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Changes in surging outlet glaciers of the Langjokull Ice Cap, Iceland.

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This study investigated the surging Hagafellsjökull outlets of the Langjökull ice cap, Iceland. It utilises digital elevation models from 1986, 1997, 2004 and 2007 in order to assess topographic change. These changes are linked to the surging outlets in terms of alteration of the subglacial hydrological system. Flux of water through the subglacial system is considered using a degree day surface melt model. Possible mechanisms of surging are considered and linked to the apparent disparity in surging between the neighbouring outlets Hagafellsjökull Eystri and Hagafellsjökull Vestari. It is found that accumulation in the upper reaches of both outlets led to increased overburden pressure of ice. This resulted in a partial flow switch from the southern Hagafallsjokull outlets to more northern outlets. The loss of flow is considered to have led to instability in the subglacial drainage system resulting in a surge of Hagafellsjökull Eystri and a partial, but failed, surge of Hagafellsjökull Vestari in 1998. Modelled changes in neighbouring subglacial hydrological systems are linked to historic evidence that more outlets of Langjökull ice cap may be, or may have been, surge type. The possibility is suggested that Sudurjökull and Þrístapajökull may well have been subject to surging through alteration of their subglacial hydrological systems, most likely related to the Hagafellsjökull system. The future of Langjökull is considered and agreement is made that the ice cap is retreating with the potential to melt completely within the next 150 years. Future surges seem likely: primarily Hagafellsjökull Vestari is expected to surge within the next 5 years due to increasing imbalance and loss of subglacial meltwater flow. Hagafellsjökull Eystri, post 1998 surge, is also suggested to have returned to a period of quiescence and recent data shows moderate surface elevation increases characteristic of an outlet building up to a surge. Future surge behaviour may also be influenced by increased melting through climatic change and precipitation increases with the possibility of increased surge incidence suggested. The techniques employed are suggested to be useful and highly transable to other studies provided adequate data is available.

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- For good old Sir Tom and to "Sticking your toes in" -

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Declaration

I hereby declare that this dissertation is the result of my own work and includes nothing which is the outcome of work done in collaboration except where specifically indicated in the text. The Dissertation is no more than 20,000 words in length excluding the acknowledgements, declaration, list of references, tables, captions and appendices.

1 - Introduction

1.1: The glaciers of Iceland

At the last glacial maximum (*circa* 20 ka BP) ice caps of Iceland extended onto the continental shelf with ice thicknesses of around 2000 m before rising sea levels caused the collapse of the marine section around 13 ka BP. The Younger Dryas (*c.*10 ka BP) saw a brief re-advance followed by rapid retreat and readvance to similar positions around 9.8 ka BP (Norðdahl, 2008).

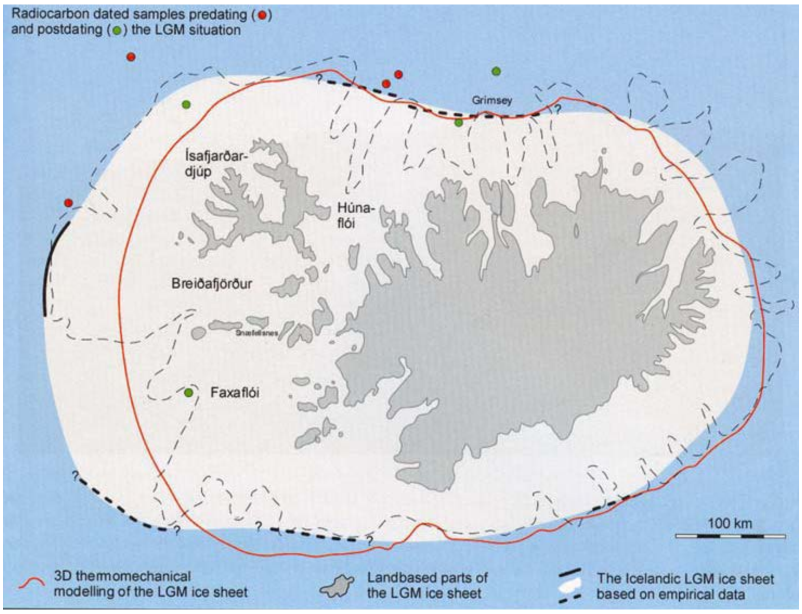


Figure 1.1: The maximum extent of glaciation in Iceland (c. 20ka BP) showing physical evidence and modelled outlines (from Norðahl et al., 2008)

Retreat from this extent saw Icelandic ice caps reach similar positions to their contemporaries around 8 ka B.P. This retreat continued, and during the climatic optimum of the mid Holocene Icelandic ice caps were substantially reduced. Neoglacial cooling (beginning c. 5-6 ka BP) saw a series of advances of Icelandic ice caps between this date and the Little Ice Age, which began in the 16th century. (Gudmundsson, 1997). The Little Ice Age maximum was reached in the 18th or 19th century (Kirkbride & Dugmore, 2006). Worldwide, smaller ice caps and glaciers (other than the Antarctica and Greenland ice sheets) have been estimated to have contributed around 0.25 mm a^{-1} to global sea levels between 1961 and 1990 (Dyurgerov & Meier, 1997). Measuring glacial variation associated with climate change is important considering the effect changes can cause. At a wide scale even a small sea level rise poses a threat to millions of people worldwide who inhabit low-lying land, and to entire island nations. In has been estimated that if they were to melt completely small ice caps and glaciers would raise sea level by about 0.5 m (Jóhannesson , 2006) – a disastrous prospect for areas already struggling with problems of flooding and increasing soil salinity. At a more local scale these small ice caps and glaciers are vital for resources such as agricultural irrigation, water for human consumption, supplying major navigable rivers for transport and, increasingly, for hydroelectric power production. For example, the Svartisen ice cap other smaller ice caps and other smaller surrounding ice caps cover about 50% of the drainage area that is utilised by the Svartisen hydropower plant in northern Norway (Jóhannesson *et al.,* 2006). Glaciers have provided Iceland with valuable resources. Iceland's main agricultural areas of the south and west which are constructed of glacial/fluvioglacial sediments in the early Holocene (Björnsson & Pálsson, 2008). Icelandic icecaps and geothermal activity have also provide vital water supplies and have been harnessed from hydro and geothermal electricity production (Árnason *et al.,* 2001). Iceland has no fossil resources so the contribution of Icelandic glaciers of around 20% to river runoff and groundwater supplies has great importance (Jóhannesson & Sigurðsson, 1998).

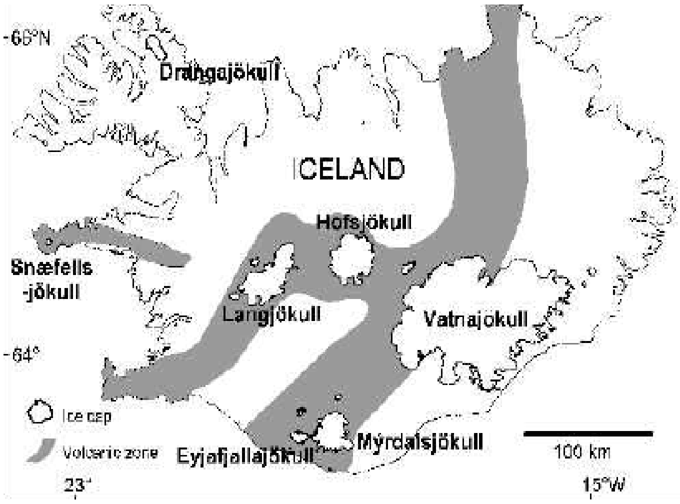


Figure 1.2: The major ice caps of Iceland, showing volcanically active areas in grey From: Bjornsson et al.,

Glacier variation in the past, present and future can, and will, document widespread changes in the climate of mountain regions that are often poorly represented by point meteorological measurements (Jóhannesson & Sigurðsson, 1998). Approximately 11% of Iceland is covered by glaciers as either ice caps or valley glaciers and they are all classified as temperate (Björnsson & Pálsson, 2008). The glaciers of Iceland are more dynamic than the majority of other Arctic glaciers due to several factors. The high volcanic activity of the island can cause to jökulhlaups induced by sudden ice melt. Jökulhlaups also occur more predictably at sites of geothermal heat flux that cause more gradual but high rates of melting accumulation and periodic release of water from subglacial lakes. 60% of Iceland's current glacial area is underlain by a volcanic/geothermally active bed (Björnsson & Pálsson, 2008) and this gives rise to frequent threats to inhabited areas. The Grímsvötn area of Vatnajökull is especially effected by geothermal activity (Nye, 1976; Björnsson, 1992; Jóhannesson, 2002) with the most recent outburst in 1996 releasing up to 50,000 m^3 s^{-1} of water from beneath the outlet Skeiðarárjökull at peak flow , which almost completely flooded the Skeiðarársandur plains.

Response to climatic variation also affects the dynamics of Iceland's glaciers. Precipitation in Iceland generally arrives from the south on prevailing winds: hence the greatest amounts of precipitation are found on the southern highlands and decease towards the north (Flowers *et al.,* 2007). Precipitation levels of up to 7000 mm a^{-1} are recorded on the southern slopes of Vatnajökull. These heavy falls are a result of the elevation and also due to the convergence of warm, moist tropical air and cold arctic air (Björnsson & Pálsson, 2008). This maritime climate leads to Icelandic glaciers being characterised by high precipitation and large mass turnover compared with glaciers in continental environments (de Woul *et al.,* 2006). Whilst summer balance levels on the highest ice caps are usually negative, a cooler summer can give a marginally positive summer balance in the highest reaches resulting in a marked increase in albedo and an associated reduction in ablation. This interannual variability demonstrates how small climatic changes may have a considerable effect on Iceland's glaciers. Responses to climate shifts are seen at the glacier snout of an Icelandic icecap within approximately a decade, depending upon the size and location of the ice cap (Sigurðsson & Jónsson, 1995). Due to the Irminger current (a maritime current of warm water) Iceland benefits from a relatively small variation in temperature with average winter temperatures on the southern coast close to $0^{o}C$ and a mean annual temperature of $5^{o}C$ (Einarsson, 1984).

Another reason the glaciers of Iceland are so dynamic is their surge behaviour. This is discussed in the next section.

1.2: Surging Glaciers in Iceland

Surging glaciers in Iceland have a relatively high incidence and cover all ranges of climatic conditions. They occur on all of the major Icelandic ice caps as outlet glaciers and also in individual mountain glaciers. They show no spatial relationship to geothermal heat sources (Björnsson *et al.*, 2003). Surging glaciers are also well distributed in terms of basal geology; Vatnajökull glaciers are underlain by impermeable bedrock (basalts) contrasting with those from Mysdalsjökull and parts of Langjökull which may be underlain by porous lavas – young formations of the late Tertiary/early Quaternary (Björnsson *et al.*, 2003; Sigurðsson, 1990). In total there are 26 identified surging glaciers in Iceland with size ranges from 0.5 to 1,500 km^2 (Björnsson and Pálsson, 2008). 80 surges are recorded through history with advances ranging from tens of metres to around ten thousand metres (Sigurðsson, 1998). Icelandic surging glaciers are characterised by more gently sloping surfaces of 1.6 - 4° - somewhat lower than the *c.*12° average of the non-surging Icelandic glaciers (Björnsson *et al.*, 2003). As characterises surging glaciers worldwide they flow too slowly to maintain a balance between accumulation in the upper reaches and ablation due to ice flowing to the glacier snout. As a result they become out of equilibrium and this is theorised to be the cause of their surging characteristic (see Section 2).

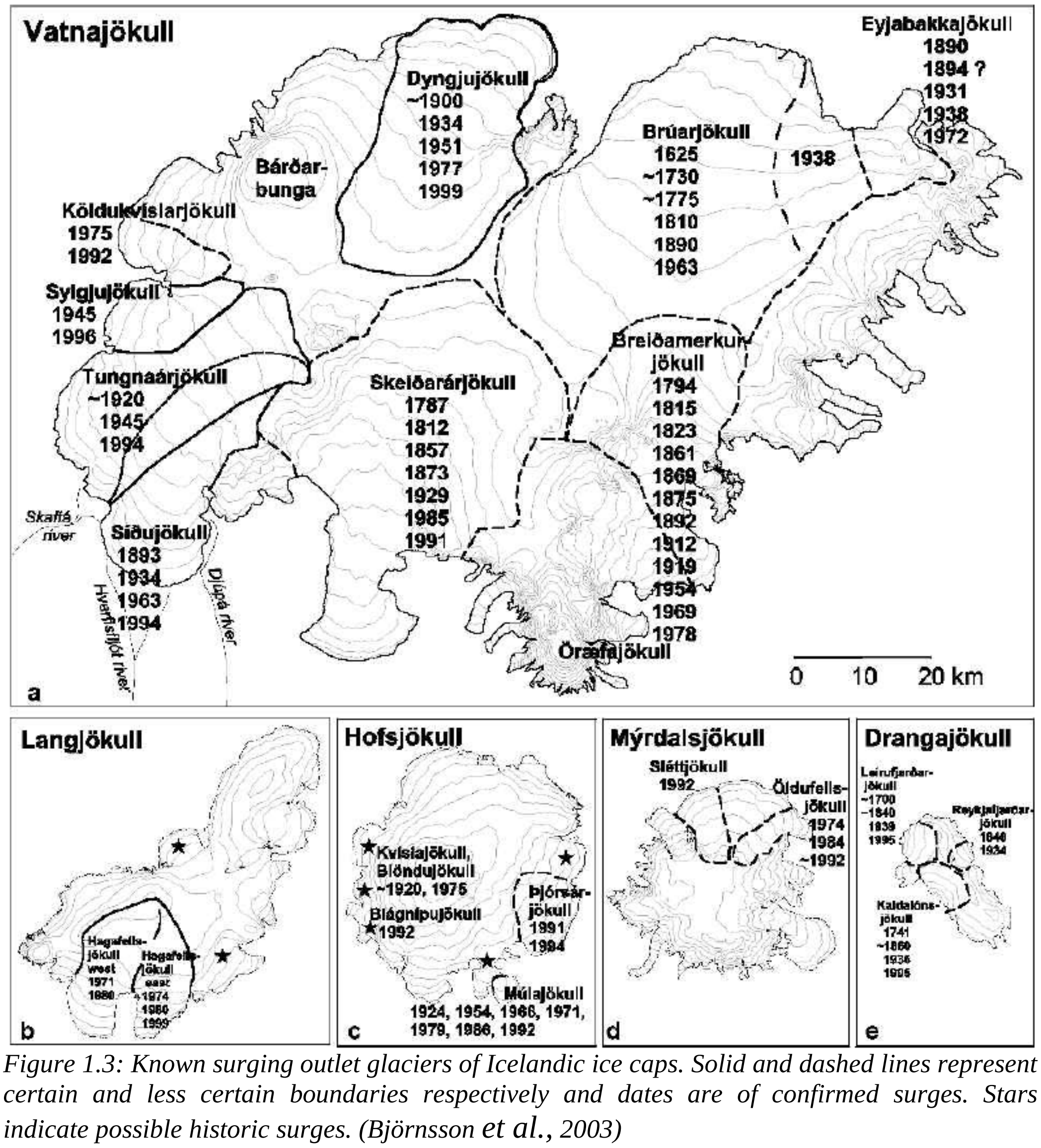


Figure 1.3: Known surging outlet glaciers of Icelandic ice caps. Solid and dashed lines represent certain and less certain boundaries respectively and dates are of confirmed surges. Stars indicate possible historic surges. (Björnsson et al., 2003)

Surges therefore play an important role in transporting ice and restoring the equilibrium of many Icelandic glaciers. For example, surging outlets of Vatnajökull are estimated to be responsible for the transfer of at least 10% of total ice flux to ablation areas. Surges also increase out flowing water sediment loads markedly (Sharp, 1985). In particular the finest grain sizes are increased with loads of 7-10 kg m^{-3} recorded following surges of Vatnajökull (Björnsson & Pálsson, 2008). This is a substantial increase over more normal sediment loads of ~1.5 kg m^{-3} (Flowers *et al.,* 2007).

1.3: The Langjökull Ice Cap

Langjökull is Iceland's second largest ice cap with an area of around 925 km^2 and a total ice volume of ~195 km^3 (Palmer *et al.,* 2009). This gives the potential for 0.5 mm of eustatic sea level rise. The ice cap is located in central western Iceland approximately 85 km north east of the capital, Reykjavík and is orientated SW-NE. Radio echo sounding studies have revealed a mean ice thickness of ~200 m and a

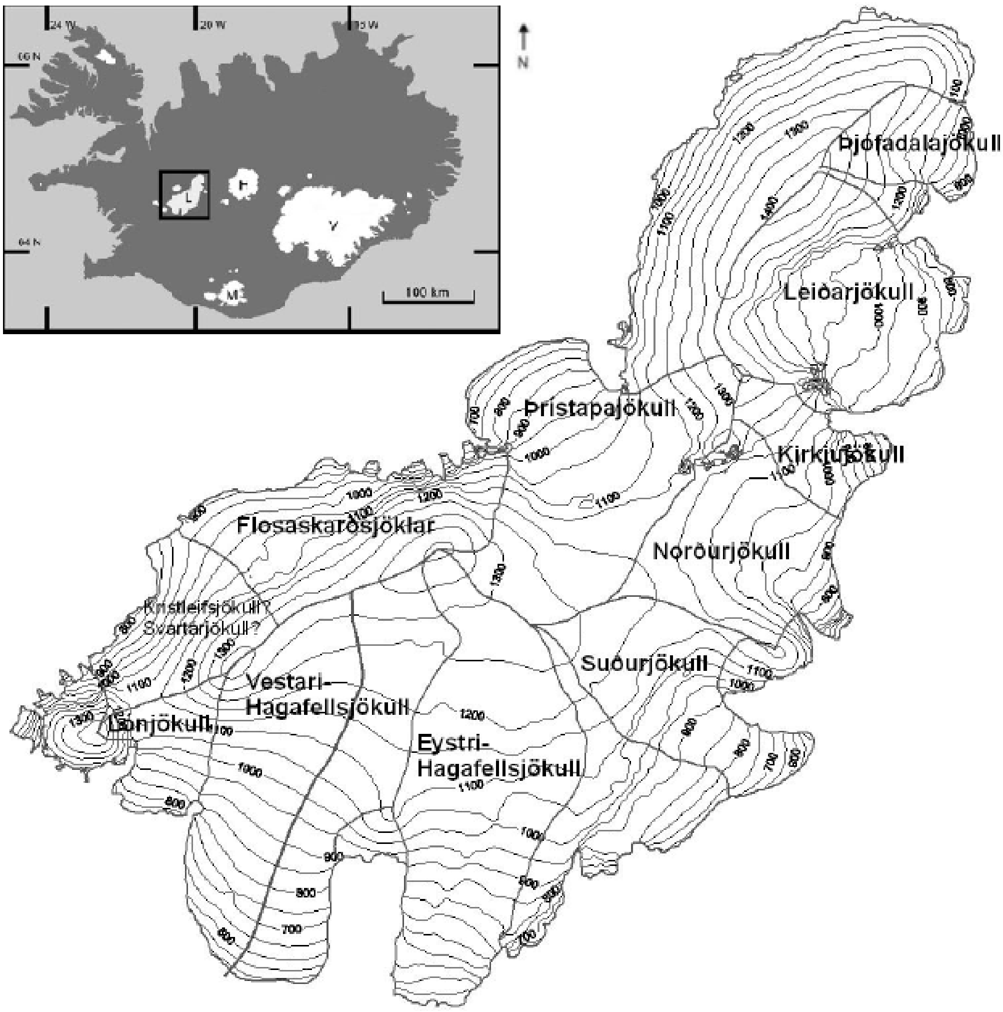


Figure 1.4: Map of the Langjökull Ice Cap showing elevation (m a.s.l.), ice divide, and central flow lines. Large map courtesy of Finnur Pálsson and Helgi Björnsson. Inset map from Eyre et al., (2005). L, H, M and V refer to Langjökull, Hofsjökull, Mýrdalsjökull and Vatnajökull respectively.

maximum of 580 m (Björnsson *et al.,* 2006). Surface elevations extend from around 500 m up to around 1500 m a.s.l. (Figure 1.4) with the equilibrium line altitude of the Langjökull ice cap at around 1000 m. Langjökull is considered to be a completely temperate ice mass throughout, with moulins signifying melt water is freely able to reach the glacier bed - believed to be of deformable sediments of a porous lava bedrock material (Eyre *et al.,* 2005). Around 69% percent of drainage from the entire ice cap is calculated to flow out as groundwater (Flowers *et al.,* 2007) and the remainder subglacially. No postglacial volcanic activity is in evidence beneath Langjökull, despite the widespread volcanic activity of Iceland. No jökulhlaup has ever been recorded to have issued from the ice cap (Sigurðsson, 1998).

Precipitation on Langjökull is similar to that of neighbouring Hofsjökull with lower amounts than the most southerly ice caps due to the prevailing wind direction. Precipitation of around 3,500 mm a^{-1} are recorded. A negative mass balance from observations during the period 1996-2006 caused a total mass loss of 13.1 km^{-3} (7% of the total) during this period (Bjornsson & Palsson, 2008). Model predictions of a 2.8°C temperature rise and 6% precipitation increase towards the end of the 21st century suggest rapid changes for the future of the Langjökull ice cap (Bjornsson & Palsson, 2008). These climatic changes would diminish Langjökull by around 35-40% of its present volume during 50 years and would see the ice cap disappear in 150 years (Jóhannesson & Sigurðsson, 1998; Björnsson & Pálsson, 2008). Studies of sediments from lake Hvítárvatn imply Langjökull may have completely melted during the warmest Holocene period around 10 ka BP before accumulation began again around 5-6 ka BP (Black *et al.,* 2006). The Holocene maximum of the ice cap is thought to have been attained around 250 years ago. Melt since this time has led to sedimentation of the Little Ice Age portion of proglacial lake Hvitarvatn (Flowers *et al.,* 2007).

The Langjökull ice cap has several outlet glaciers (figure 1.4). A recent study utilising Interferometric Synthetic Aperture Radar (InSAR) identified eight principle outlet glaciers with flow speeds of up to 75 m a^{-1} (Palmer *et al.,* 2009). Of these, two are recorded to be surge type: Hagafellsjökull Eystri and Hagafellsjökull Vestari (Sigurðsson, 1998). These Hagafellsjökull outlets are separated by the Hagafell ridge and are the two main outlets of the Langjökull ice cap. As described previously the southern outlets of Langjökull are believed to be underlain by porous lavas (Bjornsson *et al.,* 2003). In addition to the Hagafellsjökull outlets the InSAR study described above also suggests the smaller, easterly glacier Suðurjökull may be surge type after observing increased flow velocity in 1994 (Palmer *et al.*, 2009). Björnsson *et al.*, (2003) also suggest Suðurjökull and the more northerly Þrístapajökull (figure 1.4) may also be surge type although evidence is anecdotal only. The surges of the Hagafellsjökull glaciers after the 1970s (described below) seem to have been preceded by at least 40 years of quiescence (Björnsson *et al.*, 2003). It is suggested the surges may be a reaction to positive mass balance from the 1960s-1980s.

1.3.1: Hagafellsjökull Eystri (East)

Hagafellsjökull Eystri is around 4 km wide and 25km long. It currently terminates in proglacial lake Hagavatn. It is constrained to the by a volcanic ridge (Jarlhettur) and in places overflows the ridge and forms a series of small piedmont lobes in the Jarlhettukvísl Valley (Bennett *et al.*, 2005). To the west it is constrained by the Hagafell ridge. Surges have been recorded in 1974, 1980, and 1998. No earlier surges are recorded although landforms that mark the ice maximum may be the result of surge-like oscillations (Bennett *et al.*, 2005). Monitoring of this outlet glacier started in the 1930s (unsystematically) and by professional surveyors during the 1960s (Sigurðsson, 1998 ; Björnsson & Pálsson, 2008). During the three recorded surges the glacier advanced between 900 and 1500 m during the period of late winter or early spring (Sigurðsson, 1998 ; Bennett *et al.*, 2005). The latest surge began slowly in 1998 with the advance of the piedmont lobes in the Jarlhettukvísl valley with a rapid increase in velocity in April 1999 (30 m in 24 hours). It then advanced 1165 m during the subsequent six weeks.

1.3.2: Hagafellsjökull Vestari (West)

Hagafellsjökull Vestari is approximately 7 km wide and 25 km long. It is constrained to the east by the Hagafell Ridge. This outlet surged in 1971 and 1980 with no surge recorded since. The recorded terminus advances were in the region of 650-720 m and like those for Hagafellsjökull Eystri, the surges began in late winter to early spring. The 1980 surge of both Hagafellsjökull outlet glaciers coincided with each other. Although there appeared to be no surge (*i.e.* an advance of the ice front) of Hagafellsjökull Vestari to coincide with the 1998 surge of Eystri there is some evidence that an incipient surge may have been prematurely halted. Finnur Pálsson (2010, personal communication) recorded highly increased flow velocity that led to some transport of ice from the accumulation area to the ablation area through greatly increased flow velocities. However, the surge wave of Hagafellsjökull Vestari did not reach the glacier margin and the surge came to a halt. Fluvial erosion in front of the glacier in autumn 1998 suggests a sudden flood of water from beneath the glacier may have terminated the surge early (Björnsson *et al.*, 2003).

2: Literature Review

2.1: Glacier surging

Surging glaciers are glaciers that exhibit cyclic flow instability: flow is usually at a relatively constant rate but is punctuated by periods of rapid flow. This rapid flow (surging phase) is usually between 10 and 1000 times that of the normal flow (quiescent phase) (Murray *et al.*, 2003). Around 1% of worldwide glaciers have been classified as surge type (Jiskoot *et al.*, 2001) and the distribution of surging glaciers worldwide is described as non-random (Clarke, 1976). Understanding this non-random distribution would be useful for placing constraints on the mechanisms responsible for surging but this has so far proved difficult to understand (Harrison & Post, 2003). Bedrock parent material has also been the subject of study with sedimentary rocks seemingly providing a greater likelihood of surging in Svalbard (Hamilton & Dowdeswell, 1996). These supports the suggestion that surging glaciers are more likely in glaciers overlying 'easily eroded materials' (Harrison & Post, 2003 : 1). This implies surging may be related to till at the glacier bed, which seems to be present, at least in part, at the beds of all surging glaciers. Other studies of the distribution of surge type glaciers suggest length may be a causal factor in some regions: longer glaciers in Svalbard and the Saint Elias mountains of Alaska/Canada seeming more likely to surge (Clarke, 1991). For some surging glaciers the length of surging and quiescent phases tends to be relatively constant allowing reasonably accurate prediction of when a surge is likely to occur. Others, however, have show more variation – Variegated glacier, Alaska, U.S.A, has experienced quiescent phase variation of 10-18 years during the 20th century as discussed later (Eisen *et al.,* 2001). There is much variation between surge phase lengths in differing areas of the world. Regions characterised by temperate glaciers have much shorter surging phases that may last only last for periods of years or even months (Joughin *et al.*, 1996). Conversely, subpolar Svalbard glaciers are characterised by longer surges than others typically lasting >5 years with lower flow velocities, (Murray *et al.*, 2002). This more subdued style of surging compared to temperate glaciers is likely linked to the thermal regime of the glacier. Clearly, the thermal regime is one element of the glacial system that could be linked to climate. The effects of a changing climate make understanding the distribution of surging glaciers even more complex because of the constant evolution of glacier topography through time which might alter the suseceptability of glaciers to surging – glaciers which surge now may not necessarily surge in 100 years (Harrison & Post, 2003) For example, Vernagtferner in the Tyrol region of the European Alps, underwent a series of large surges (c. 2km) from the late 16th century until the late 19th century and then ceased to surge (Hoinkes, 1969). During this period glaciers in the Alps were relatively large due to climatic conditions. As climate changed the ability of the glacier to surge was diminished. Quiescent phases exhibit a similar variability that could be of the order of several to over 100 years (Kamb *et al.,* 1985). As briefly mentioned above previous climate has the potential to affect the length of quiescence (or the frequency of surge recurrence) through alteration of the mass balance of the glacier system (Eisen *et al.,* 2001). The surges of Variegated Glacier, Alaska, and Medivizhiy glacier, Russia have been correlated with cumulative mass balances in their respective reservoir areas (Dyurgerov & Meier, 1997, Harrison & Post, 2003). A changing climate could influence the mass balance of a glacier thereby altering the surge frequency. This influence can be dependant upon where surges are initiated within the glacier system. Glaciers with surges that are initiated in the higher reaches may be more slowly affected by climatic alteration than glaciers with surges initiating in lower reaches. Surges often commence in early winter when the least meltwater is available (Harrison & Post, 2003). Intuitively it would be expected that a changing climate could influence the timing, or even possibility of a surge through changing patterns and quantities of melt.

The identification of surging glaciers is possible during both surging and quiescent phase although via differing indicators. During the surging phase identification is possible due to rapid advance, high surface velocities, heavy crevassing, shear margins between rapid and non rapid ice flow zones, a steep ice front and stranded ice on bedrock topographic high points due to rapid surface lowering (Post, 1969 ; Copland *et al.,* 2003). Characteristics identifiable during quiescent phases are fewer -

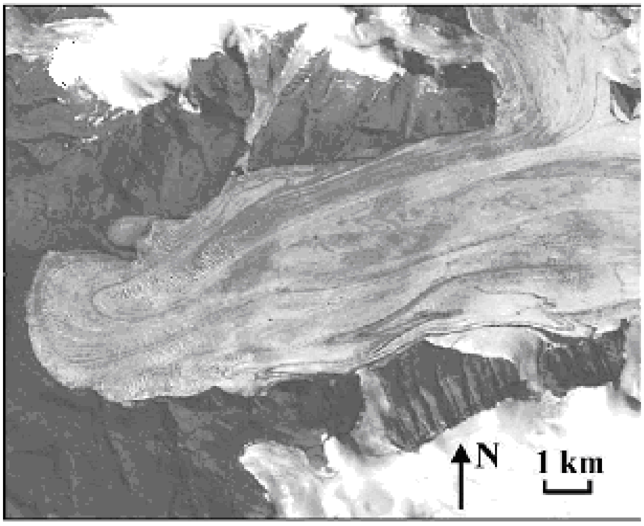
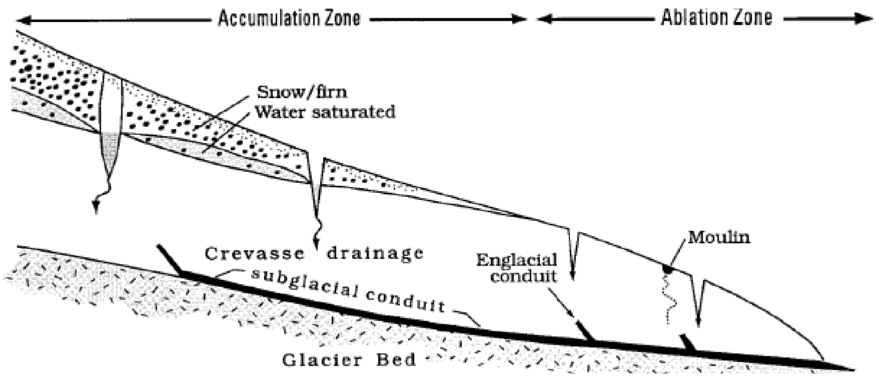


Fig. 2.1: Looped moraines characteristic of surging glaciers seen here in Landsat 7 imagery of Airdrop glacier, Canadian High Arctic. Also evident is a highly crevassed surface near the terminus (Copland et al.,

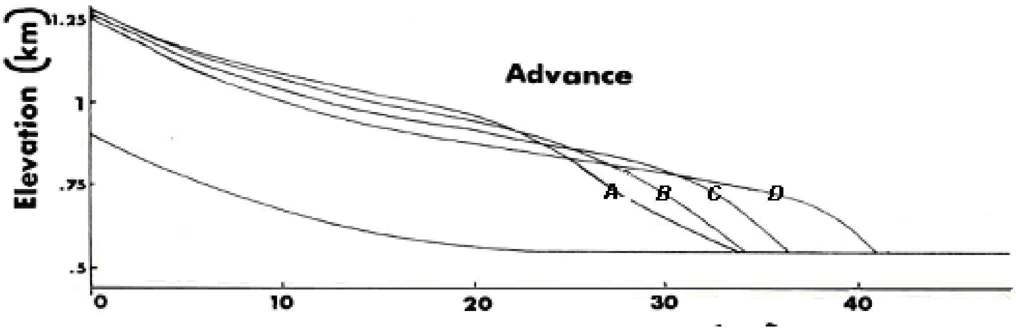
push moraines due to advance of the snout are possible identifiers. The most clearly identifiable features are looped moraines (Figure 2.1) formed when tributary glaciers advance into a main glacier. Surges can pose hazards to agricultural land and property and also more widespread hazards such as outbursts from ice dammed lakes. The surge of the Grande del Nevado glacier Argentina led to a rapid advance of several kilometres in 1933 (Espizua, 1986). This formed an ice dammed lake which in turn caused a catastrophic flood of the Rio Mendoza when released.

2.2: Surging mechanisms

Two main mechanisms that lead to surging have been proposed. These may explain why some glaciers exhibit different characteristic during surging phases - such as many Svalbard glaciers that surge more slowly than most. Surging glaciers are caused by 'dynamic instability of the glacier systems themselves and are unrelated to external factors' (Dolgoushin & Osipova, 1975: 1) although the effect of external factors from a changing climate may be influential as this work seeks to investigate. It has long been noted that surging glaciers, regardless of surge mechanism, exhibit an increase of mass in their accumulation zone that is greater than the loss in the ablation



Figs 2.2 (above) and 2.3 (Below): Idealised non-surging and surging glacier systems. In figure 2.3 the letters A-D symbolise how the glaciers profile changes as the surge transports mass from the accumulation zone to the ablation zone causing advance. Adapted from Fountain & Walder (1998) and Budd (1975) respectively.



area. Figure 2.2 demonstrates an idealised temperate glacial system in which the glacier is in balance with ice flow to the ablation area equalling the accumulation. Figure 2.3 demonstrates how a theoretical surging glacier increases mass in the accumulation area and becomes unbalanced because flow is not sufficient to balance it. The surface slope is visibly steeper because of this imbalance. The evolution of the glacier profile through a surge is then shown. During this period mass is rapidly transported to the ablation zone via increased flow velocities to return the glacier to a balanced state with an advance of the snout position. It has long been recognised (Robin, 1955) that the only mechanism by which such flow velocities could occur is increased basal sliding but the causal mechanisms leading to such considerable increases remained elusive. The theorised mechanisms that lead to surges of glaciers are summarised below.

2.2.1: Link Cavity System surging

The first mechanism is applicable to temperate glaciers, which have water available at the ice-bed interface. One of the best documented glacier surges is Variegated Glacier, Alaska, which was heavily monitored during a surge in 1982-83. The ice mass was known to be at melting point throughout (Kamb *et al.*, 1985) ruling out possible thermally controlled glacial surging (see section 2.2.3). Kamb (1987) proposed a model of surging that involving a linked cavity configuration of the basal hydrological system. Put simply, during non-surging periods water flows out freely through a channelised system, which drains the glacier efficiently. A surge might be initiated when water becomes trapped in linked cavities at the glacier bed and the glacier is no longer efficiently drained. This occurs because ice pressure throttles the drainage channels. If meltwater flux remains constant the throttling of drainage channels can be due to *increased pressure of ice* causing channels to collapse. Conversely, if the pressure of ice remains constant the throttling can be due to a *decrease in*

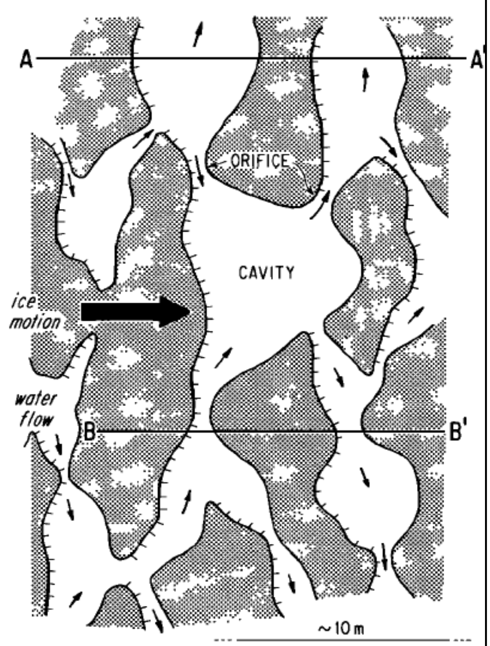


Figure 2.4: The linked cavity system as proposed by Kamb et al., (1985)

meltwater flux failing to maintain the channels through melting thus allowing them to collapse. Once drainage channels collapse, trapped water increases the basal water pressure greatly and increases the lubrication at the glacier bed by effectively forming a layer of water at the ice/bed interface. Thus the ice flow velocity increases greatly, also driven by the increased slope of the glacier due to accumulation in the upper reaches. The surge ceases when the water is released and the throttling by ice reduced sufficiently for drainage channels to reform. Many observations support this theory. During the 1982-3 surge of Variegated Glacier the basal water pressure was greatly increased reaching totals occasionally equal to the overburden pressure (generally 2-5 bar below overburden pressure). Immediately prior to the surge basal water pressures were much lower (4 -16 bar below the overburden pressure) and any peaks in these values corresponded to peaks in sliding motion. These peaks were termed 'minisurges' and were observed mainly in the melt season of the years before the onset of the major surge (Raymond, 1987). They were characterised by an abrupt increase in velocity over a period of a few hours with a slowdown of around a day. These minisurges pulsed down glacier as waves at ~0.1 – 0.6 km h^{-1}. A rapid onset and termination of the surge was recorded and major slowdowns in ice flow velocity corresponded to outflow floods of much more turbid water than normally produced (more than ten times normal amounts) and also resulted in a drop in the glacier surface height of 0.1-0.7 m. Dye tracing experiments showed water flow through the glacier to be much slower during surging phases with a drop from around 0.7 m s^{-1} to 0.025 m s^{-1}. These dye experiments also showed much greater dispersal of the dye during the surging phase with dye emerging from all the outflow streams. The transition between the channelised flow system and the linked cavity system is modelled by Kamb (1985) as:

$$\boldsymbol{\psi = aw^{3/2}\, h^{7/6}\, (\acute{\eta}/v)^{1/2}\, (Pi - Pw)^{-1/2}\, M^{-1}} \qquad \text{(Eq. 1)}$$

where aw is the hydraulic gradient, h is the orifice step height (as in figure 2.4), ή is ice viscosity, Pi is the ice overburden pressure, Pw is the subglacial water pressure and M is the Manning roughness coefficient. Values of ψ of less than 0.8 are suggested as forming a stable cavity system. Fowler (1987) links channel instability to the increase in velocity. His theory states that the stability of each drainage system depends mainly upon the velocity. Velocity increases as the glacier thickens in the accumulation area (and thus gains a steeper profile overall). At a critical velocity the system switches from a tunnel based system to a linked cavity system. This results in an activation wave of hydraulic transition front travelling up and down glacier at approximately 50 m hr^{-1}. At the wave front the transition from one system to another occurs rapidly through the collapse of the tunnel system leading to the high pressure, link cavity system. The starting point of this wave is likely to be the area termed the 'zone of enhanced velocity' (Björnsson *et al.*, 2003: 87). This zone experiences increased velocity in the months prior to a surge leading to a step-like bulge in the lower part of the zone. On Vatnajökull and Langjökull the zone has been observed to be approximately 10km long in the upper ablation zone. Movement of this bulge is often the first sign of a surge with rapid velocities (several tens of metres per day). Fowler (1989) also applies this theory of velocity relation to a surge of Hubbard Glacier, Alaska where similar observations were made and the theory of the linked cavity system is supported. It is also suggested a wave of deactivation exists which spreads in the opposite direction to the activation wave, restoring the efficient channelised drainage of the glacier.

2.2.2: Thermally Regulated Surging

Alongside the linked cavity mechanism another theory of surging is proposed – thermally regulated surging. This provides an alternative mechanism of surging that is relevant to cold based/polythermal glaciers, and often glaciers overlying frozen or partially sediments. These glaciers are generally found at high latitudes, such as those found in Svalbard. A positive feedback system exists that involves thicker ice warming the bed of the glacier thus permitting increased basal sliding. This idea was first suggested by Robin (1955) and was called 'thermal instability'. The basic principle is that during the quiescent phase ice is frozen to the bed and that ice flow by sliding/ice deformation is limited. Hence, mass accumulates and increases the pressure of ice. The surging phase is initiated when pressure warms the glacier bed to melting point through both frictional heating from sliding and also sheer stress due to the ice pressure. This enhances sliding and ice deformation increasing the flow velocity of the glacier. This increased sliding feeds back into the system in the form of increased frictional heating thus causing more melt and increasing the ice flow velocity further. The surge ceases when the heat is lost by advective cooling of the ice-bed interface and when ice mass transfer to the ablation zone reduces the driving force for ice to flow. This theory was later altered (Robin, 1969) and featured a thicker layer of warm ice and the surge halting for mechanical reasons rather than the advection of heat. Clarke (1976) suggested this 'thermal instability' was a viable mechanism for surging of cold based glaciers with basal temperatures at or close to freezing but stressed that another mechanism for the surging of temperate glaciers must exist, as the previous section explains. The concept of thermally regulated surging was considered further by Fowler *et al.,* (2001). This study involved the glaciers Bakaninbreen, Svalbard, and Trapridge, North West Canada, which both overlie several metres of sediments and considered the effects of temperature increase on frozen sediments beneath glaciers. The warming of the bed causes increased velocity not only by sliding/deformation of ice at the ice-bed interface but also through deformation and/or shearing of the subglacial sediments when they reached melting point. The warming of the bed is theorised to create activation waves, which spread up and down glacier. If the surging wave speed is faster than the activation wave speed an advancing wall of ice may propagate down glacier as observed at Trapridge and Bakaninbreen. This mechanism initiates a slower surge initiation and overall ice velocity as it relies on the weakening of underlying till by the pore water resulting from pressure melting once sufficient ice has built up in during the quiescent phase. A gradually accelerating till deformation releases steadily more heat and a positive feedback exists until sufficient ice thinning allows heat to dissipate and the glacier to refreeze to its bed.

Of course, many of the systems described above are idealised and theoretical. At one extreme are glaciers that would potentially be temperate and overlying a bed of impermeable bedrock. At the other would be a cold based glacier with a thick, entirely frozen sediment bed. Neither of these two perfect extremes is likely to be encountered – particularly considering that sediment seems to be found at the base of all surging glaciers (and of course non-surging ones). The thermal regime of glaciers is also highly variable with many being polythermal and featuring areas of both frozen bed and temperate bed. The interaction of hydrological and sediment systems is inevitable. As described above the hydrological system may play an important role in the till strength beneath glaciers but conversely the hydrology is equally likely to be influenced, or even controlled, by the sediment present on the bed (Harrison & Post, 2003). Part of the lack of understanding of these interactions is due to difficulties of observations at the ice bed interface.

2.3: Glacial topography influences on subglacial hydrology – flow switching and water piracy

As the previous section has described changes in glacial surface topography can have important influences on subglacial hydrology due to the effects of increasing/decreasing ice mass on processes occurring at the glacier bed. Ice surface topography changes are also present in the ice streams of Antarctica and have considerable, although interestingly different, consequences for subglacial conditions. Ice streams are not generally believed to surge although it has been suggested that the Kamb Ice Stream may be subject to surging and is currently in a quiescent phase (Rose, 1979). Alley *et al.*, (1994) reason that this is unlikely because surge type glaciers spend most of their time in the quiescent phase yet four out of five of the Antarctic Siple Coast ice streams are currently flowing quickly. Topographic differences between the ice streams are also few with no sign of the characteristic increase of mass in the accumulation area – Kamb Ice Stream is very similar in appearance to its neighbouring ice streams. Despite this, the fact remains that Kamb Ice Stream is currently flowing much slower than its neighbouring ice steams with velocities of <10 m yr^{-1} compared to neighbouring Whillans Ice Stream with velocities ~800 ma^{-1} (Anandakrishnan & Alley, 1997). This apparent cessation occurred approximately 140 years ago. One explanation for this rapid change is explained by Alley *et al.*, (1994) and Anandakrishnan & Alley (1997). Holocene surface warming and subsequent basal warming allowed the headward extension of the Kamb Ice Stream. This lowered the surface slope of the ice stream thereby changing its topography considerably. This allowed the

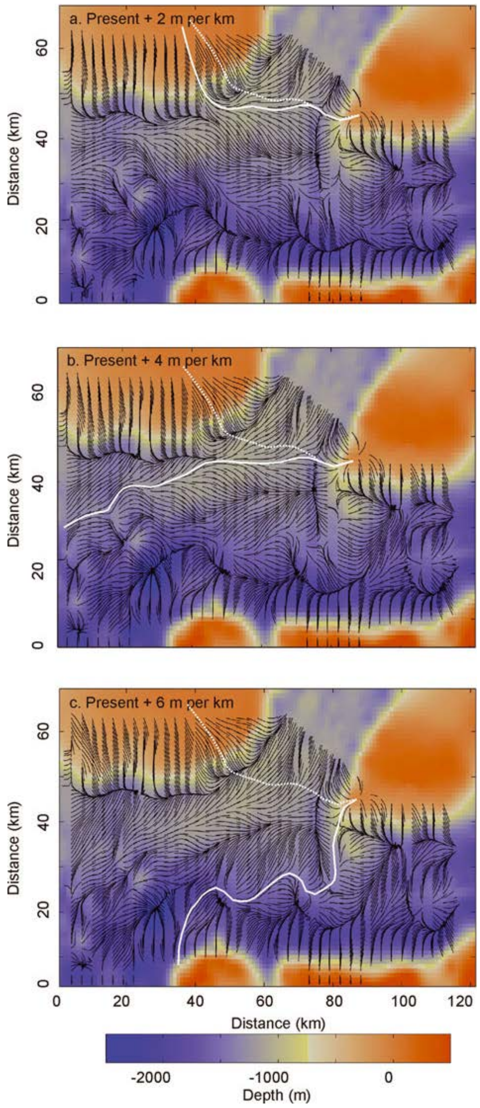


Figure 2.5: The changing subglacial hydrological potential at the heads of Rutford Ice Stream and Carlson Inlet, Antarctica. The dashed line is the current flow divide and the solid line represents the change with Rutford Ice Stream thickening 40, 80 and 120 m. (Vaughan et al., 2008).

subglacial hydrological system to change – particularly the hydrological potential. Basal slopes allowed the water to flow beneath the neighbouring Whillans Ice Stream that may have also been extending headward towards the drainage catchment of the Kamb Ice Stream. In effect this removed the lubricating water beneath Ice Stream C (required for basal sliding and also to enable till deformation) thereby causing stagnation of the Kamb Ice Stream.

In a similar study in Antarctica (Vaughan *et al.*, 2008) analysed the contrast between two neighbouring ice streams; the Rutford Ice Stream and the Carlson Inlet. These glaciers are similar in size and driving stress and yet Carlson inlet exhibits flow speeds approximately 10-50 times lower than Rutford Ice Stream, which flows at ~350m a^{-1}. Carlson inlet is believed to have been an active ice stream until around 250 a^{-1} BP. Evidence to support this is in the form of relict shear margins (Doake *et al.*, 2001) and water content in basal sediments despite steady state calculations showing that the glacier should be frozen to its bed (Frolich *et al.*, 1989). Its is suggested that Carlson Inlet has stagnated because of a flow switch to Rutford Ice stream – possibly caused by a thinning of the Rutford Ice Stream altering the hydraulic potential beneath it. Modelling has shown that even small thickness increases in the topography of Rutford Ice Stream could divert significant amounts of water towards Carlson Inlet (Figure 2.5). An ~ 4% thickness increase (around 120 m) would potentially divert all of the subglacial water towards Carlson Inlet and could reactivate it, although there is no sign of any thickening at present.

The behaviour of the Kamb Ice Stream and Carlson Inlet is very interesting and somewhat different from the behaviour of glaciers such as Variegated Glacier described previously. Absence of basal water from two Antarctic glaciers may have had the effect of substantially decreasing their flow. By contrast, although on a much more rapid timescale, a lack of water in the basal system of the valley glaciers had the effect of initiating a surge and therefore greatly increasing velocity. This difference seems to lie in conditions at the bed. The Antarctic ice streams are reported to be underlain by poorly consolidated and easily eroded sediments around 400 m thick thinner in places (Anandakrishnan & Bentley, 1993). Deformation of this sediment is essential for ice stream flow and reduced water content would decrease sediment deformation thus slowing ice stream flow. Reduced water in the basal system would also allow 'sticky spots' to exert more drag on the ice stream causing reduced ice stream flow or stagnation as described in the cases of Kamb Ice Stream and Carlson

Inlet. These sticky spots can come in various forms including areas of bedrock bumps, till-free areas and areas of ‘strong’ (i.e. well drained) till (Stokes *et al.,* 2007).

3 - Key Study Aims

There are several areas into which this study may be able to provide insight. This study intends to incorporate several different methods that should allow the assessment of numerous glaciological processes occurring within Langjökull – in particular the surging behaviour of the southern Hagafellsjökull outlets.

Topographic Change

Primarily it should be possible to bring together all of the available Digital Elevation Model data from the University of Iceland and the Scott Polar Research Institute. This will allow a view of the evolution of the Langjökull Ice Cap from 1986 to 2007 through a period of quiescence, a period of surging of at least one outlet and back into a period of quiescence. The data from 1997 and 2004 provide useful temporal reference points to gauge the magnitude of topographic change following the surge in 1998. Previous studies did not have the advantage of the 2004 data.

Surface Melt

Using a simple degree day melt model from the Scott Polar Research Institute model the surface melt across the ice cap. This model will be driven using corrected precipitation data from a nearby weather station to compliment automatic weather station temperature data from the Langjökull Ice Cap. From this it should be possible to estimate the total amount of water flux through the subglacial hydrological/groundwater system.

Subglacial Hydrology

Using the Digital Elevation Models recreate the subglacial flow system. This recreation can then be related to the surface melt model and the flux of melt water through the system assessed. The topographic change of the glacier calculated from the ice cap DEMs – particularly in the build up to the 1998 surge of Hagafellsjökull Eystri and the apparent failure of Hagafellsjökull Vestari to surge – can be assessed relating to effects upon subglacial processes. Assessing how changes in hydrological flow pattern due to topographic change may have led to the disparity between the neighbouring Hagafellsjökull outlets. It is hypothesised this is due to a switch in subglacial hydrology to an unstable system beneath Hagafellsjökull Eystri leading to increased surface velocities.

Utilising DEMs of Langjokull since the 1998 surge it will be possible to assess how changes have affected the system since. Patterns of continuing change may allow predictions into when surges may occur again and how surging behaviour may into the future – with possible links to a changing climate considered.

4 - Data sources

4.1 - Digital Elevation Models (DEMs)

*2007 LiDAR data. (**Li**ght **d**etection **a**nd **r**anging)*

The initial LiDAR data was collected on 2nd August 2007. The instrument used for data collection was an Optech ALTM3033 LiDAR system belonging to Cambridge University's Unit for Landscape Modelling flown aboard a Dornier 228 aircraft provided by the Airborne Research and Survey Facility of the UK Natural Environment Research Council. The vertical accuracy of the data was c.10 cm after processing and the data was gridded to a 10m resolution. Details of similar collection

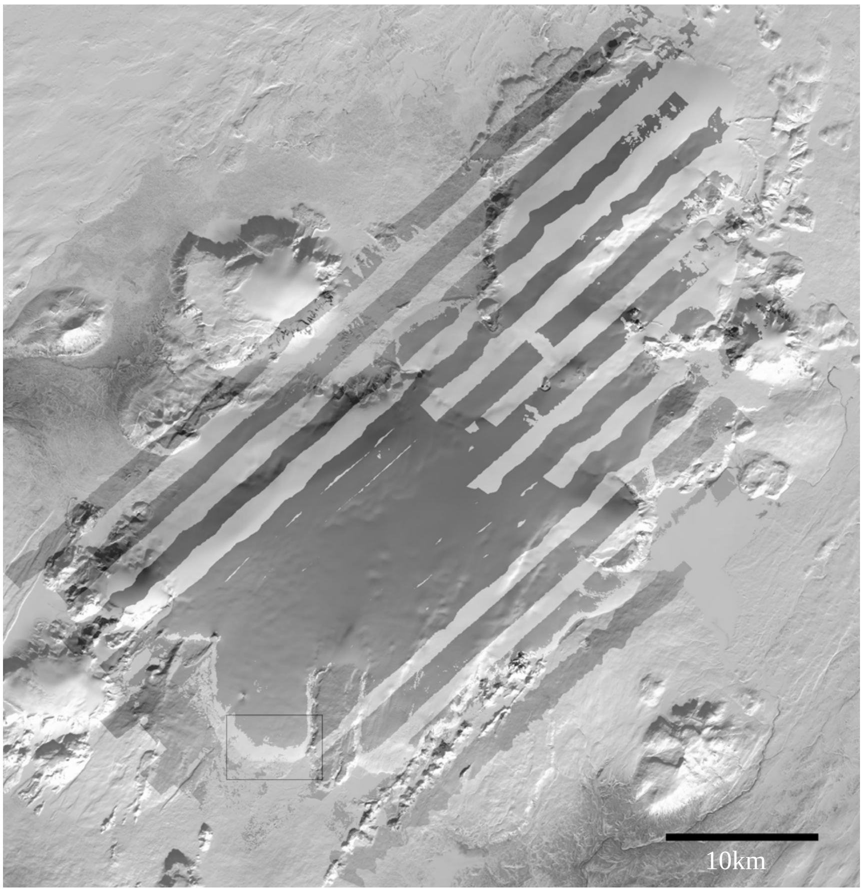


Figure 4.1: August 2007 LiDAR survey overlayed on a 2002 Landsat-derived image.

and processing techniques of data are given by Arnold *et al.*, (2006). However, due to logistical constrains the LiDAR flight data did not supply a complete DEM, as can be seen in figure 4.1. Although coverage is mostly continuous in the southeast other areas show *c.*3 km strips lacking LiDAR coverage. Comparison of the LiDAR data and overlapping summer 2007 elevation data from differential GPS tracks (collected on snowmobile) was undertaken. Differential GPS has up to *c.* 2 cm vertical accuracy depending on the distance to the base station. The data used here was provided by Finnur Pálsson and Helgi Björnsson at the University of Iceland's Earth Science Institute. This comparison was used to confirm that there was no systematic offset between the data sources, thereby acting as a check of the accuracy of the LiDAR data. The comparison showed the LiDAR data to be accurate with no offset from the GPS data.

Consequently, all differential GPS data were used to supplement the LiDAR DEM, this work was done by Allen Pope of the Scott Polar Research Institute. The DEM of Langjökull was the subject of a study using the technique of *photoclinometry* to provide a complete DEM for the ice cap. Photoclinometry is a method which unifies visible light imagery with elevation data. Basically, photoclinometry transforms the brightness of a given pixel in a visible light image into a surface slope parallel to the solar azimuth for that image. It is also known as 'shape from shading' for these reasons. Pope's study utilised a Landsat ETM+ band 4 image collected on 19 March 2002 to produce the completed DEM (henceforth referred to as LiDAR v.2) – essentially a hybrid of LiDAR, skidoo and photoclinometric data as seen in figure 4.2. A geoid correction was also produced to be applied the DEM to account for differences in the WGS84 datum: our Icelandic colleagues use a pre geoid corrected datum whereas SPRI does not.

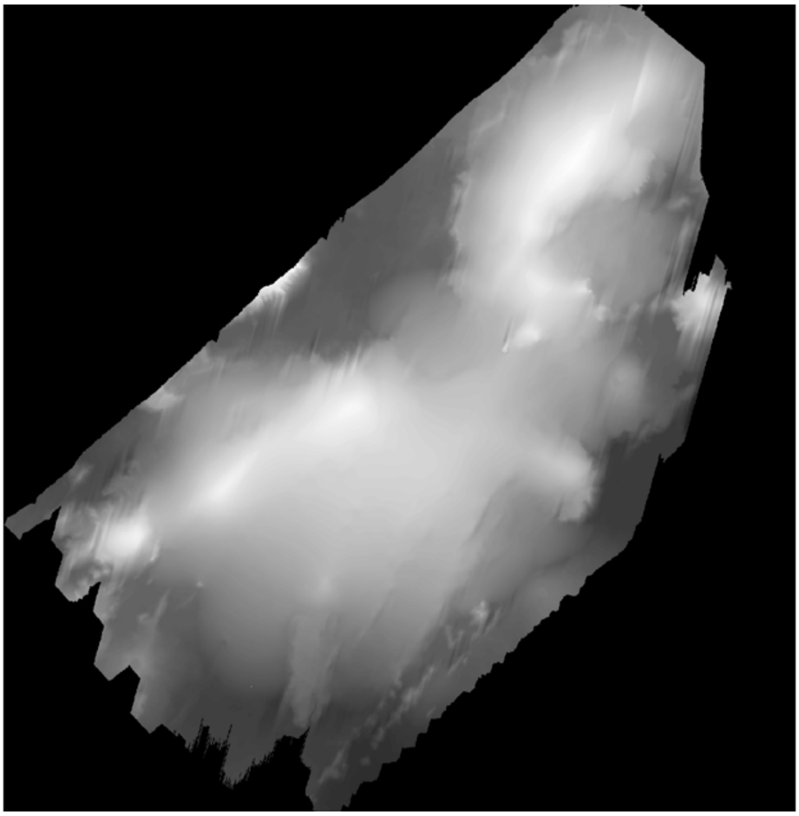


Figure 4.2: 2007 LiDAR v.2 DEM produced by Allen Pope (SPRI) utilising 2007 LiDAR data, 2001 skidoo data and photoclinometry of 2002 Landsat ETM+ band 4 imagery. Scale as figure 4.1.

1997 DEM

A DEM of the entire 1997 Langjökull surface and surrounding ice free topography was processed by Dr. Ian Willis (SPRI) from data provided by Finnur Pálsson and Helgi Björnsson (University of Iceland). The data was based on an extensive network of differential GPS snowmobile tracks and the 'kriging' method of interpolation was used to grid the data at 100m resolution and is presented in figure 4.3.

Figure 4.3: 1997 DEM and surroundings produced from differential GPS data collected via skidoo tracks. Courtesy of Helgi Björnsson and Finnur Pálsson with processing by Dr. Ian Willis. Scale as figure 4.1.

1986 and 2004 DEMs

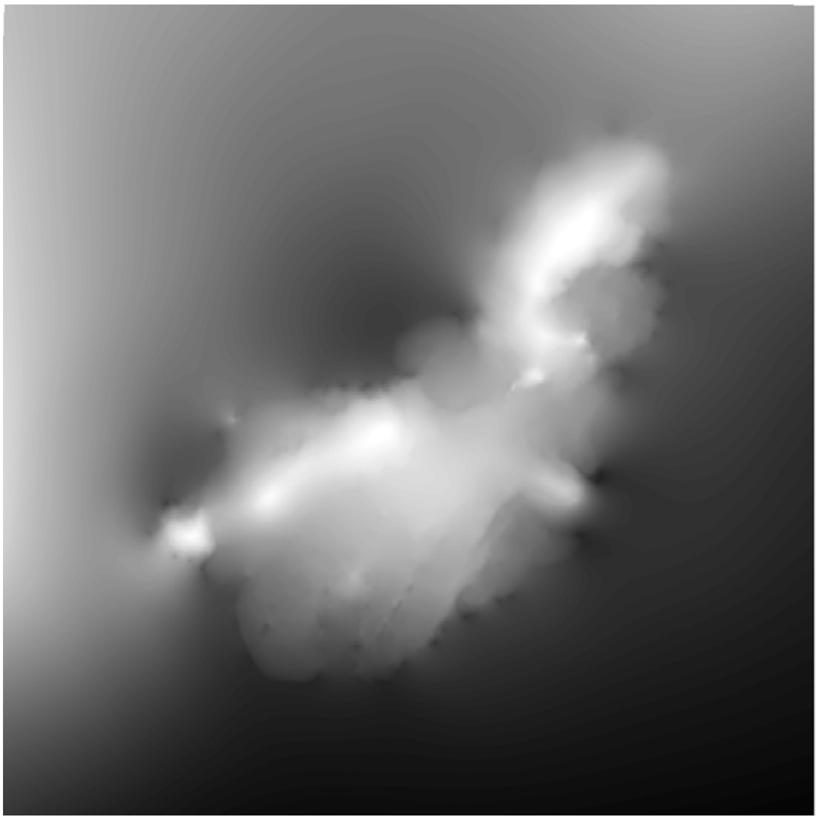


Figure 4.4: 1986 DEM created from photgrammetry.

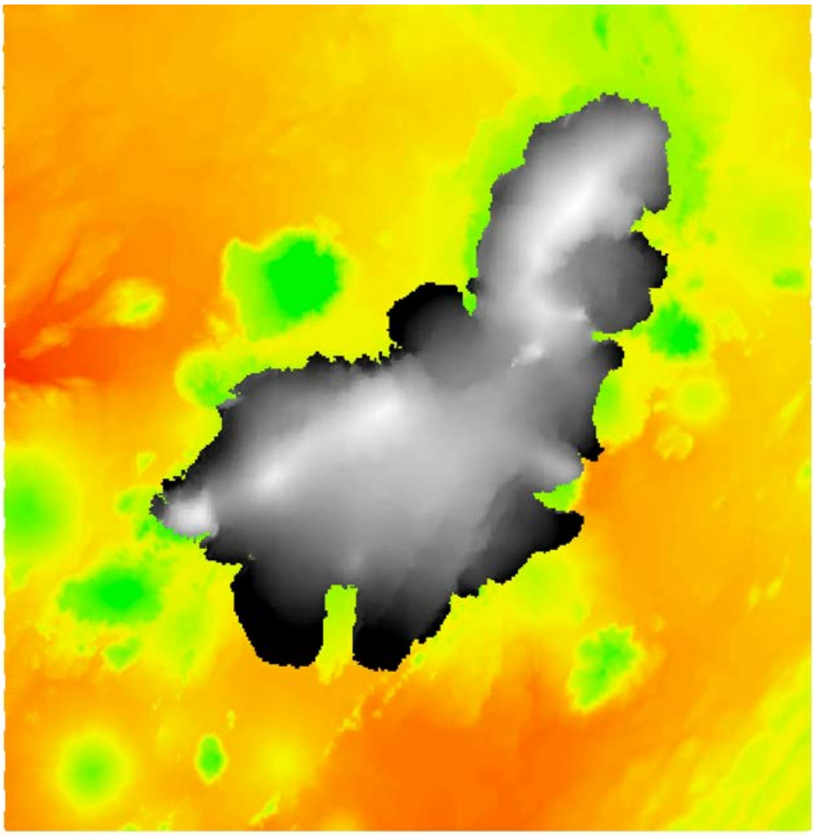


Figure 4.5: 2004 DEM overlain on the surrounding topography

Both the 1986 (figure 4.4) and 2004 (figure 4.5) DEMs were supplied courtesy of Helgi Björnsson and Finnur Pálsson at the University of Iceland. The 2004 DEM was constructed from SPOT 5 satelitte data with a spatial resolution of 40m. Accuracy is around 10 m in elevation and around 30 m in horizontal position. As acquired the data was gridded to a resolution of 170 m. The 1986 DEM was constructed by using photogrammetry and gridded to a resolution of 30m.

Subglacial topography DEM

The subglacial topography DEM (figure 4.6) was created from data collected via ground penetrating RaDAR (GPR). It was supplied courtesy of Helgi Björnsson and Finnur Pálsson, University of Iceland. It is gridded to a resolution of 400m.

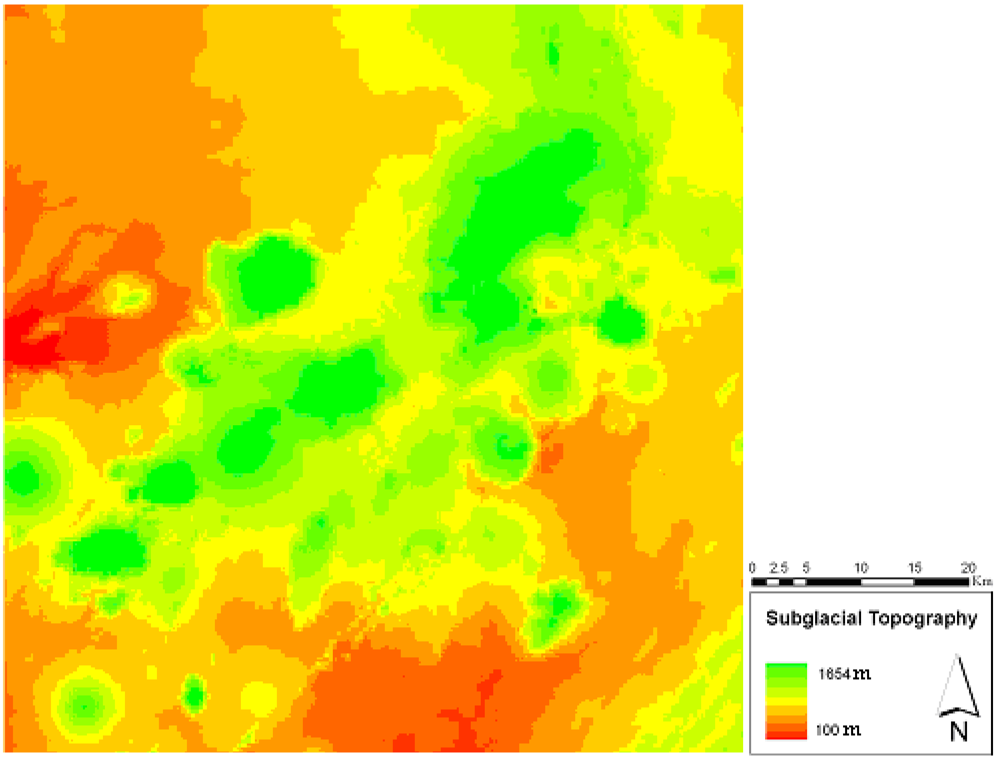


Figure 4.6: The subglacial and surrounding topography of the Langjökull ice cap

4.2 - Automatic weather station data (AWS)

Data from two automatic weather stations (AWS) located on the Langjökull ice cap was supplied by Finnur Pálsson (University of Iceland). The data is for 2006 at the locations L01 (490m) and L05 (1100m) as in figure 4.7. The data record the summer season; running from day 136 (16th May), to days 291 (18th October) for L01 and 325 (22nd November)for L05. The recording frequency was every 10 minutes. The data recorded included; temperature, relative humidity, solar radiation, albedo, long wave radiation, wind speed, wind direction and snow elevation.

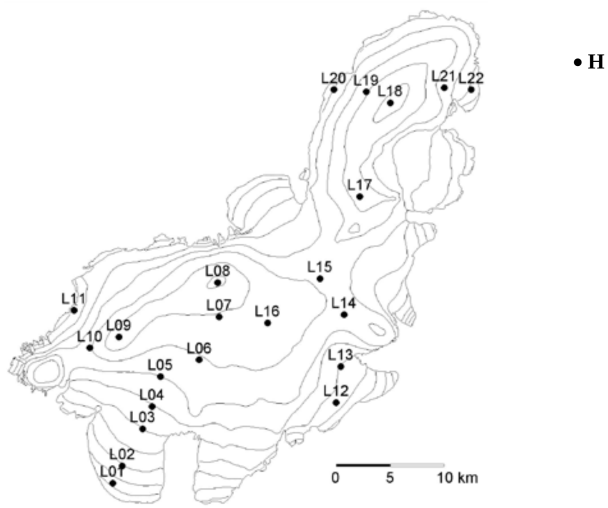


*Figure 4.7: Location of AWS stations on Langjökull. (Map courtesy of Finnur Palsson, University of Iceland). **H** marks the location of Hveravellir station.*

4.3 - Precipitation data

Precipitation data was supplied by Sverrir Guðmundsson of the Institute of Earth Sciences, University of Iceland. The data was from automatic weather station Hveravellir (H in figure 4.7). It is located between the Langjökull and Hofsjökull icecaps (64°52.005' N, 19°33.733'W) at around 650 m a.s.l. The data supplied was in mm d^{-1} format for the years 2005 and 2006. Due to the steep precipitation gradients found in Iceland, described in section 1, it was necessary to correct the data to altitude before it could be incorporated into the melt model (see section 5.3. Using the 30% per 100m precipitation gradient figure from de Woul *et al.*, (2006) the data was firstly regressed to sea level. It was then possible to calculate precipitation values for the automatic weather station altitudes - again using the 30% 100m precipitation increase to extrapolate.

4.4 - Mass Balance Data

Similarly, mass balance data was supplied by Finnur Pálsson of the University of Iceland. Data were available from 1997 to 2007. The mass balance was determined by a stratigraphic method through measuring changes in thickness and density relative to the summer surface. Ablation was monitored using snow stakes in the accumulation area while wires were drilled down in the ablation area. The summer balance was measured in the autumn. To measure the winter balance ice cores were drilled through the winter layer in the spring. The available data included;

- Elevation,
- Specific winter balance in m w.e. (metres water equivalent)
- Specific summer balance in m w.e.
- Specific annual balance in m w.e.
- Date of the autumn reading from stakes or wires the previous year
- Date of spring measurements and the date of autumn reading from stakes or wires.

Mass balance measurements are known to be variable from year to year: for example from drifting and redistribution of snow or predominant wind direction influencing precipitation accumulation. Sampling locations for mass balance follow outlet flowlines (for example L01-L08 in figure # represents the flow line of Hagafellsjökull Vestari). The readings from all points of the ice cap are combined and then extrapolated to give mass balance figures for the entire ice cap. Combining stake measurements from multiple outlets should theoretically account for both lateral and vertical variability in mass balance across the icecap. Error limits following integration are considered to be no lower than 15% (Björnsson & Pálsson, (2005). Personal communication)

This data was used to calculate the snow gradient for the ice cap to be used in the degree day melt model (section 5.3). The available total winter accumulation data from the weather stations across the glacier (in this case from L01-L08) was plotted against the altitude and the gradient which precipitation changed with progression up glacier was calculated. The mass balance data used was for the year 2006 as this is also the year for which the temperature and precipitation data are available. The calculated gradient for Langjökull was 4.45 mm m^{-1} (accumulation increasing from ~325 mm at 490 m a.s.l. to ~3650 mm at 1287 m a.s.l.).

5 - Methodology

5.1 - Ice surface topographic changes

After gathering and processing all of the DEMs and climate data into the correct format analysis began using primarily using ESRI's ArcGIS package. The initial analysis was to calculate the change in ice topography between the dates available. This was to provide a chronology of ice thickness. The ice margin was defined by the known extent for the various DEMs. Where this was not known (for example the 1986 does not have ice margins defined) available but where this was not possible the most appropriate margin was chosen. For example, the comparison for the 1997 DEM - 2007 LiDAR DEM was constrained by the 2004 margin. The rationale for this decision was that changes in the glacier margin are removed from the analysis. This aids clarity when assessing topographic changes. Although this method does not always allow advances in snout position due to surging to be measured it does, in the context of this study, avoid confusion. Ice mass transported down glacier during a surge is still easily detectable due to elevation increases around the margins. If required the snout position can be analysed separately although yearly snout position data is already well documented by Sigurðsson (1998) and also Pope *et al.*, (2009).

The differences between the DEMs, calculated for the various timescales, are presented in figures 6.1, 6.2, 6.3 and 6.4.

5.2 - Subglacial Hydrological System Reconstruction

Once changes in topography of the Langjökull ice cap had been assessed the next step was to analyse how these changes may have affected the hydrology of the subglacial system. As explained in section 2 the mass of ice is critical to the conditions at the ice/bed interface with a critical level of ice pressure suggested to be critical to surge initiation. In the case of a temperate glacier like Langjökull the switch from an efficient, channelised drainage system to an inefficient linked cavity system is the widely suggested mechanism that leads to surging of outlet glaciers.

The first step in this process is to calculate the hydraulic potential at every point beneath the ice cap. Utilising the DEM for each year and the DEM of the subglacial topography this is possible using equation 2, below:

Equation 2: $\mathbf{\Phi = Es - E\beta * \rho_i * g + (E\beta * \rho_w * g)}$

Where;

Φ = Hydraulic potential in Pascals

Es = Surface Elevation

$E\beta$ = Bed Elevation

ρ_i = Density of ice (900 kg m^{-3})

ρ_w = Density of water (1000 kg m^{-3})

g = Gravitational constant (9.81 m s^{-2})

Once the hydraulic potential across the entire ice cap was calculated it was necessary to evaluate any potential sinks within the glacier. In terms of the DEM this would constitute a cell with a value sufficiently low to allow all surrounding cells to flow into it. Once these sinks are identified it is necessary to fill them to a minimum level to allow the flow to move outwards to the lowest surrounding cell. For example see figure 5.1:

Figure 5.1: A visualisation of a potential sink being filled. On the left all cells flow in to the lowest cell. On the right the lowest cell is filled to a level to allow it to flow outwards to the lowest surrounding cell

With the potential sinks filled the direction of flow is then considered. The filled potential DEM was analysed so each cell finds the lowest surrounding cell - replicating the flow down the hydraulic potential gradient. The accumulation of flow can then be added to this by analysing how many upstream cells flow into the next cell. A visual interpretation of the flow is produced from this representing the direction of flow and also the total amount of flux through the cells – often resulting in a branched system as would be intuitively expected.

5.3 - Degree Day Melt Model

The subglacial hydrological system is the key element responsible for surging. Once the main flow pathways are calculated (as section 5.2) it was necessary to calculate values of water flux through the system. This makes it possible to assess how changes in the flux and the surface topography may interact to result in a surge of the glacier (or conversely to delay a surge). In order to evaluate the potential water flux through the Langjökull subglacial hydrological system a distributed glacier surface degree day model was used and was implemented using the MATLAB software package. The code for the model was originally written by Cameron Rye of the Scott Polar Research Institute and has been adapted from a study in Norway.

The model involves numerous input data which were taken from existing literature or, where unavailable or specific to Langjökull, calculated from other available data. Degree day factors are input for a single point on the glacier (the location of an AWS) and melt is then distributed over the glacier using a DEM. The following data were input into the model:

- *AWS data*: this included the Julian day, the average air temperature and the daily average precipitation data.
- *Ice density*: For ice sheet modelling purposes the density of ice is assumed to be **900 km m^3**. Previous studies have also used this value specifically for Langjökull (Flowers *et al.,* 2007, Eyre *et al.,* 2005).
- *Snow density:* Snow density in Iceland has been found to be relatively independent of altitude (Jóhannesson *et al.,* 1998). The density of snow at the end of the accumulation season is, of course, variable. Lundberg *et al.,* (2000) suggest a typical value of 350 kg m^3 as do Björnsson & Pálsson (2010, Personal communication) when modelling on Langjokull. However, Jóhannesson *et al.,* (1998) noted mid winter snowpack density values in Iceland to be higher than other noted values such as from the European Alps. They measured values of around 400-450kg m^{-3}, similar to Norwegian values. Due to this disagreement between studies a value of **400 kg m^3** is assumed.

- *Temperature lapse rate:* in a previous study using degree day models on Langjökull Guðmundsson *et al.,* (2003) give the figure **0.0062°C m^{-1}** and as this is relatively constant is utilised in this study.
- *Precipitation gradient*: The precipitation characteristics of Icelandic ice caps are controlled by a maritime climate with the highest precipitation found towards the south (see section 1). Precipitation on Langjökull is similar to that of neighbouring Hofsjökull. (De Woul *et al.,* 2006) justify a precipitation gradient of **30% per 100m** when modelling melt of Hofsjökull therefore this gradient is assumed here.
- *Precipitation threshold:* It is necessary to consider whether precipitation is rain or snow. Previous studies in Iceland (de Woul *et al.*, 2006) have assumed a mixture of snow and rain is found in a transition from ranging from -1 °C to to +1 °C. Using this as a guide a precipitation ratio of 50% snow and 50% rain at +1°C is expected which would potentially lead to some accumulation. Consequently the value is set at **+1°C** for this study.
- *Snow gradient:* As described previously (section 4.4) the gradient with which precipitation changed with progression up glacier was calculated from the mass balance change related to altitude. The calculated figure is **4.45 mm m^{-1}**.
- *Degree Day Factors (ice and snow):* Guðmundsson *et al.,* (2003) calculate several degree day factors (ddf) linking the summer balance to the weather parameters and surface albedo via different calculations. They are calculated for two weather stations on Langjökull – 490m and 1090m. Four empirical models are used to calculate the degree day factors with various successes at modelling; one being discounted as it modelled the ablation poorly. The model most suited to this study gives values of **5.3mm °C^{-1} d^{-1} for snow** and **6.0mm °C^{-1} d^{-1} for ice**. Guðmundsson *et al.,* (2003: 12) describe this model as coming 'nearer to depending solely on conditions at the glacier surface'. It also varies more gradually and is less sensitive to changes in weather parameters than other models and better represented the melt through the early part of the melt season (May) when net radiation, strong winds and relatively high temperatures combine. The values selected here are also within 1 mm °C^{-1} d^{-1} to those used when modelling melt in the later part of the 20th century on Langjökull by Flowers *et al.,* (2007).

5.3.1 - Flux of melt water through system

Using the model to calculate the total melt across the ice cap surface during the ablation season it is then important to consider how much of this water enters the subglacial system. Flowers *et al.*, (2007) estimate a total input of meltwater of around $80m^{-3}s^{-1}$ from the glacier. They simulate hydrogeologic parameters beneath Langjökull and suggest that 69% of melt water from the Langjökull ice cap is transported through the groundwater system with the remainder transported subglacially. Although studies are sparse a similar figure of groundwater transport (70%) is given by Sigurðsson's (1990) geochemical estimate. The study estimates a groundwater recharge figure 50-80 m^3 s^{-1}. Therefore in this study 30% of the calculated melt will be assumed to be routed through the subglacial system at the ice/bed interface. It is also assumed that the meltwater enters the system immediately via moulins and is does not flow over the glacier surface. Figure 6.5 shows the final DEM of melt for the summer of 2006.

6 – Description of results

6.1 – Topographic change

1986 – 1997 (Figure 6.1)

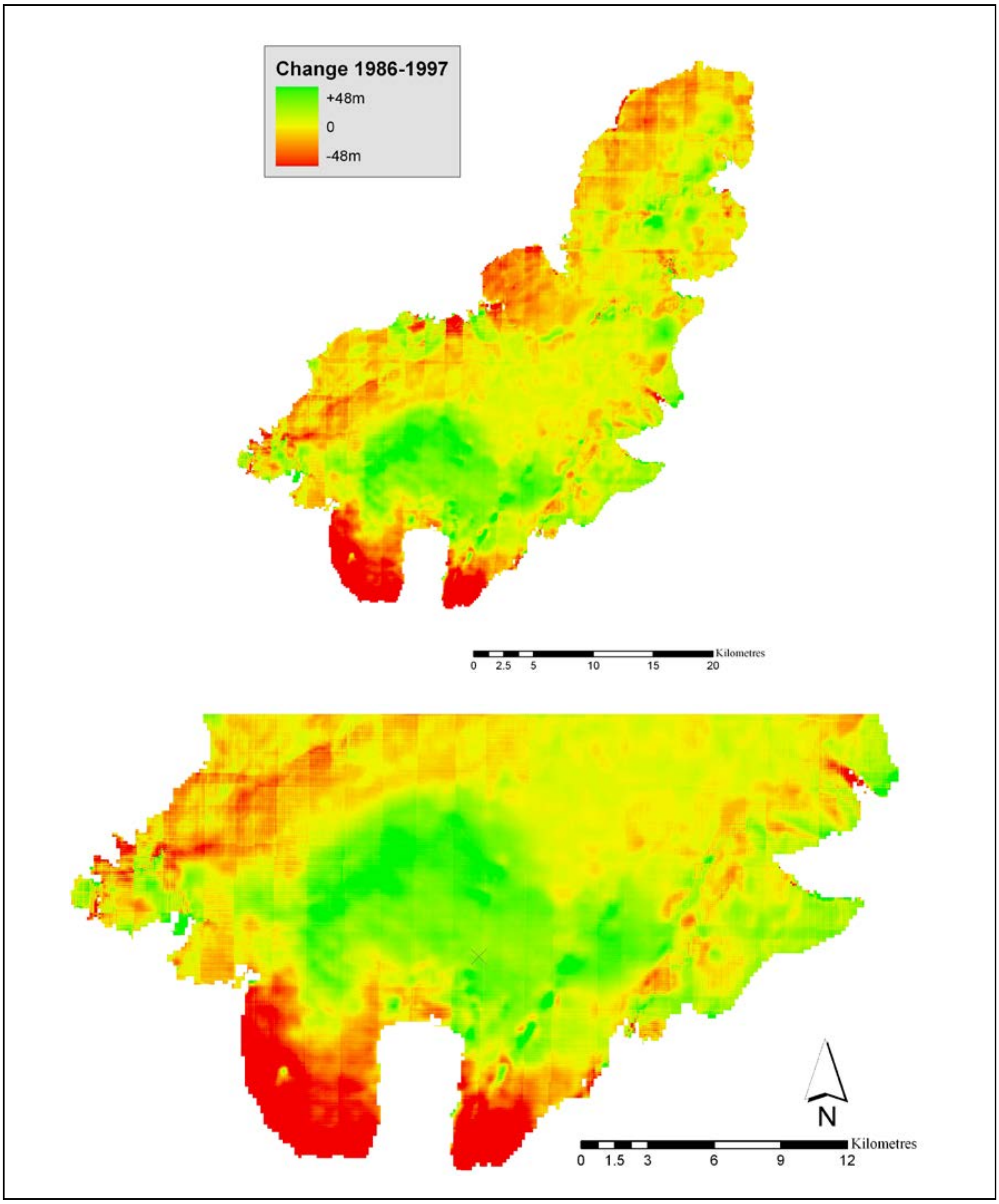


Figure 6.1: Topographic change of the Langjökull ice cap (top) and the southern outlets (bottom) from 1986 – 1997.

Topographic changes of the Hagafellsjökull outlet glaciers during the period 1986 to 1997 reflect outlets that are in the quiescent phase and are building up to a surge. The ablation areas of the two outlets are clearly and strongly identified as loosing mass by the red colouring. The amount of ablation in these areas is generally >40m with a gradual progression up glacier seeing a reducing rate of ablation leading into a brief area of equilibrium before a considerable area of accumulation is reached at *c*. 900 -1000 m a.s.l (matching the equilibrium line altitude of Langjökull given by Eyre *et al.,* (2005). The area that accumulated mass (c. 10 km^2) shows a gain of around 20-50 m and is distributed all across the upper reaches of both Hagafellsjökull Eystri and Hagafellsjökull Vestari (for comparison glacier divides are shown in figure 1.4).

1997 – 2004 and 1997-2007 (figures 6.2 and 6.3)

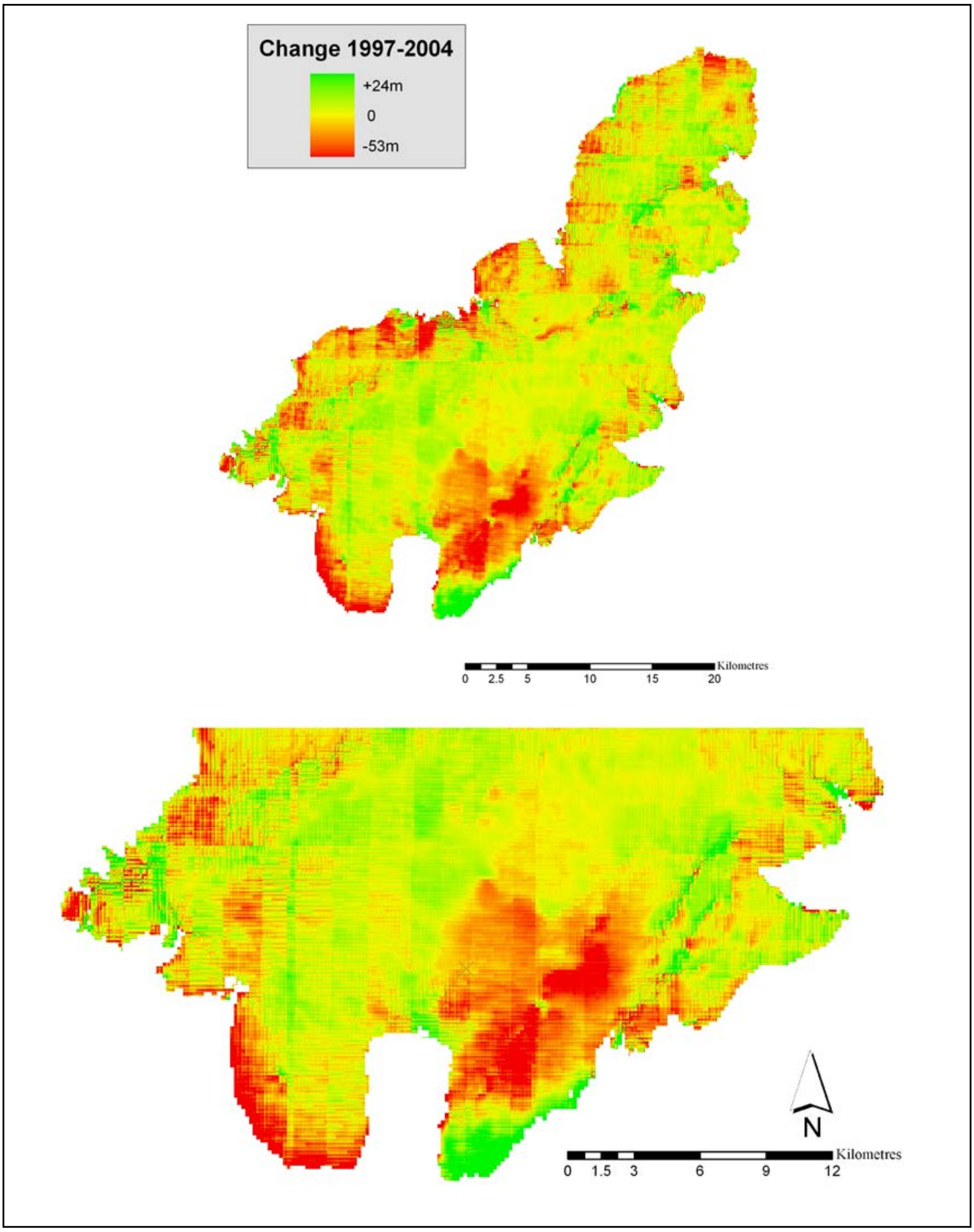


Figure 6.2: Topographic change of the Langjökull ice cap (top) and the southern outlets (bottom) from 1997 – 2004.

Following the surge of Hagafellsjökull Eystri in 1998 the topography underwent some considerable change. Ideally a DEM from 1999/2000 would allow the yearly change to be analysed but unfortunately this is not available. However, the changes are still very evident when compared with the 2004 data. Figure 6.2 shows considerable losses in the upper reaches of Hagafellsjökull Eystri with generally c.40m of surface lowering. Conversely there are strong gains (the green section) in ice surface elevation in the lowest reaches (generally in the mid 20s of metres). It is worth considering that the advance of the terminus is not visible as the ice margin does not take this into account as explained previously. To quantify this advance: the 1998/1999 surge gave an advance of around 1100m which will have extended the glacier margin somewhat. Similarly this figure does not take into account the ablation during the c. 6 year period between DEMs so the initial surface height increase would have been greater still. Pope *et al.,* (2009) calculate a figure of -2.28 m yr^{-1} w.e. for the period 1997 – 2001.

During the same period Hagafellsjökull Vestari demonstrates very different patterns. The opposing colouring to its neighbouring outlet shows Vestari continued to loose mass in the ablation area at similar levels to the previous comparison. The upper reaches do not generally show any ice surface lowering – instead they border between stagnant and moderate gain. Figure 6.3 also shows this pattern continuing to 2007 with some gain in surface height.

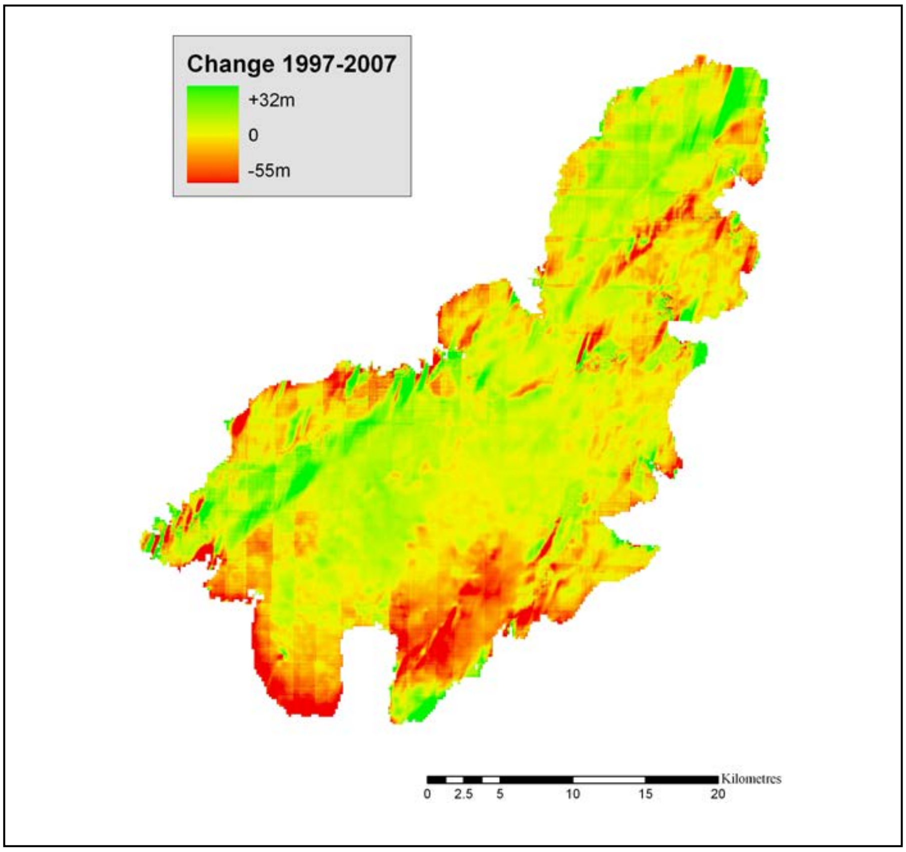


Figure 6.3: Topographic change of the Langjökull ice cap from 1997 – 2007.

2004-2007 (Figure 6.4)

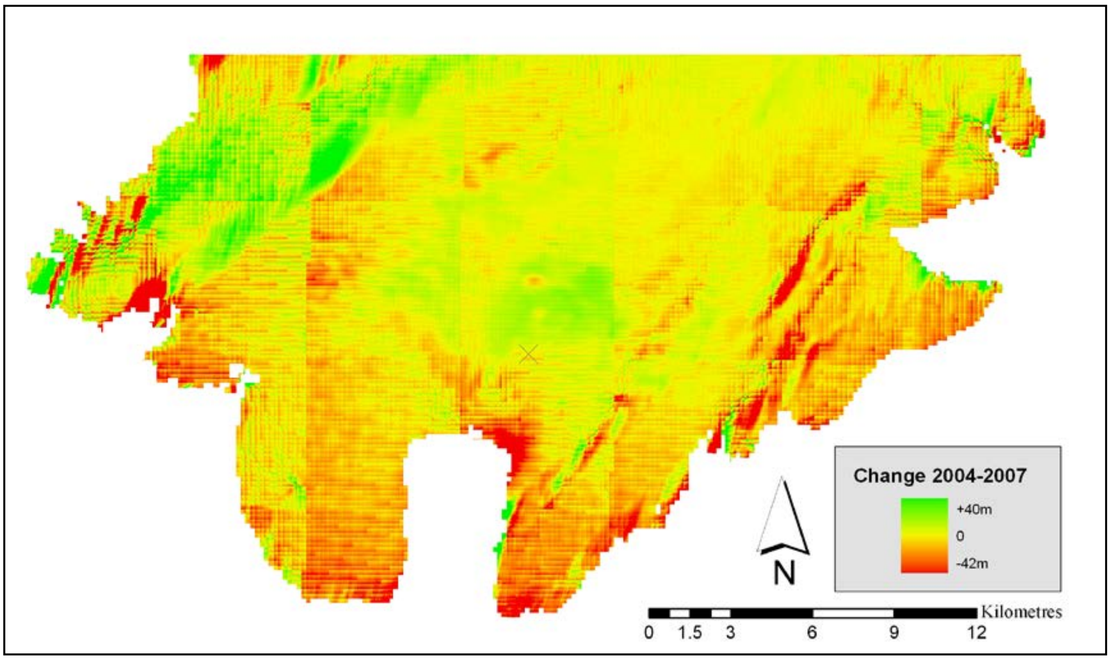


Figure 6.4: Topographic change of the southern outlets from 2004 – 2007.

The 2004-2007 comparison of surface height show the outlets in what should be the quiescent phase. Indeed, the topography over this short three year period shows signs of a pattern of topography change that may well develop to be similar to that seen in figure (86-97). Accumulation is apparent in the upper areas of both outlets and ablation is predominant in the lower reaches as expected during normal ice flow conditions in the quiescent phase.

6.2 – Surface melt model

The degree day surface melt model described in section 5.3 for run using the acquired data from 2006. The modelled melt is displayed in figure 6.5.

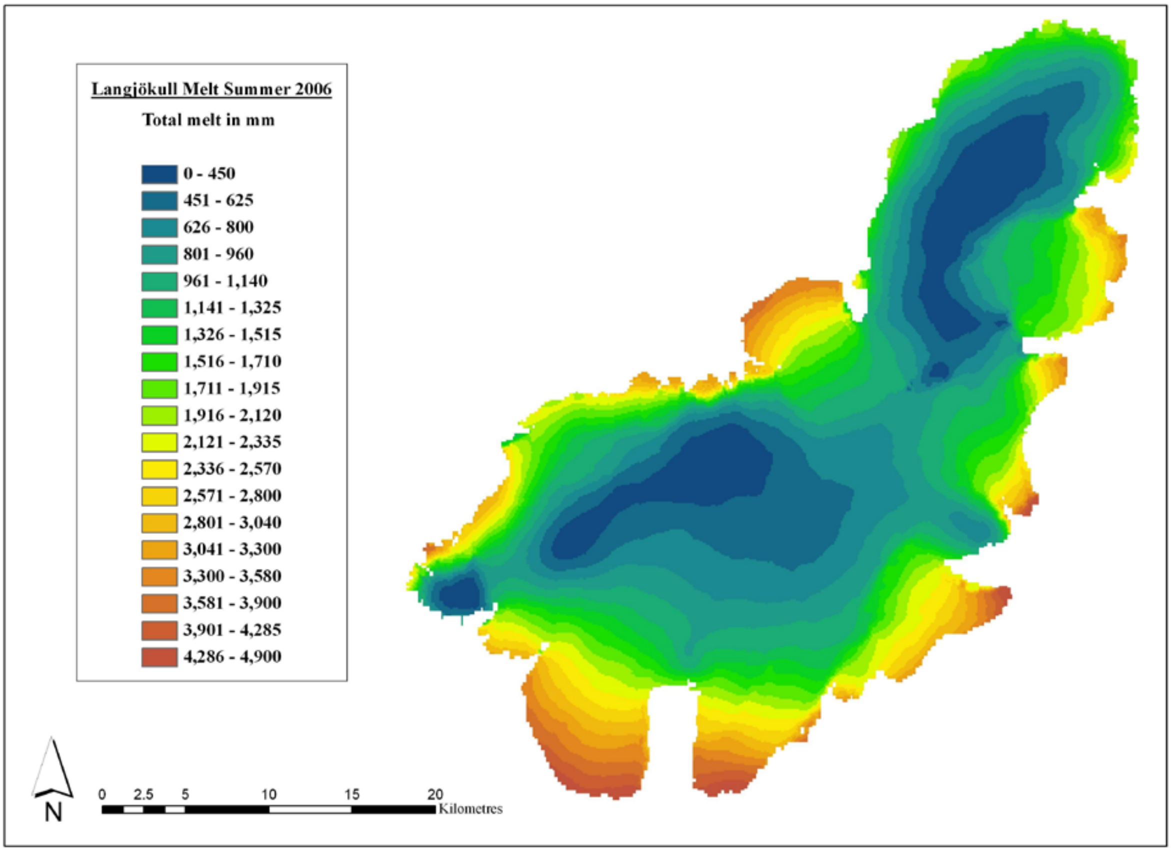


Figure 6.5: Modelled melt for the degree day model. Values show the total calculated summer melt for the year 2006

The melt model shows good relation to the areas of greatest ablation as identified in this chapter. Hagafellsjökull Eystri and Vestari show the highest levels of melt with maximums almost reaching 5 m.w.e a^{-1}. The lowest levels of melt are the highest elevations in the west and north-east. These areas do not correspond entirely with the greatest areas of topographic change (for example those shown in figure 6.1). The greatest areas of elevation increase in the Hagafellsjökull accumulation areas is south of the highest elevations of the ice cap.

6.3 – Subglacial Hydrological reconstruction

Following the reconstruction of the hydrological system the meltwater DEM was routed through the subglacial system. This gave each grid square in the hydrological system values of water flux, in mm, over the entire melt period. For flow comparisons in was, of course, preferable to convert this value into a more useable value of $m^3 s^{-1}$. However, for the purposes of display and comparison much better results are obtained using by displaying the total melt period values. Equivalent values in $m^3 s^{-1}$ are given as a guide where necessary. After processing the flow accumulation the following figures 6.6 to 6.9 were produced for each of the DEMs from 1986 – 2007.

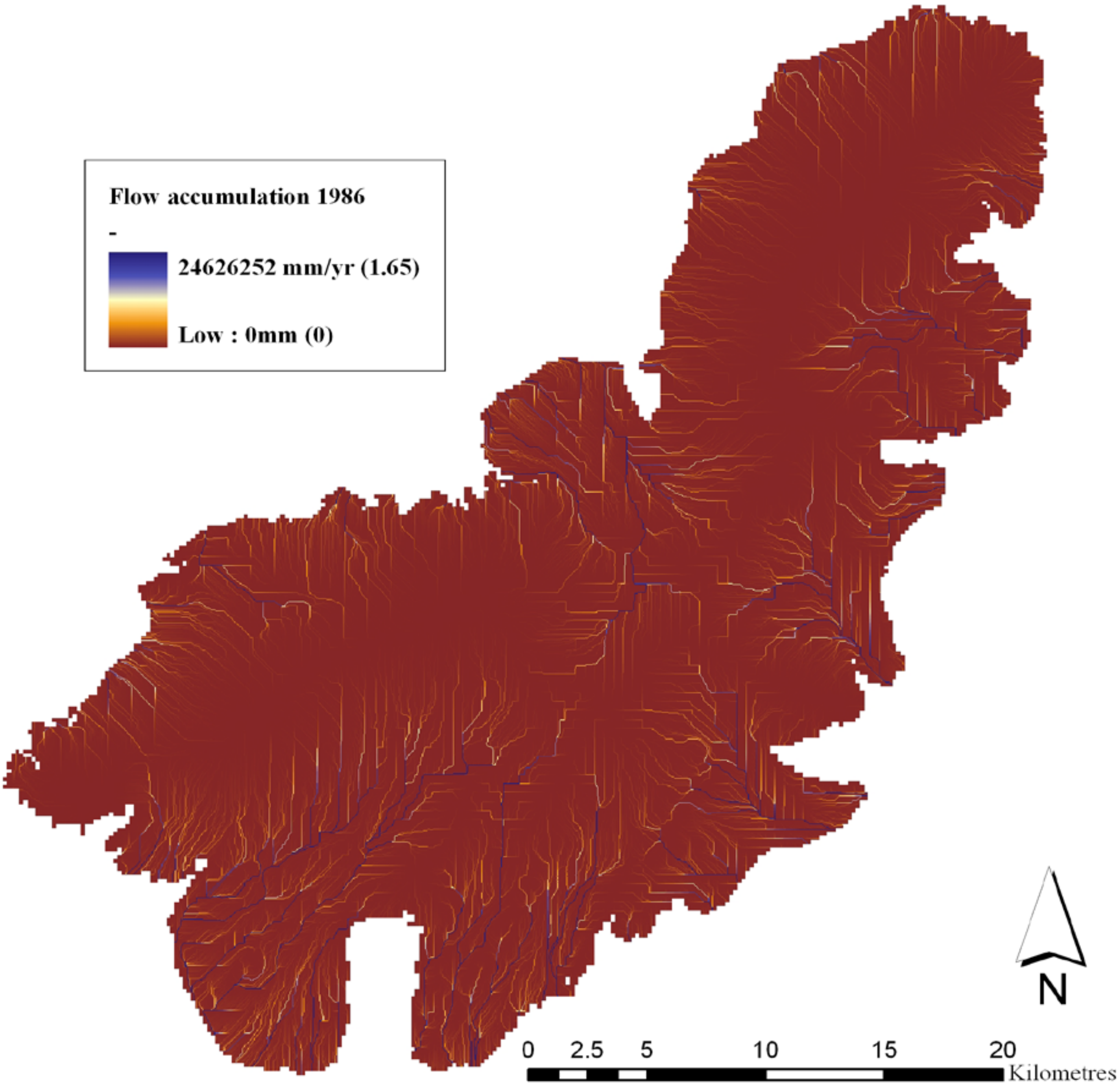


Figure 6.6: Subglacial hydrological reconstruction 1986. Bracketed values in $m^3 yr^{-1}$.

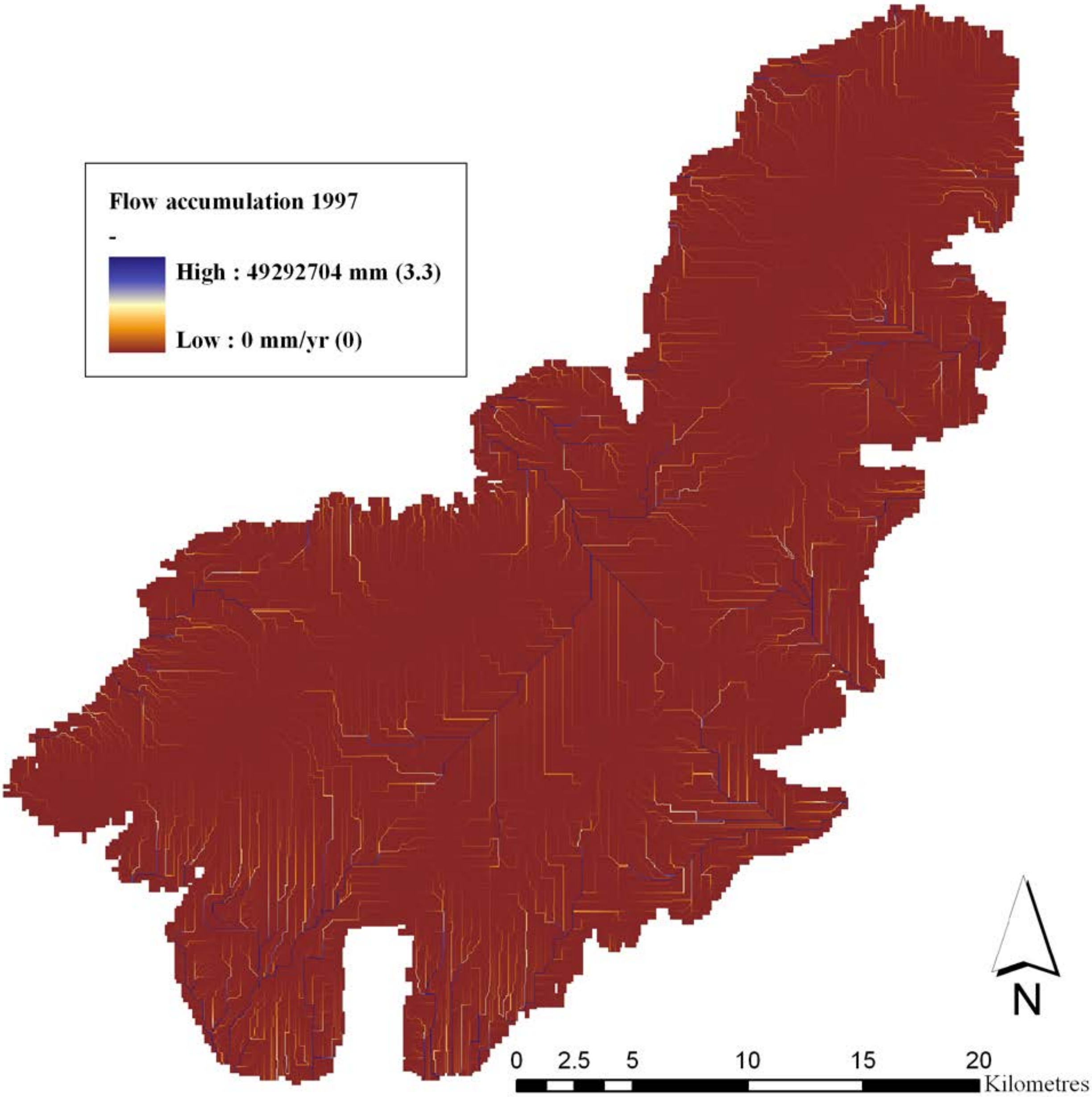


Figure 6.7: Subglacial hydrological reconstruction 1997. Bracketed values in $m^3\ yr^{-1}$.

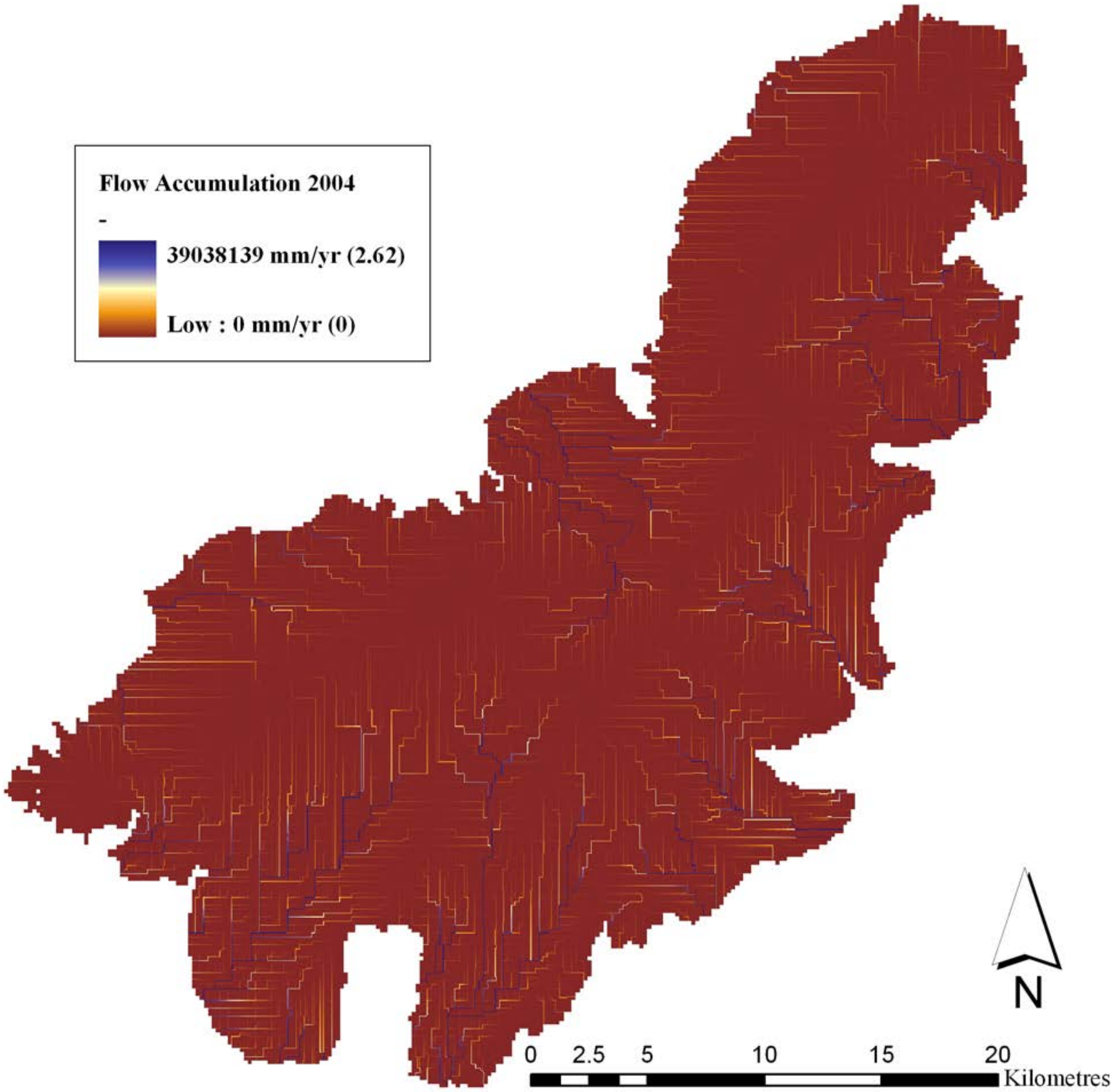


Figure 6.8: Subglacial hydrological reconstruction 2004. Bracketed values in m^3 yr^{-1}.

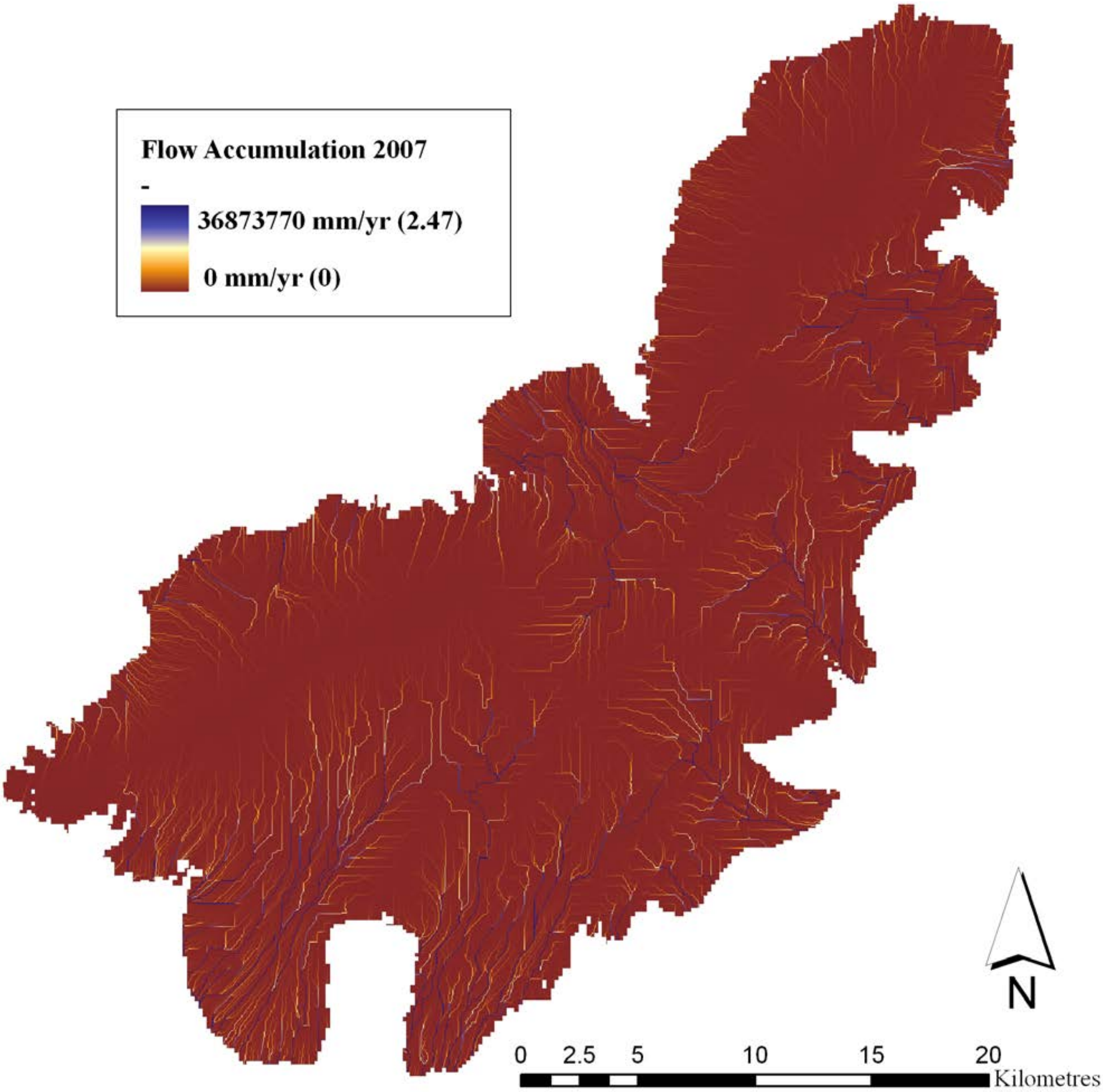


Figure 6.9: Subglacial hydrological reconstruction 2007. Bracketed values in m^3 yr^{-1}.

Figures 6.6 to 6.9 show that widely speaking the subglacial hydrological system of the ice cap has remained relatively consistent in most areas. However, as intuitively expected, changes are apparent in the Hagafellsjökull outlets – particularly in the accumulation area between 1986 and 1997. These changes are described in the following section.

6.4: Description of changes: 1986 – 1997

Following subglacial hydrological system reconstruction of Langjokull for both 1986 and 1997 comparison reveals a mixture of changes. Parts of the system remain very similar – for example the modelled hydrological system beneath Noðurjökull (figure 6.10) shows little change with a good agreement in drainage area and drainage pathway size and position. Similarly, Flosaskardsjöklar (top, figure 6.11) shows an almost overlapping system accumulating meltwater from a similar sized catchment area. Main outlets are in generally matching positions for both 1986 and 1997. There are, however, some particularly pronounced changes.

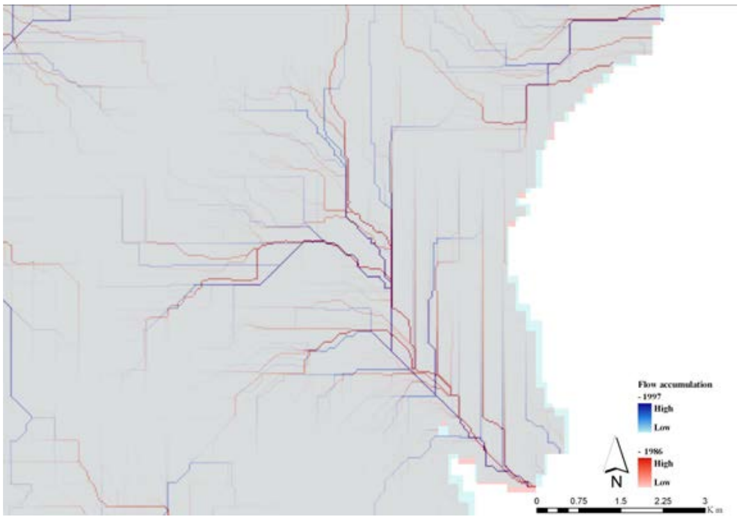


Fig 6.10: Agreement of modelled hydrology of Noðurjökull for 1986 and 1997.

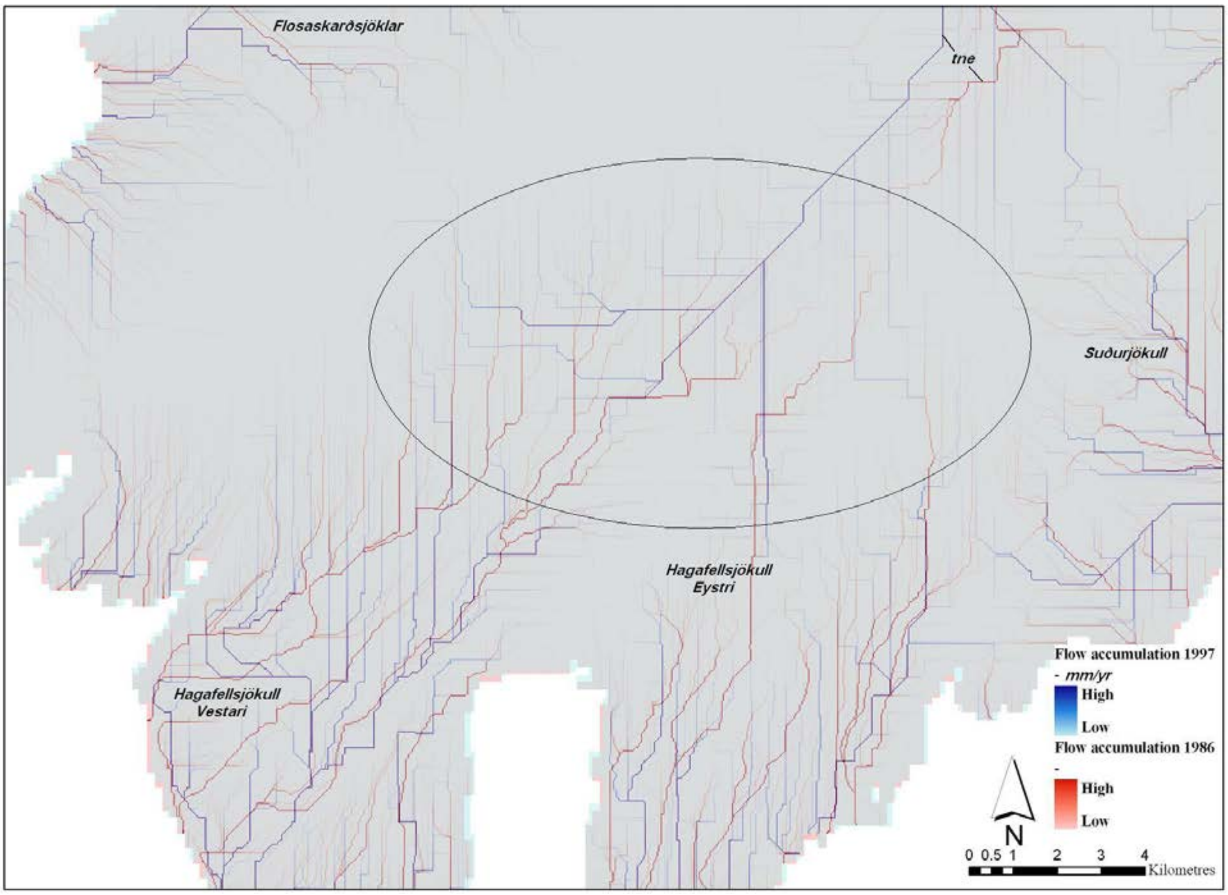


Figure 6.11: A composite of the 1986 and 1997 modelled hydrological systems (southern outlets shown)

Figure 6.11 shows the upper reaches of the southern Langjökull outlets. It is a composite image with the modelled accumulation from both 1986 and 1997 overlain. There is considerable overlap between the 1997 system (blue) and the 1986 system (red) in the circled area and the direction of flow is reversed. The flow of meltwater to the north-east appears to have extended south west by *c.* 6-7 km in places. The effect of this is to have reduced the size of the drainage catchment to the Hagafellsjökull glaciers considerably. The model shows meltwater instead being directed initially north-east before joining the catchment of Þrístapajökull (see figure 1.4) and flowing out towards the north-west.

In terms of estimates of meltwater flow the accumulated value from the tributary draining the circled area to the north-east (labelled 'tne') increases from approximately 0.2 m^3s^{-1} in 1986 to 1.4 m^3s^{-1} in 1997 – a considerable increase of 700% the original value.

6.4.1 - Hagafellsjökull Eystri

The modelled subglacial hydrological system of Hagafellsjökull Eystri appears to have a similar level number and relatively similar pattern of drainage pathways. However, it is apparent that in figure 6.12 that the catchment area of the outlet is somewhat less because the pathways are shorter by several kilometres. As described,

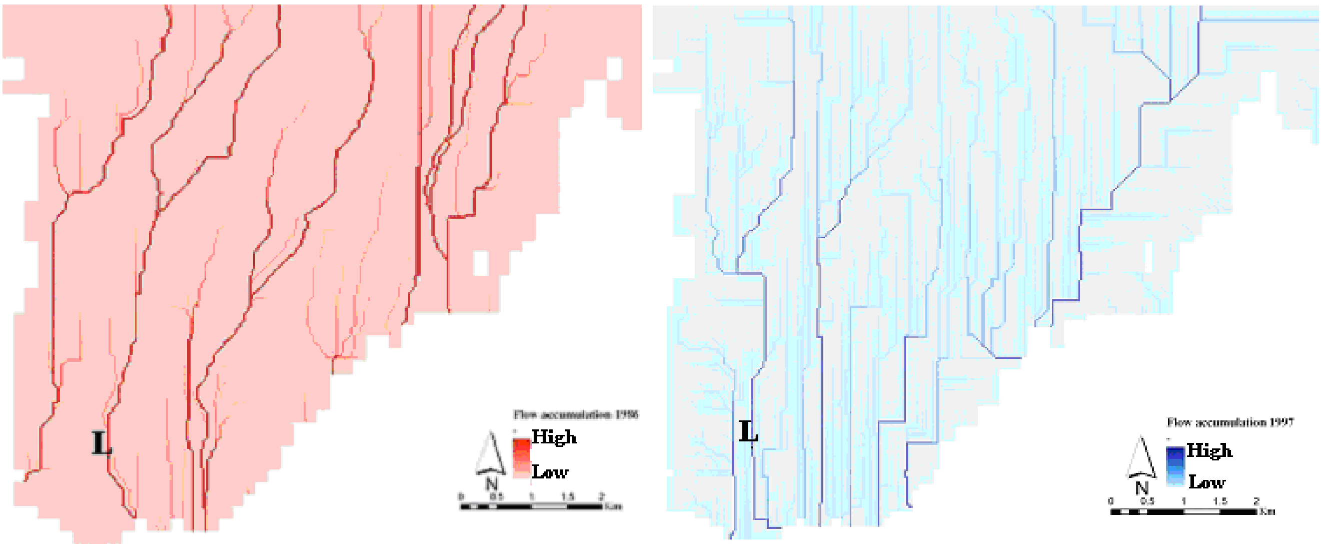


Figure 6.12: Comparison of the 1986 and 1997 hydrological systems of Hagafellsjökull Eystri.

more flow appears to be directed towards the north-east in 1997 compared to 1986. The accumulation of flow lost from the catchment of Hagafellsjökull Eystri is well visualised in figure # (on the following page). This affects practically the entire system and is reflected in the modelled values of accumulated flow from the pathways. For example – at point 'L', which marks one of the main pathways flowing beneath Eystri in both 1986 and 1997, modelled flow is reduced from $0.7 m^3s^{-1}$ to $0.4 m^3s^{-1}$.

6.4.2 - Hagafellsjökull Vestari

Similarly, Hagafellsjokull Vestari exhibits a modelled reduction in its catchment area when comparing the 1986 model to the 1997 model.

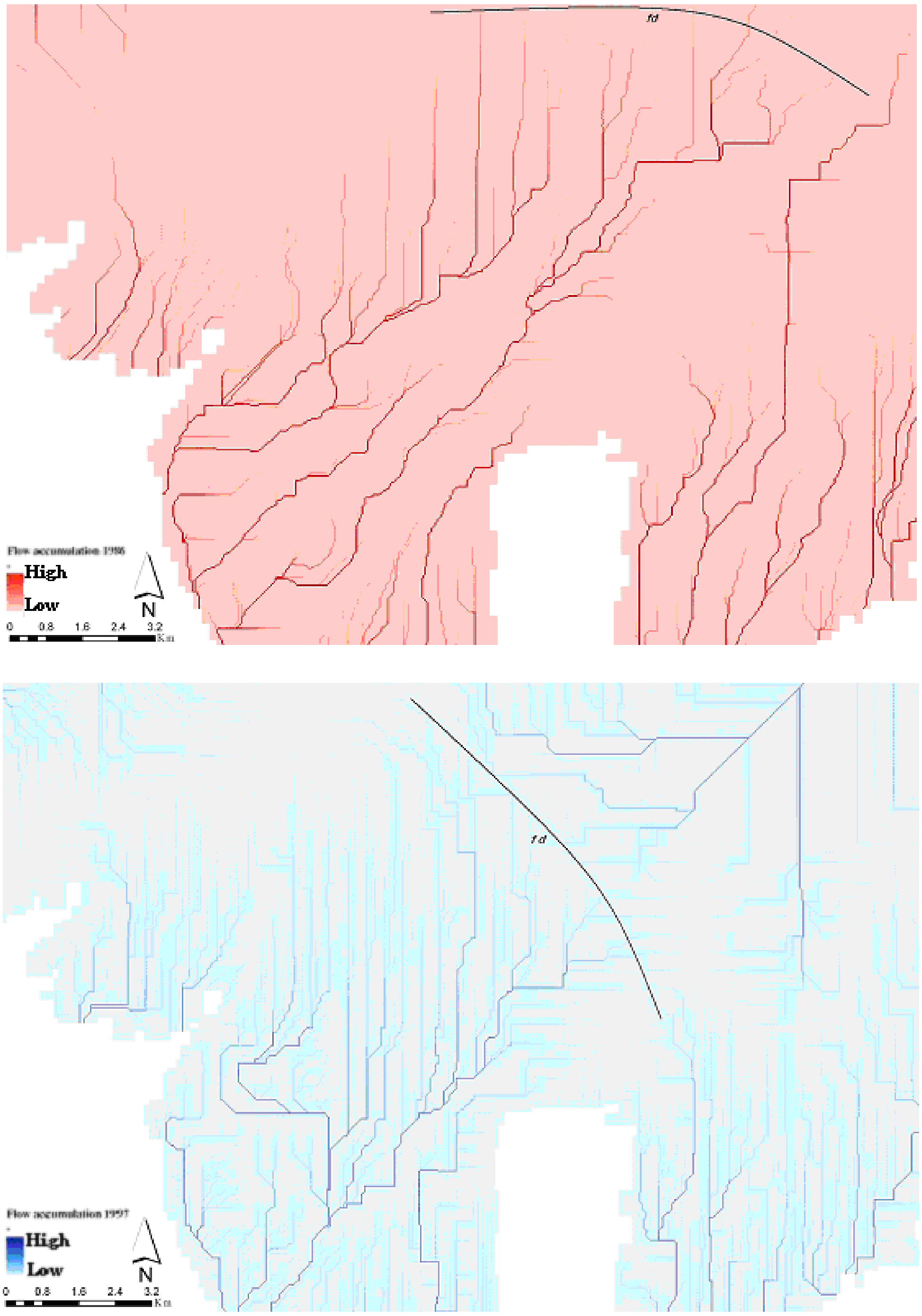
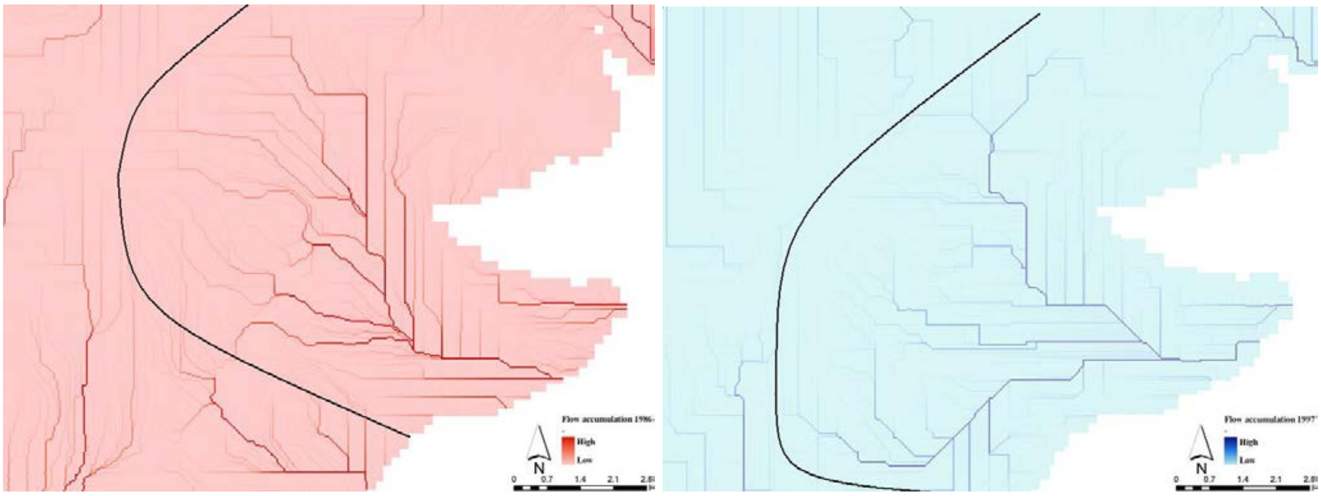


Figure 6.13: The changing flow of the upper Hagafellsjökull area from 1986 (top) to 1997(bottom)

The black lines labelled 'fd' in figure 6.13 give a basic comparison of how the apparent flow divide between the south-west and north east is modelled to alter. As with Hagafellsjökull Eystri this results in the drainage pathways becoming shorter by several kilometres. Worthy of note at this point is the change in the pattern of drainage pathways. As described in the previous section the pathways of Hagafellsjökull Eystri become shorter in 1997 but maintain a similar pattern with a comparable number of pathways. Hagafellsjökull Vestari shows similar levels of pathway shortening but also shows some considerable differences in the form of the drainage system. The modelled 1986 system shows numerous (*c.* 6-7) main drainage pathways reaching the edge of the outlet. However, in the 1997 model many of these pathways coalesce in the centre of the outlet to form a branched system with one main outlet. As a result comparison of the modelled flow values is somewhat misleading. The largest pathway in 1986 has a calculated discharge of approximately 1.4 $m^3 s^{-1}$ whereas in 1997 it is approximately 2.2 $m^3 s^{-1}$ - despite the drainage catchment in 1986 being somewhat larger. Overall the total out flow of meltwater from beneath Hagafellsjökull Vestari through the series of smaller, individual systems in 1986 is a higher value compared to total for 1997.

6.4.3 - Suðurjökull

Most of the subglacial drainage beneath the other outlets of Langjökull exhibit similar patterns and levels of flow in comparison between 1986 and 1997. One exception to this is the neighbouring Suðurjökull outlet, located to the north-east of Hagafellsjökull Eystri (figure 1.4). As a result of the modelled increase in flow to the north-east the upper reaches of Suðurjökull see some increase in the modelled catchment area - the black line in figure 6.14 demonstrate approximately where the flow divide is located. The pattern of flow - numerous tributaries feeding a single main outlet - exhibits little change. One key change is apparent however. Due a modelled increase in the catchment area the volume of flow increases. Figure 6.14 shows the flow divide of Suðurjökull. In 1986 the catchment area stays broadly within this area. However, in 1997 an increased amount of flow is modelled as being collected by Suðurjökull from the south-west. The area is not defined as an outlet; it forms an unnamed area between Suðurjökull and Hagafellsjökull Eystri confined by topography. In 1997 this means that despite the model showing a smaller catchment in the upper reaches compared to 1986 the overall output from the main outlet does increase. The calculated output figures are 1.4 $m^3 s^{-1}$ and 2.3 $m^3 s^{-1}$ for 1997 and 1986 respectively.



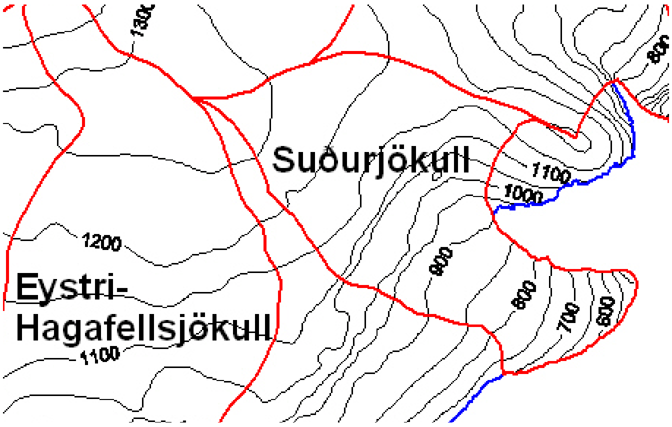


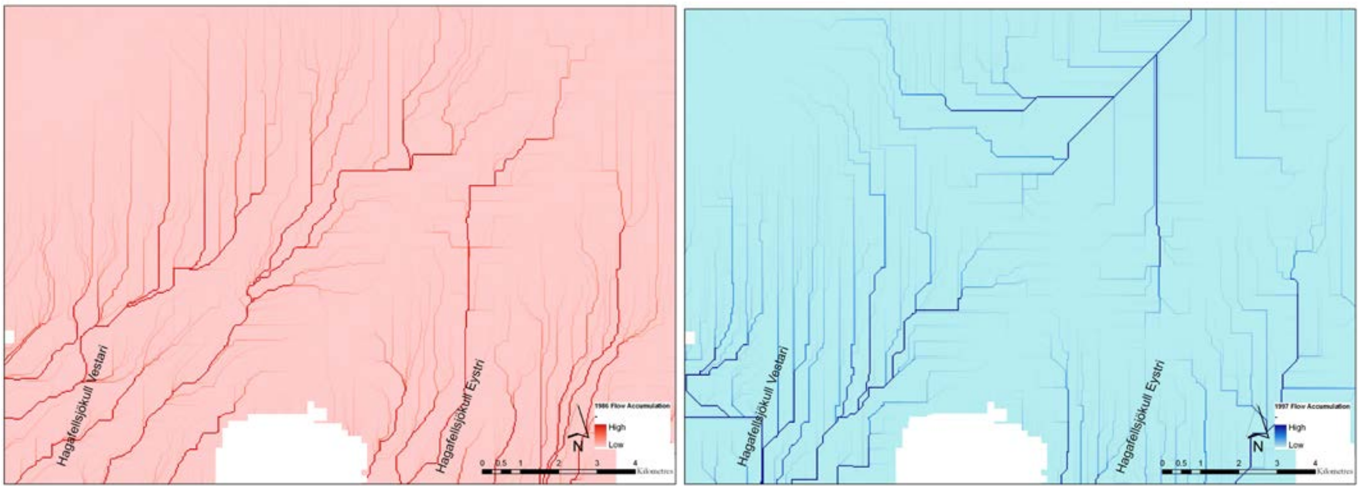
Figure 6.14: Comparison of Sudurjökull and the surrounding area. Clockwise from left: 1: Sudurjökull flow margins, 2: 1986 modelled subglacial hydrology and 3: 1997 subglacial hydrology. Black lines in 2 & 3 indicate approximate flow divides.

6.5 - Post 1998 surge changes

The 1997 DEM represents a model of Langjökull in a stage now known to be immediately prior to a surge of Hagafellsjökull Eystri in 1998. As expected during a glacial surge this rapid flow led to rapid transfer of accumulated ice mass from the upper area of Hagfellsjokull Eystri to the ablation area. This increased the ablation area altitude by around 25 m when the 1997 and 2004 DEMs are compared (figure 6.2). Following the surge the glacier returned to a similar pattern to that prior to the surge. Surface lowering was once again predominant in the ablation area and comparison between the 2004 and 2007 DEMs showed accumulation was again causing moderate increases in surface elevation in the upper reaches (Figure 6.4).

6.5.1 - Post 1998 surge changes: Hagafellsjökull Eystri and Vestari

Comparisons of the modelled subglacial drainage from 1997 and 2004/2007 again show many areas of little change. Conversely, they also demonstrate some areas with change as marked as the period 1997 – 1986. Perhaps unsurprisingly, given the large amount of change in topography instigated by the Hagafellsjökull Eystri surge, the key area is once again the upper reaches of the Hagafellsjökull outlets. The modelled change in topography in 2004 returns the hydrological system to, what appears to be, a similar arrangement to that modelled for 1986. The percentage of flow 'lost' to Þrístapajökull flows back to the Hagafellsjökull outlets. This is evident in figure 6.15. These changes reduced the flow to the Þrístapajökull branch (labelled 'tne' in figure 6.11) back to volumes comparable to 1986. The 1986 model calculated *c.* 0.2 $m^3 s^{-1}$ and the 2004 calculation estimates *c.* 0.24 $m^{-3} s^{-1}$.



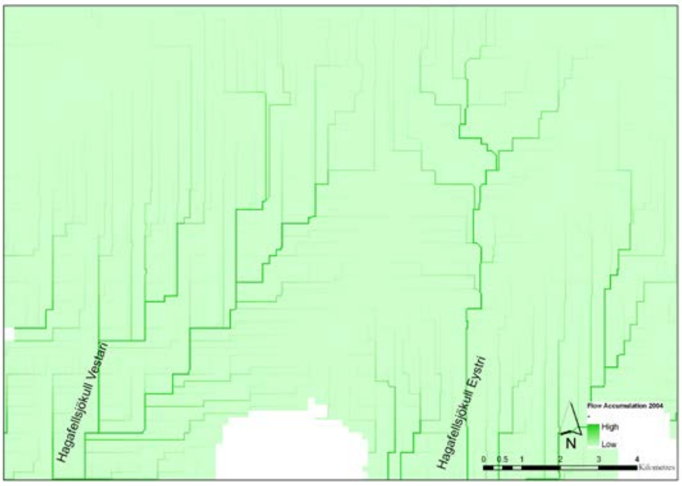


Figure 6.15: The evolution of the upper Hagafellsjökull outlets. Clockwise from above: 1986, 1997 and 2004. The comparison shows a significant switch between 1986 and 1997. This switch is mainly reversed by 2004.

This modelled return switch of this meltwater increases the flow of melt water beneath Hagafellsjökull Eystri and Hagafellsjökull Vestari. The largest outlet from Eystri is modelled as having ~0.75 $m^3 s^{-1}$ of output in 2007 which is a comparable flow rate to 1986 values. Values from the 2004 model were considerably higher than this but are not considered accurate because the modelled system provides only a very basic flow system in which most of the tributaries coalesce. This seems to be a consequence of the lower resolution of the 2004 imagery combined with the relatively low resolution of the subglacial DEM. Hence it was decided to use the considerably higher resolution 2007 model for comparison of outflow. Vestari also shows increases in flow - although the changes are more subtle and difficult to compare as there are some changes in the flow patterns (discussed overleaf).

The modelled subglacial hydrology of Hagafellsjökull Vestari following the surge of its neighbouring outlet shows some slight but influencial differences,

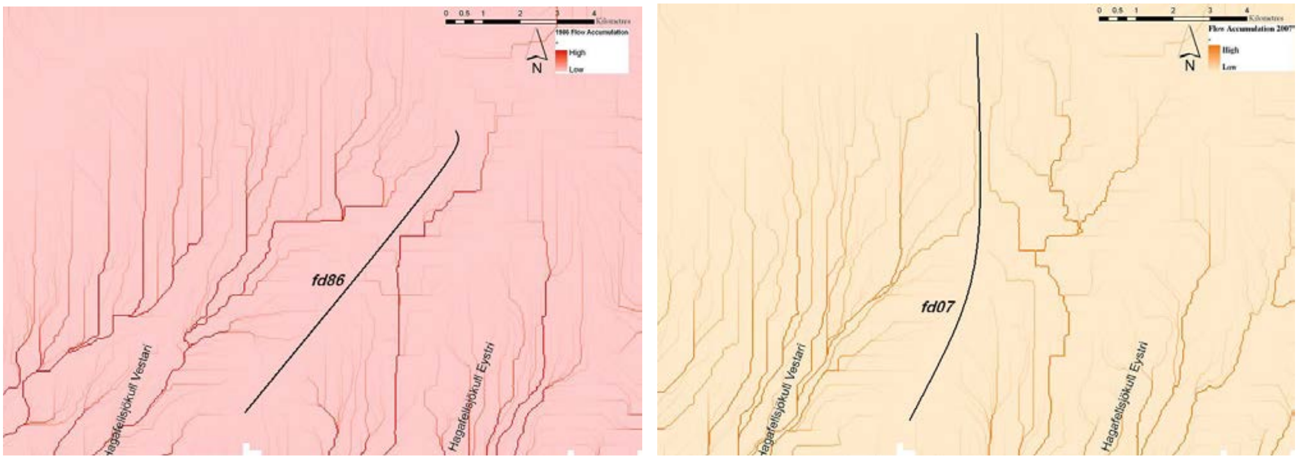


Figure 6.16: Comparison of pre (1986) and post surge (2007) hydrological systems of the upper Hagafellsjökull. Black lines indicate flow divides

particularly in the upper reaches. Figure 6.16 shows a certain amount of flow divide switch – the lines ‘fd86’ and ‘fd07’ demonstrate the extent of this change. The branch of the system modelled to switch has an output in the Eystri system of ~0.41 $m^{-3} s^{-1}$. This implies a gain to the hydrological system of Eystri and a loss to that of Vestari.

There are also some differences in the patterns of flow modelled for the Vestari hydrological system. Figure 6.17 shows how the pattern of the hydrological system has changed from 1986 to 2007. In the 1986 model there are essentially two main branch systems draining the upper accumulation area of Hagafellsjökull Vestari.

Figure 6.17: Pre (1986) and post surge (2007) comparison of the upper Hagafellsjökull Vestari system. Black arrows indicate shift in flow patterns

These branches form two main channels which flow approximately parallel to each other in a southerly direction. In the 2007 model much of the melt water from the most westerly of these branches is routed further south beneath the outlet before being routed out towards the south west. This branch is calculated to collect ~2m^3s^{-1} of melt water that is now flowing in a different direction - causing much more flow through the central section of the Vestari outlet. Despite this change in direction the model does not show the branches coalescing and the systems remain in separate channels that flow in closer proximity.

6.5.2 - Post 1997 changes to other outlets

The other outlets of the Langjokull ice cap again exhibit little change with similar flow levels and patterns. One exception to this is Þrístapajökull, which because described above, is modelled to experience reduced levels of flow as less meltwater is collected from the Hagafellsjökull accumulation area. Another exception to this is Suðurjökull which is shown in the 2007 model to have lost part of the flow increase in the 1997 model.

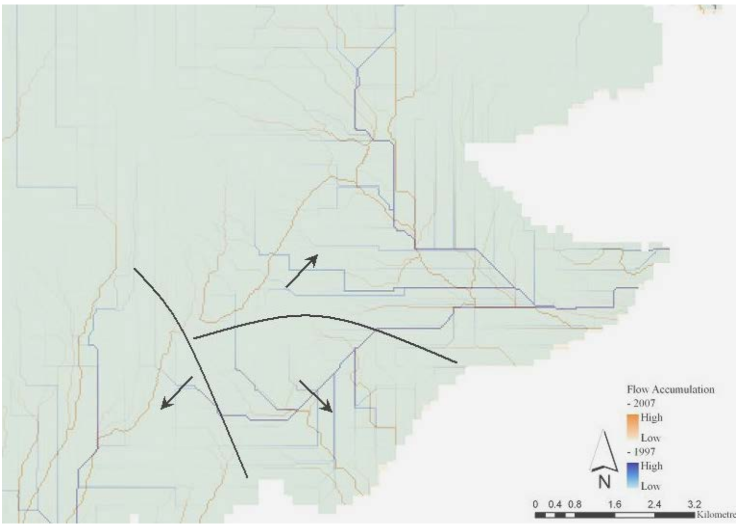


Figure 6.18: Composite of the 1997 and 2007 hydrological systems beneath Sudurjökull. Black lines and arrows approximate flow divides

Figure 6.18 shows a composite demonstrating this change. The black lines approximate the flow divided as they are modelled. Notably, they fall very close to the glacier flow margins shown in figure 6.14. This partial loss is calculated as a decrease in meltwater flow from *c.* 2.3 $m^3 s^{-1}$ to *c.* 1.4 $m^3 s^{-1}$ from 1997 – 2007. Most of this flow is lost to the unidentified area to the south. The remainder is lost to Hagafellsjökull Eystri.

7 - Discussion of results

The figures and description presented in the previous chapter show some considerable changes in the modelled drainage system of the Langjökull ice cap with progression from 1986 to 2007. This section will discuss why these changes occurred and link, wherever possible, to mechanisms and processes from the existing literature. Encompassed within this section will be deliberation of changes since the surge of Hagafellsjökull Eystri. This will include how the most recent modelled changes may suggest Langjökull is evolving into the future; particularly with reference to future surge behaviour. It also considers the methods used to acquire these results in terms of their confidence and potential applicability to other studies.

It should be considered at this point that while every effort was taken to model the melt across Langjökull as accurately as possible any figures for meltwater flux serve as a guide only. The melt model was calculated from data for 2006 – clearly annual variation in meltwater flux due to varying winter accumulation and summer melt means this cannot be representative for all of the time period considered here. Rather than supplying absolute figures for summer melt the meltwater flux model is more useful for analysing changes in patterns of flow due to topographic change under assumed reasonable meltwater flux estimates.

7.1 - Hagafellsjökull Eystri and Vestari: 1986 to 1997

The largest modelled change in the system was the considerable change in flow direction in the upper reaches of the Hagafellsjökull outlets. Over the entire modelled summer melt period of 155 days this is calculated from the model to have resulted in the loss of approximately 18.7×10^{-6} m^{-3} of meltwater from the Hagafellsjökull outlets. This flow was redirected to the north-east where it flowed out beneath Þrístapajökull. In terms of subglacial topography this switch seems entire feasible. The DEM of the subglacial area shows this area to be a relatively flat SW-NE orientated valley with a slight surface slope in a similar direction before turning to the NW beneath Þrístapajökull.

Figure 7.1 compares the modelled drainage systems in 1986 and 1997. Both systems are overlain by the calculated surface height change during the 1986-1997 period – characterised by accumulation in the upper reaches and ablation in the lower reaches. The change in the overburden pressure along with ice thickness change is also shown in this figure. The comparison shows that in 1986 a considerable amount of meltwater input to the Hagafellsjökull drainage systems is from the area accumulated the greatest amount of ice by 1997. By 1997 the drainage system has altered and the flow divide (approximated by the black lines) appears to have shifted. The position of this divide across the area of the greatest surface elevation increase

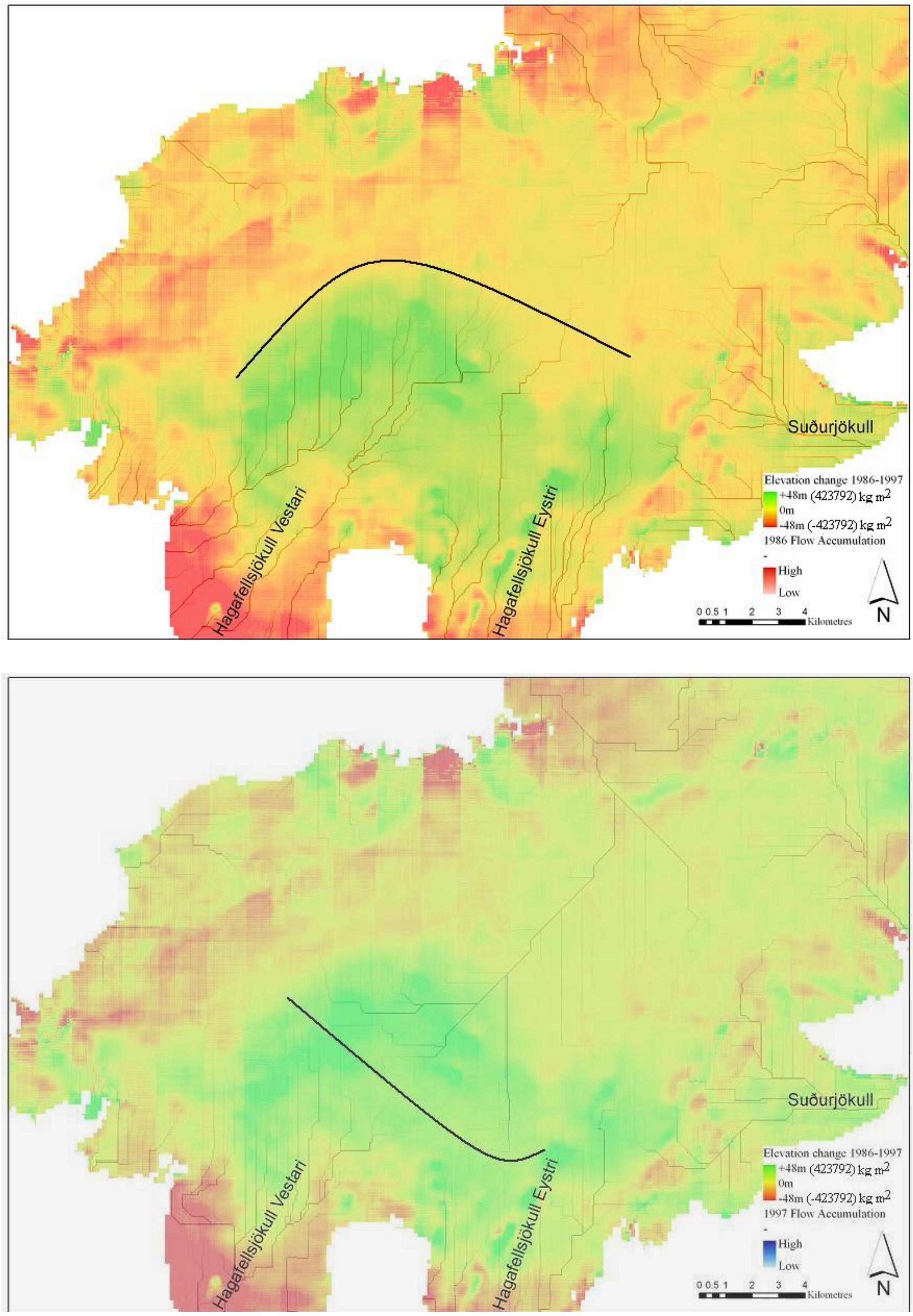


Figure 7.1: The change in flow divide from 1986 (top) to 1997 (bottom) overlain by surface elevation change. Black lines indicate approximate flow divides

suggests increase in surface elevation of 20-30 m had a substantial effect upon the subglacial drainage system. Comparatively, the area to the north-east is characterised by little or no change in ice mass from 1986 to 1997. It therefore seems that the relative increase in overburden pressure in the upper area of the Hagafellsjökull outlets had the effect of causing some of the flow to switch to the north. This concept

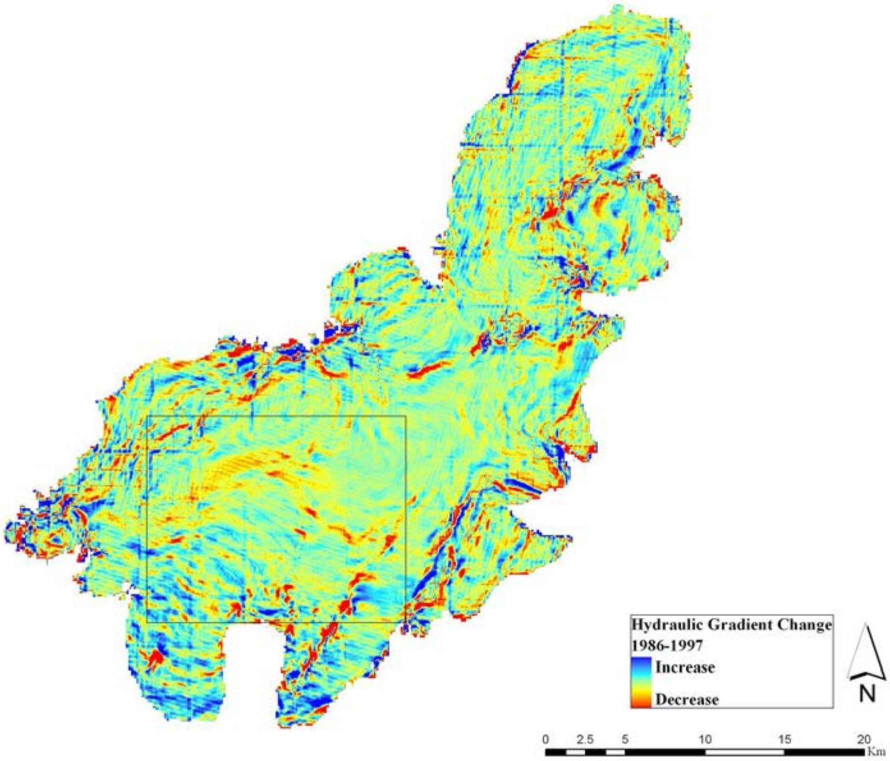


Figure 7.2: Overview of the change in the calculated hydraulic gradient from 1986 to 1997. The box indicate the area in figure 7.3(overleaf).

is similar to the flow switching hypothesised to have occurred between the Rutford Ice Stream and the Carlson Inlet, Antarctica (Vaughan *et al.,* 2008). Here, the suggested reason for this switch in flow is alteration of the hydraulic gradient at the glacier bed. Figure 7.2 shows the calculated change in the hydraulic gradient from 1986 to 1997. The hydraulic gradient was calculable using equation 3, as utilised by Arnold & Sharp (2002):

Equation 3: $\boldsymbol{\varphi = \alpha + [(\rho_w - \rho_i)/\rho_w]\beta}$.

Here: $\boldsymbol{\varphi}$ is hydraulic gradient, $\boldsymbol{\alpha}$ the surface slope, $\boldsymbol{\rho_w}$ the density of water (1000 kg m^3), $\boldsymbol{\rho_i}$ the density of ice (900 kg m^3) and $\boldsymbol{\beta}$ the bed slope. The increased overburden pressure of ice associated with the slope increase in the upper reaches of the

Hagafellsjökull outlets from 1986 to 1997 is modelled to have created changes that certainly had the potential to alter the flow of subglacial water. The areas of the greatest overburden pressure increase show, as expected, the greatest loss in hydraulic gradient. A loss of hydraulic gradient in a particular area results in meltwater flowing

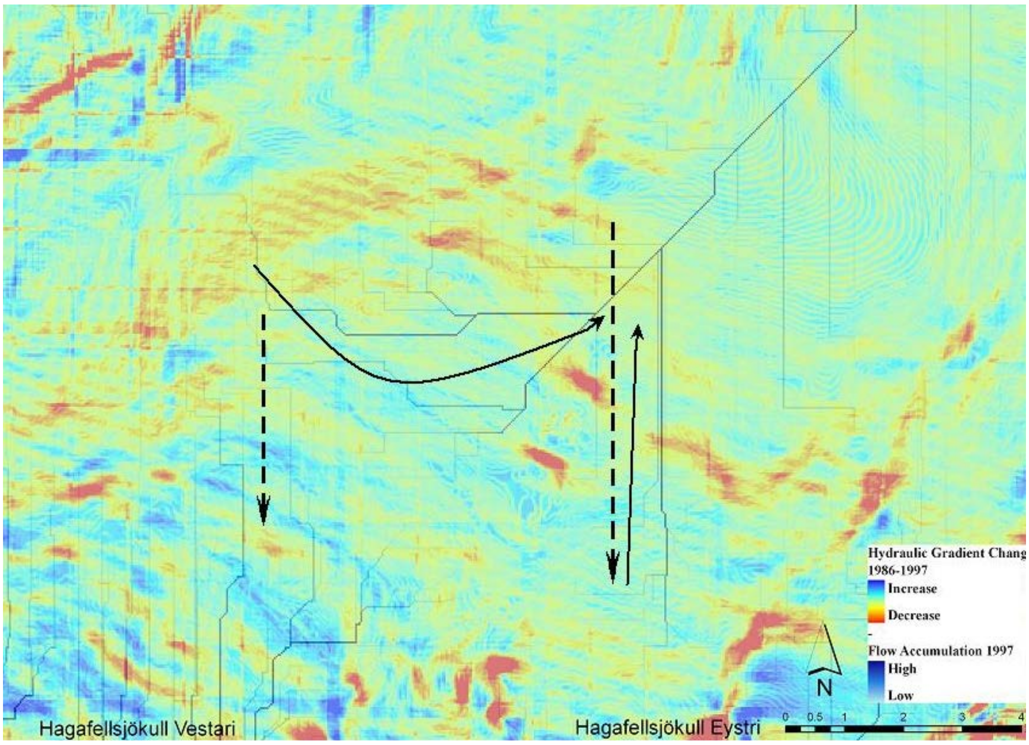


Figure 7.3: 1986-1997 change in hydraulic gradient in upper Hagafellsjökull. The solid arrows indicate possible flow paths to the north-east in 1997. The dashed arrows indicate the general 1986 flow direction.

away from this area towards an area of lower overburden pressure – meltwater will always take the route of least resistance. Towards the north-east where overburden pressures showed little change from 1986 to 1997 the trend of hydraulic gradient matches accordingly. This lack of change relative to the loss in the hydraulic gradient in the upper Hagafellsjökull glaciers would certainly provide a distinct hydraulic gradient towards the north-east. An area of lower hydraulic potential in the upper reaches of Hagafellsjökull Eystri primarily appears to form a barrier to flow to the north-east. However, the solid black arrows indicate inconsistent gaps in this ‘barrier’ that have higher hydraulic gradient and could potentially draw water from areas of lower gradient. These gaps could well be the key to allowing meltwater to flow to the north-east. These arrows also bear a good spatial relation to the modelled main flow channels for 1997. Further to this it is clear in figure 7.1 that the area of surface height increase of Eystri extends further south than that of Vestari. This would result in higher overburden pressure restricting flow towards Eystri. It is therefore suggested that this increased hydraulic gradient running south-west to north-east is the mechanism resulting in the loss of some of the flow from the Hagafellsjökull glaciers in 1997.

If this mechanism is correct the obvious question is why did Hagafellsjökull Eystri surge in 1998 whereas Hagafellsjökull Vestari failed to do so? Clearly both outlets were subject to accumulation in their upper reaches and yet only Hagafellsjökull Eystri surged successfully. However, noting the observations of increased surface velocities on Hagafellsjökull Vestari in 1998 (Pálsson, 2010. Personal communication) it seems apparent that a surge of Vestari was near but failed to initialise fully. The increased velocity noted in the upper areas may well have been the enhanced velocity zone described by Björnsson *et al.,* (2003). The formation of a bulge at the lower area of this zone precedes a surge. The rapid movement of this bulge is described as "first unquestionable sign of a surge" (Björnsson *et al.,* 2003 : pg. 87). Hagafellsjökull Vestari seem to have reached the stage of the increased velocity zone (with the formation of a bulge probable) but then failed to progress to the next stage – *i.e.* surging. The noted flood of water from the subglacial system implies a linked cavity system may have formed temporarily, but it was not stable or did not dominate a sufficient area of the bed to cause a surge. Hence, the water was released as an efficient channel drainage system was reformed. Linking this back to Kamb's stability criteria (Kamb *et al.,* 1985 – equation 1) this would imply a ψ value less than 0.8 was not reached across a sufficiently wide area of the bed to sustain a surge. Attempts to produce a model displaying values of ψ were made here but conclusions could not be reached – partly due to lack of knowledge of basal velocity fields and orifice step heights specific to Langjökull. The cause of the failure to sustain the surge of Vestari could lie with several observed factors. Figure 6.13 shows the tributary branches feeding Hagafellsjökull Eystri are somewhat shorter and accumulate less than those of Hagafellsjökull Vestari. Eystri does, however, have one branch noticeably longer than most which is modelled to be lost to the Þrístapajökull hydrological system in 1997 (labelled L in figure 6.12). Surges usually begin in winter (Harrison & Post, 2003) due to lower water fluxes as melt rates decrease. The 1998 surge of Eystri was no exception with the surge beginning in the last months of the year. The loss of the large branch '*L*' may have been pivotal in the initiation of the surge. The decrease of accumulated flow through this main drainage channel may have resulted in the collapse of the system and a switch to a linked cavity system. This change, coupled with less flow through the smaller flow tributaries during winter, could well have sparked a change to a linked cavity system beneath a sufficient proportion of the glacier to initiate a surge. The overburden pressure increase between 1986 and 1997 (figure 7.1) shows the increase spreads considerably further south on Eystri than Vestari and this could also have been important causing drainage channel closure.

Hagafellsjökull Vestari was similarly affected by the modelled switch in flow towards Þrístapajökull but a surge was not sustained. Figure 6.13 shows a considerable pattern change in the hydrological system beneath Vestari with fewer, larger channels. Unfortunately this may again be the result of the lower resolution of the 1997 surface DEM combined with the lower resolution subglacial DEM. Intriguingly, however, the modelled 1997 system beneath Eystri shows no such change – indeed Eystri appears to have a greater number of smaller channels despite the same data resolutions being applied. Consideration is therefore given to the idea that a surge of Vestari was averted due to a change in the hydrological flow pattern. A single channel with larger flow would potentially be more stable against collapse through overburden pressure than numerous small channels because increased flow levels would be more likely to sustain channels via melting. The possibility of a larger channel forming from confluence of other channels, particularly beneath the rapidly thinning ablation area of Vestari, could be possible considering the ever reducing overburden pressure due to surface lowering. The change could also be driven by the pattern of accumulation above Hagafellsjökull Vestari (see section 7.4.1). The change in the system to fewer, larger channels is modelled to be carried over to the 2004 reconstruction (figure 6.15) although the lower resolution makes its reliability questionable. The 2007 DEM shows a switch to smaller channels but the pattern evolves somewhat differently (also discussed in section 7.4.1).

7.2: Suðurjökull *changes: 1986-1997*

The modelled changes in Suðurjökull from 1986 are somewhat difficult to explain. The incorporation into the Suðurjökull system of flow from the unnamed area between Suðurjökull and Hagafellsjökull Eystri (and some very limited flow capture from Hagafelljokull Eystri) seems to have had little effect upon the system. One slight change in the topography constrained area between Eystri and Suðurjökull is some marginal thickening against this constraining feature at the south-east edge. This appears to have led to a drop in hydraulic gradient at this bounded area, visible around 'Th' in figure 7.4. Conversely the area 'D' shows slightly increased hydraulic gradient. Although this area is not classed as an outlet it could be possible that flow

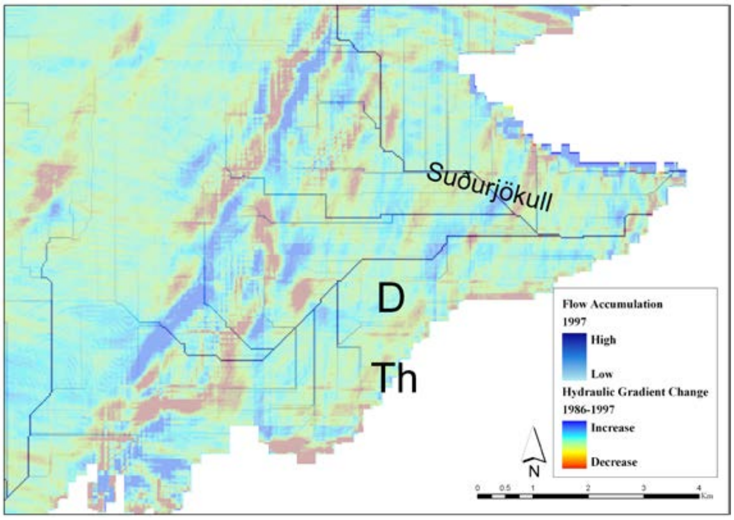


Figure7.4: Changes in hydraulic gradient of Sudurjökull overlain by 1997 flow system

velocities increased slightly due to increased pressure caused by moderate surface elevation increases in the upper reaches. The slight change of hydraulic gradient from the outside to the centre may explain the slight flow migration north towards Suðurjökull – the overlain 1997 modelled hydrological system follows this pattern as figure 7.4 shows. The effects these modelled changes had upon Suðurjökull and the neighbouring area seem to have been nil or small. Again, these it is difficult to assess such relatively small changes as they may be the result of DEM resolution variability.

7.3 - Climatic impacts upon accumulation

The accumulation of mass in the upper reaches of the Hagafellsjökull appears to have had a considerable impact upon the flow direction and the modelled system implies a link to the surge of Hagafellsjökull Eystri. The pattern of surface elevation change from 1986 to 1997 is interesting. The highest levels of accumulation are not necessarily the highest elevations. The accumulation data from 2006 shows that the highest levels of precipitation (*c.* 3.5m a^{-1}) are reached on the SW facing slopes at around 1300m: the Hagafellsjökull accumulation area. This corresponds to L07 in figure 4.7. The highest point of the ice cap actually lies to the north (L08 – 1413 m a.s.l.) but actually receives *c.* 1.4m a^{-1} less precipitation. This is visible in the surface elevation change (figure 7.1) where the greatest surface increases do not correspond to the highest points of the glacier. This demonstrates how Langjökull is influenced by the prevailing winds which bring the greatest levels of precipitation to the south. This may have had important influence upon the flow switching from the Hagafellsjökull outlets to a north-easterly direction. The accumulation on the southern slopes coupled with the decrease in precipitation with northerly progression appears to have changed the hydraulic gradient sufficiently to switch the meltwater flow direction. Had the accumulation of mass on Langjökull been equal this gradient may have formed. The influence of precipitation seems to have been important in redefining part of the hydraulic gradient from NE-SW to SW-NE.

7.4: Hagafellsjökull Eystri and Vestari: 1997-2007

Following the surge of Hagafellsjökull Eystri in 1998 the increased surface velocities led to significant transfer of ice mass to the ablation zone with the effect of rapidly increasing the surface velocity. Figure 7.5, below, shows this change clearly with a pronounced switch from loss to gain from pre 1997 (top) to post 1997 (bottom). Hagafellsjökull Vestari, which apparently did not surge, shows no such change and retains generally a similar pattern. There are however, some subtle

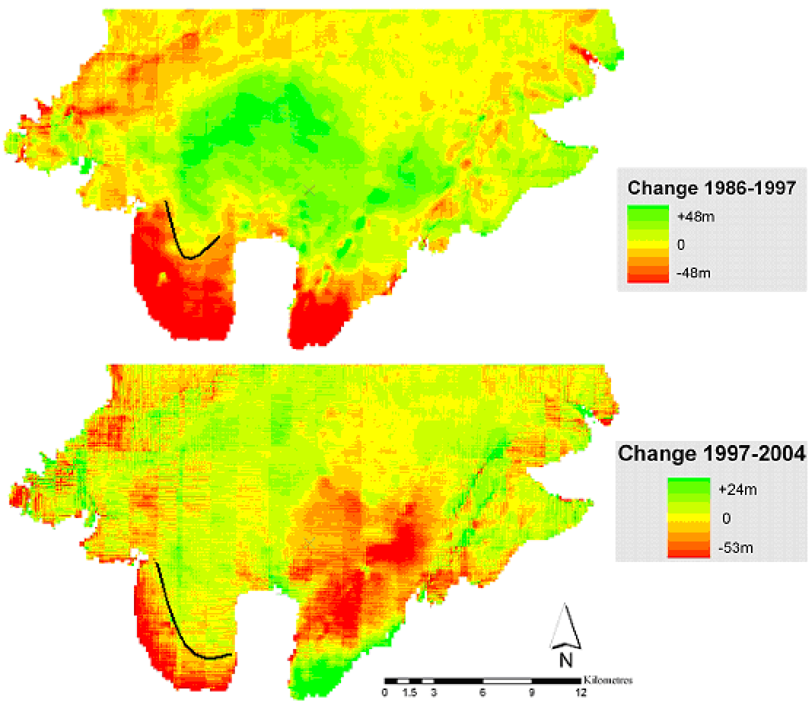


Figure 7.5: Changes in the apparent equilibrium line of Hagafellsjökull Vestari from 1986-1997(top) and 1997-2004 (bottom). The lowering line suggest some mass transfer on Vestari – possibly during the failed 1998 surge.

changes in the topography of Vestari. The black lines indicate essentially the equilibrium line – the point between the accumulation and ablation zones. There is a decided shift to the south in this line when compared pre and post surge. The cause of this is likely to be the increased surface velocities noted by Pálsson (2010, personal communication) that appear to have lead to led to some transfer of mass. This appears to have been a failed surge as explained in the previous section – assuming a surge is defined by the advance of the glacier terminus. If a surge were to be defined as a rapid, considerable transfer of mass down glacier (but not necessarily a terminus advance) then arguably Hagafellsjökull Vestari *did* surge in 1998. The transfer of mass during the early part of the surge (in the *enhanced velocity zone* of Björnsson *et al.*, 2003) could well have restored the balance of the glacier sufficiently to allow an efficient channelised drainage system to reform in the upper area. The slope angle of Hagafellsjökull Eystri shows a clear post surge reduction – the lighter colouring of the lower reaches in figure 7.6 demonstrate this well. Comparatively, Hagafellsjökull Vestari seems to show little change with a similar slope across the whole outlet for both years. This implies the 1998 increased flow of Hagafellsjökull Vestari did little to rebalance the outlet - slope is a key characteristic that alters as a surging glacier becomes out of balance with accumulation only to be restored it during surging. Comparison with a DEM closer to 1997 may allow a change in surface slope to be assessed better – the 2004 DEM unfortunately leaves a considerable temporal gap from the 1998 surge and the data resolution is less.

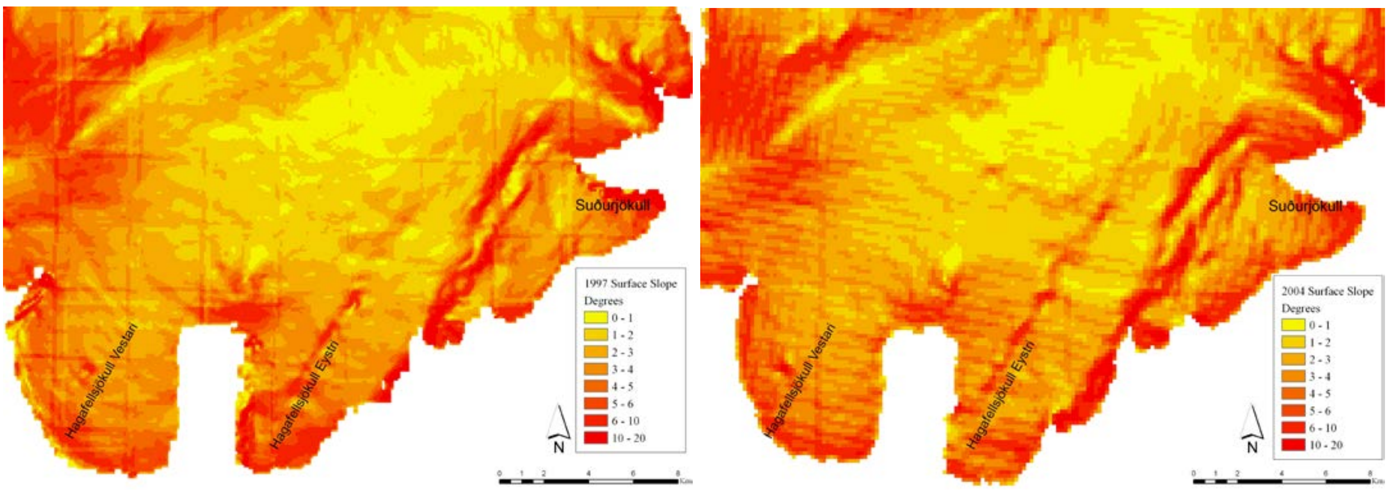


Figure 7.6: Slope angle comparison of the southern outlets, 1997 (left) and 2004 (right).

As described in the results chapter the key modelled change from 1997 – 2004 was the switch of the flow lost to Þrístapajökull back to the Hagafellsjökull outlets. The previous section explains this change was caused initially by alterations in the subglacial hydrological gradient through increased overburden pressure of accumulating ice. The switch back of a modelled 1.2 $m^3\ s^{-1}$ back the Hagafellsjökull outlets would therefore instinctively be due to a lowering of the overburden pressure and restoration of a similar hydraulic gradient to that of 1986. Figure 7.7 certainly shows an increase in gradient towards Hagafellsjökull Eystri (the blue areas in the upper reaches). The increases in the surface elevation in the lower reaches of Hagafellsjökull Eystri are also well pronounced with a loss of potential clearly visible (red/yellow areas). Hagafellsjökull Vestari shows little change in hydraulic gradient - although the continued ablation of the lower reaches has led some increase in this area. The upper reaches are marked by little change – in-keeping with the lack of a full surge to restore the balance of the outlet. Post 1998 the overburden pressure in upper Vestari is still higher than in 1986 – also suggesting the glacier remains out of balance, although presumably insufficiently to cause a surge. Due to this lower hydraulic gradient caused it therefore seems unlikely that the flow modelled to be lost in 1997 would be capable of re-establishing drainage beneath Hagafellsjökull Vestari. Indeed this does not seem to have occurred. The conditions surrounding this are discussed in the next section. A post surge decrease in overburden pressure and increased potential for flow is therefore suggested as the mechanism for the switch of flow back to the Hagafellsjökull glaciers.

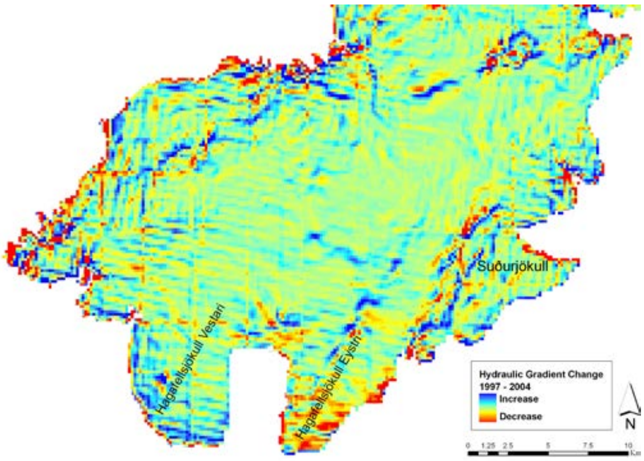


Figure 7.7: Hydraulic gradient changes of the southern outlets from 1997-2004.

7.4.1 – Effects of failed surge of Hagafellsjökull Vestari

The modelled flow switching due to mass build-up prior to the 1998 surge is not modelled to have affected the flow distribution between the two Hagafellsjökull outlets: each outlet is modelled to have lost a similar share of meltwater to Þrístapajökull. However, following the 1998 surge and the modelled return of flow to the Hagafellsjökull outlets there are some differences in the flow patterns and volumes as described in the previous chapter (figure 6.16). Figure 7.8 shows the

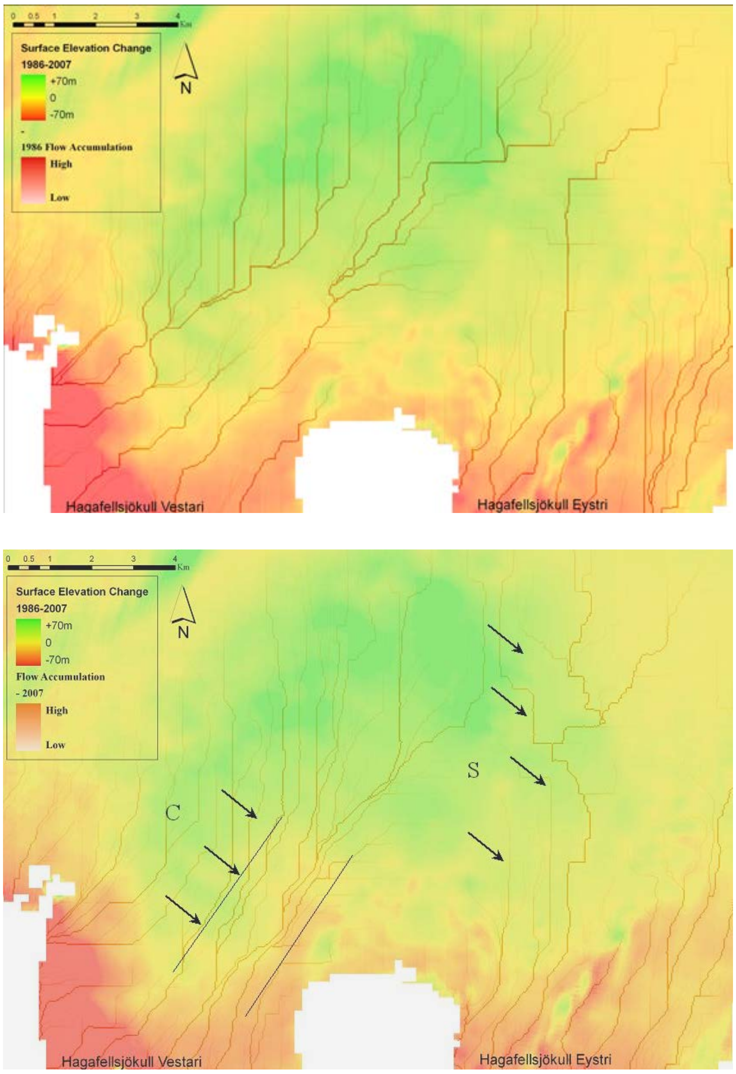


Figure 7.8: 1986 (top) and 2007 (bottom) modelled hydrological systems overlain by the change in surface elevation for the period

modelled hydrological systems for 1986 and 2007 overlain by the total change in surface elevation during this period. The effects of the surge disparity are clear. Hagafellsjökull Eystri shows seemingly little change - it has rebalanced by the surge in 1998. On the contrary, Hagafellsjökull Vestari shows considerable surface elevation gains in the accumulation area because it has not surged. Both outlets have net loss of mass in their ablation areas (showing that the increase in elevation caused by rapid discharge during the 1998 surge was quickly melted). The effects of this elevation difference, and thus overburden pressure and hydraulic gradient, are clear to see in comparisons of the modelled 1986 and 2007 subglacial drainage systems. The arrows labelled ‘S’ in figure 7.8 indicate a flow shift from Hagafellsjökull Vestari to Hagafellsjökull Eystri. Comparison with the overlain surface elevation change shows a good relation between this shift and areas of surface elevation increase. This appears to be due to the failure of Vestari to surge fully in 1998 which caused it to remain out of balance. The increased pressure of ice is causing displacement of subglacial water towards Hagafellsjökull Eystri. Following the 1998 surge and a decrease in overburden pressure of Hagafellsjökull Eystri an increased hydraulic gradient will exist. It therefore appears to the author that Hagafellsjökull Vestari is not only out of balance within its own glacial system but is also out of balance with its neighbouring glacier and this imbalance is causing changing interaction between the two. Further to this, the increased mass of ice is causing changes in the flow pattern beneath Hagafellsjökull Vestari itself. The arrows labelled ‘C’ in figure 7.8 suggest a shift in flow towards the east (also see figure 6.17). Again, consideration of the elevation changes from 1986-2007 show a relation to these changes. The greatest elevation increase areas (green) see less accumulation of flow as flow is being forced east by increased overburden pressure. Similarly, this likely to be being increased by the surface lowering (red) in the east close to the Hagafel ridge leading to an increased hydraulic gradient.

7.5: Post 1997 changes to other outlets

The outlet Suðurjökull is modelled in figure 6.18 to lose the flux of meltwater gained from the area between itself and Hagafellsjökull Eystri. The reasons for this are not widely apparent with little change in ice thickness exhibited after 1997– again this suggests the accuracy of modelling such small changes may be limited by the resolution of the DEM. One possible reason that can be suggested is that the increased hydraulic gradient towards the subglacial system of Hagafellsjökull Eystri following the surge in 1997 may have removed potential meltwater sources. The only other notable change to an outlet glacier is Þrístapajökull which, as described, returned to a flow volume similar to 1986 when the flow gained from the

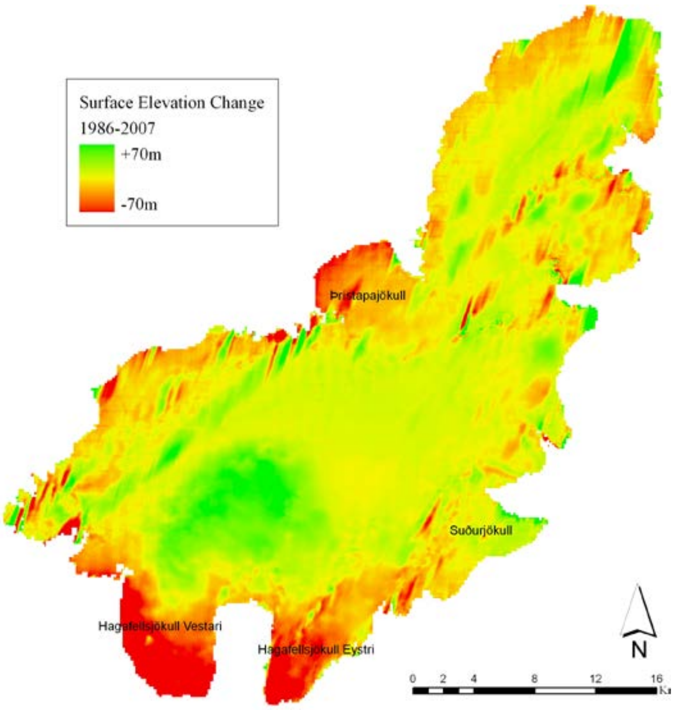
Hagafellsjökull hydrological system switched back following the 1997 surge of Hagafeljokull Eystri.

The changes in the hydrological systems of Suðurjökull and Þrístapajökull are small, difficult to deduce clear mechanisms for and seem to have little effect upon the ice flow of the outlets. There is, however, one particular reason they are worthy of note. With the exception of Hagafellsjökull Eystri and Vestari these outlets are the only outlets modelled to have subglacial hydrology affected by the changing surface topography. These changes are both it direct (as Suðurjökull which shows some surface elevation changes) or indirectly via changing meltwater flow divides as Þrístapajökull experienced. Intriguingly, both of these outlets are identified by Björnsson *et al.,* (2003) as having potentially surged in the past (figure 1.3). Evidence for this is described as anecdotal only and unfortunately no further evidence is given. Palmer *et al.,* (2009) also imply Suðurjökull may be a surge type outlet after recording increased surface velocities in 2004 (although this study finds no change in surface elevation to suggest a transfer of mass down glacier). Þrístapajökull is described by Flowers *et al.,* (2007) as having a 'tendency to advance from its current position over the gently sloping plain' during modelling experiments of the Little Ice Age. The occurrence of the modelled changes in hydrological systems found here and the suggestion of these glaciers once surged seems an unlikely coincidence. The author therefore suggests that under past conditions of surface melt and icecap topography changes in the hydrological system may have been responsible for surges of these other outlets of the Langjökull ice cap. A suggested method is a periodic switching of flow from the southern Hagafellsjökull and Suðurjökull systems to the northerly Þrístapajökull. This could potentially have resulted in a periodic switch in surging of southern and northern facing outlets.

7.6 : The future of the Langjökull Ice Cap

The changes modelled here show a retreating ice cap characterised by a general decrease in volume through continued ablation of outlet glaciers. Comparison of the earliest (1986) and latest (2007) DEMs (figure 7.9) show the most prominently retreating areas are Hagafellsjökull Vestari, Eystri and Þrístapajökull. This study sees no reason to disagree with estimates of a *c.* 7% loss from 1997-2006 and that Langjökull could disappear entirely within 150 years (Björnsson & Palsson, 2008).

7.6.1 - Hagafellsjökull Vestari: Ice volume gains across the entire ice cap are restricted to modest increases in the interior, possibly due to increased precipitation, and substantial gains in the accumulation area of Hagafelljokull Vestari (again, as shown in figure 7.9). This increase is the result of the failure of Hagafellsjökull Vestari to surge fully along with its neighbouring outlet in 1998. The continued ablation in of lower Vestari coupled with continuing gains of upper Vestari are resulting in the system becoming increasingly imbalanced.



Hydrological changes resulting from this, as described in the previous section, are modelled to be forcing meltwater to switch flow away from Vestari into the system of Eystri. This imbalance and reduced meltwater flux suggest a surge of Hagafellsjökull Vestari is increasingly likely and inevitable. Estimation of a precise date is difficult but the author would suggest, due the presently increasing imbalance and meltwater flow switching to neighbouring Eystri it will certainly occur within the next 5 years.

7.6.2: Hagafellsjökull Eystri: Recent observation from 2004 - 2007 (figure 6.4) shows the surge cycle of this outlet appears to have restarted placing the outlet in the quiescent phase. Ablation in the lower reaches and modest accumulation in the upper reaches over this brief period of monitoring suggest the outlet is currently building up to another surge. This could be delayed by increasing meltwater flow levels feeding into the system from neighbouring Vestari caused by ice cap topography changes described. If a surge of Vestari occurs first, which seems likely, this change could be reversed making a surge of Eystri increasingly likely after this juncture. This demonstrates the dynamic relationship these outlets are currently modelled to sustain.

7.6.3 – Climate effects into the future

Surges of both glaciers could be influenced by modelled climatic changes. Modelled predictions of a 2.8°C temperature rise and 6% precipitation increase towards the end of the 21st century (Bjornsson & Palsson, 2008) could result in considerable changes. The rising temperatures will surely lead to an increase in ablation of the outlets cited above and also the other lower altitude areas of the ice cap, resulting in higher levels of meltwater production. Meltwater discharge is estimated to peak within 50 years and then decrease to present levels within 100 years. Increased precipitation levels have the potential to cause greater accumulation in the higher altitude areas (assuming of course these changes are not counteracted by increased melt). These changes in melt and precipitation considered together give the possibility of an increased incidence of surging in the Hagafellsjökull outlets due to greater imbalance of the system, although as yet there the frequency of surging does not show any such pattern. Varying meltwater flux is expected to become an increasingly important factor leading to surge initiation. Greater variation in levels from summer to winter could causes greater instability in subglacial drainage systems. For example: a particularly warm, meltwater abundant summer followed by a cold, melt water deprived winter could be increasingly likely to trigger a surge.

7.7 - Review of techniques employed.

The analysis technique used here provides excellent spatial cover of the Langjökull ice cap and could be readily applied to any mountain glacier, small ice cap or a section of a larger ice cap. It allows the changes within the entire system to be considered, particularly interactions between individual outlets. One of the main limiting factors to another study could be data availability; in particular the GPR derived subglacial topography which is not readily available for most glaciers. In terms of both this study and others the temporal resolution of DEMs could be problematic. This study could certainly have benefited from more frequent DEMs of the ice cap in order to consider the rapid changes that occurred in and around the surge of 1998. As considered throughout the study the spatial resolution of the DEMs themselves can be problematic. Here the most accurate DEM is the 2007 LiDAR data and the least accurate the 2004 data. This has a great effect on how much can be gleaned from modelled changes in both surface topography and subglacial hydrology as it can be difficult to assess how much changes are due to actual changes or merely the difference in DEM resolution.

Similarly the melt model employed here has the potential to be adapted for use in other areas. The model used here was adapted from a previous study in Norway – this was possible through existing knowledge of input parameters from the existing literature and calculation of unknown specific parameters to Langjokull (as in section 5.3). In order to adapt this model to different locations knowledge of specific local parameters would be necessary. Clearly the dependability of the modelled output will be affected by the accuracy of this data. A limiting factor in this study is the change in precipitation with northerly progression which characterises Iceland and the rain-shadow effect of the highest points in the south (Flowers *et al.*, 2007). While the precipitation gradient with altitude was taken into account the melt model contains no consideration of this other spatial variation in precipitation.

8 – Conclusion

Digital elevation models derived from a number of sources for the period 1986 to 2007 have been used to assess changes in the surging Hagafellsjökull outlet glaciers of the Langjökull ice cap, Iceland as well as the neighbouring outlet glaciers. A model of surface melt was used to provide an input of water flux for reconstructions of the subglacial hydrology of the years 1986, 1997, 2004 and 2007. Topographic changes in the ice cap have been considered and related to noted changes in the modelled subglacial hydrology. This interaction is considered with regard to the causal mechanisms of glacier surging in order to better understand the behaviour of the surging outlets of Langjökull.

This study concludes that accumulation of mass in the upper reaches of the Hagafellsjökull outlets between the previous recorded surge in 1980 and 1997 led to an imbalance glacier system, as characterises surging glaciers. Coupled with continued ablation in the lower reaches of the outlets this led to changes and in overburden pressure of ice and, accordingly, subglacial hydraulic gradients. These alterations eventually resulted in temporary migration of a percentage of subglacial away from the Hagafellsjökull outlets towards the northern Þrístapajökull outlet. The reduced flow levels are here linked to the concept of the collapse of the subglacial channelised drainage system and the formation of a linked cavity system. The formation of this system led to rapidly increased surface velocities of Hagafellsjökull Eystri, beginning in 1998, which resulted in a terminus advance and considerable gains in surface elevation in the lower reaches through transfer of glacier mass from the accumulation area, thus rebalancing the glacier system. Noted surface velocity increases on neighbouring Hagafellsjökull Vestari, also in 1998, are suggested to be the beginning of a similar surge which was not sustained due to the reestablishment of efficient subglacial drainage. Possible reasons for this surge failure from the modelled system are suggested; the change to a single, larger drainage system and a sufficient rebalancing of the glacier system.

Following the 1998 surge the associated change in topography reduced overburden pressure in the upper Hagafelljokull Eystri. This resulted in a change in hydraulic gradient back to a similar pattern to 1986. Partial restoration of flow previously lost to the Þrístapajökull system is modelled. Most recent observations show Hagafellsjökull Eystri to have returned to its quiescent phase with moderate surface elevation gains in the upper reaches once more. The failure of Hagsfelljokull Vestari to surge has led to a net gain in elevation over the entire study period and as a consequence patterns of flow between the Hagafellsjökull outlets have changed, with more flow being driven towards Hagafellsjökull Eystri.

It is suggested that modelled changes in the subglacial hydrology of neighbouring Suðurjökull and Þrístapajökull are linked to the surge behaviour of the Hagafellsjökull outlets. This is linked to evidence suggesting these outlets may once have been surge type: study supports these claims through observations of potential influences on their subglacial hydrology and suggests previous conditions could well have resulted in surging.

The future of Langjökull is discussed and agreement is reached with published work that Langjökull is an ice cap under going retreat with the potential to melt completely within approximately 150 years. A warming climate will increase meltwater production rates rapidly in the next 50 years. This, along with continued retreat elsewhere in Iceland, will have profound impacts upon groundwater and river runoff, which are both heavily dependant upon glaciers. The current imbalance of the Hagafellsjökull Vestari glacier system coupled with the modelled changes in subglacial hydrology suggest a surge of this outlet is increasingly likely within the next 5 years and it is suggested that increased precipitation and meltwater production rates under a changing climate have the potential to increase the likelihood of outlet surging into the next century.

The techniques employed here are found to be limited by spatial and temporal resolution of data to some extent but provide an excellent method of linking topographic changes to subglacial processes. Despite some potential limiting factors the technique is suggested to be highly applicable to studies in other areas and could be used to investigate similarly interacting glacier systems.

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