Glacier Fluctuations and Climatic Change in Iceland

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Declaration

I declare that this thesis has been composed by myself and is wholly my own work, except where other authors work is explicitly acknowledged within the text.

Abstract

This thesis aims to develop an understanding of the relationship between climate, topography and glacier fluctuations in Iceland. A mass balance/glacier flow model is applied to the Holocene fluctuations of Sólheimajökull, an outlet glacier in southern Iceland. The model is also used to predict the response of Sólheimajökull to future climatic warming. The findings provide insight into the spatial variability of glacier fluctuations in Iceland, and the dynamics of Holocene climatic changes in the North Atlantic.

The results from the model suggest that the response of Icelandic glaciers to climatic change can be related to glacier area-altitude distribution. Outlet valley glaciers located in high precipitation areas descend to elevations of 0-100 m where air temperature is mild. Ablation occurs throughout the year and glacier mass balance has a large amplitude response to temperature variations. Furthermore, outlet valley glaciers experience dynamic length variations in response to climatic change. This is a geometric effect where small changes in ice cap volume result in significant fluctuations in glacier length. In contrast, wide ice cap lobes in central Iceland exhibit a different response to climatic change. Precipitation levels are lower and glaciers terminate at altitudes of 600-800 m. Ablation is restricted to the summer months, and glacier mass-balance has a lower amplitude response to temperature variations. In addition, ice cap lobes experience smaller dynamic length fluctuations in response to climatic change. This is because ice cap margins undergo small changes in extent in response to changes in glacier volume. Finally, where ice cap lobes terminate on sandur plains, further advance leads to glacier widening and an unsustainable increase in ablation.

The numerical model is used to successfully reconstruct Holocene climatic changes over the last 5000 years from the record of glacier length variations at Sólheimajökull. Climatic changes in Iceland involve switches between two modes. During Holocene cold phases, temperatures are approximately 2°C colder than the 1966-1996 mean. Sea ice is extensive in the Greenland Sea, and the North Atlantic Oscillation is dominated by its negative phase. High pressure occurs over Iceland and the zonally averaged surface westerlies are displaced toward the equator. During Holocene warm phases, the climate of Iceland is warmer by 2°C relative to the 1966-1996 mean. The Greenland Sea is relatively ice free and the North Atlantic Oscillation is dominated by its positive phase. Atmospheric circulation in Iceland is influenced by a vigorous Icelandic Low. Switches between these two climatic states occur rapidly. The 18th, 19th and 20th Centuries were dominated by a decadal-scale climatic oscillation. This included a warming and cooling cycle in the 18th Century that was of similar magnitude to climatic changes during the 20th Century. As these changes occurred prior to the Industrial Revolution, they exemplify the frequency at which large natural climatic variations can occur. Natural variability in the Arctic must be considered in the prediction of future climatic changes. Projection of future glacier trends will be limited until this is achieved.

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Acknowledgements

My interest in glaciers was initiated as an undergraduate at the University of Melbourne by the glaciologists of the Australian Antarctic Division. My enthusiasm really developed during a field trip to Tasmania where I undertook a study of Quaternary moraines. Professor Eric Colhoun encouraged me through careful assistance in the field, and sound advice later at the University of Newcastle, Australia. After Newcastle I was interested in learning more about the interactions between glaciers and climate. I was thrilled to be granted an ORS Award and a Geography Department Studentship at the University of Edinburgh and I was in southern Iceland within a few weeks of hearing the news. I wish to thank Nick Spedding for introducing me to real glaciers, and their sediments and landforms.

I decided to focus on numerical modelling of Icelandic glaciers after a fruitless search for Holocene moraines. I thought that it might be possible to understand why some glaciers have long moraine sequences, while in most cases moraine sequences dating from before the Little Ice Age are lacking. I also thought this might help us to understand the climatic significance of glacier fluctuations in the areas where moraine records exist.

With its detailed moraine chronology, Sólheimajökull was the obvious choice for a modelling study. Before this could be achieved, a survey of the glacier needed to be undertaken. I wish to thank the Department of Geography for allowing me to purchase a new ice radar and differential GPS equipment, and Frank Jacobsen of the University of Aarhus, Denmark for instructing me in its use. I also wish to thank the National Geographic Society, the Carnegie Trust for the University of Scotland and the NABO program of the National Science Foundation of America (Directed by Prof. T. McGovern) for funding the fieldwork; the two field seasons undertaking soundings on the ice were fantastic and allowed collection of data that eventually enabled the thesis to be written.

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I dedicate the thesis to my parents, who encouraged my dreams and supported my choices.

April 1996 survey of Sólheimajökull. View down-glacier toward the south coast of Iceland.

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Chapter 1: Introduction

Aim

The aim of this study is to use numerical models to help understand the relationship between glacier fluctuations and climatic changes in Iceland. There are four specific objectives:

- Develop a model of the interactions between glacier dynamics, topography and climate.
- Apply the model to the Holocene fluctuations of Sólheimajökull.
- Use the model to predict the response of Sólheimajökull to climatic warming.
- Investigate the climatic and topographic factors responsible for the spatial variability of Holocene glacier fluctuations in Iceland.

The aim of Chapter 1 is to provide a scientific background to the specific objectives and to explain the approach used to achieve these objectives. The thesis structure is also presented.

Scientific Background

This section provides a scientific background by identifying limitations in present knowledge in the four specific areas listed above. At the end of each sub-section, specific hypotheses are listed. Subsequent chapters test the hypotheses.

Glaciers are very sensitive indicators of climatic change. This is because two components contributing towards the melting process (downward fluxes of sensible heat and longwave radiation) are directly influenced by changing temperature (Oerlemans, 1992). In addition, temperature changes affect the proportion of precipitation falling as snow or rain over a glacier. This has a direct influence on mass-balance and the presence of fresh snow causes a positive feedback in its influence on the surface albedo of the glacier. Glaciers also respond directly to changes in the amount of insolation (hence cloudiness) and precipitation. The response of a glacier front to climatic change can be considered as a two-step process. First, a change in meteorological variables manifests itself as a change in mass balance. The mass-balance perturbation then causes a modification of glacier dynamics, and may result in an advance or retreat of the glacier snout. These processes are illustrated in the classic flow diagram presented by Meier (1965) (Figure 1.1). In the next section, the relationship between local climate and mass balance is considered. This is followed by a description of the factors responsible for the eventual response of the glacier snout.

Climate and mass balance; key concepts

Mass balance defined as the annual sum of accumulation and ablation at a point on a glacier is known to depend mainly on elevation; in the upper parts of a glacier there is an annual net accumulation of snow and the mass balance is positive. In the lower regions where there is an annual net ablation of ice, the mass balance is negative. Interannual variability in precipitation, temperature, cloudiness and other meteorological variables are responsible for the year to year variations in mass balance. As glaciologists, one of our aims is to understand how variability in climate can lead to changes in mass balance (Figure 1.1). This requires the terms and methodology for describing mass balance to be precisely defined. Several useful concepts are available: the *equilibrium-line altitude* (ELA) defines a surface on the glacier where accumulation and ablation are equal over an annual cycle. Although the ELA is a mathematical concept and cannot be directly measured, it can be estimated by the *snowline*, the transition from the previous winters snow pack to bare ice or older snow. In autumn or early winter the snowline moves down-glacier, often in steps or occasionally in one large snowfall that covers the entire glacier. In spring, the snowline migrates up-glacier as melt proceeds. At the end of the summer melt season, the height of the transient snowline marks the approximate location of the ELA.

It was initially thought that the ELA could be linked to simple meteorological variables such as the mean freezing level, but recent work has shown that the relationship is more complicated. This is mainly because much of the melt on glaciers is caused directly by solar radiation. The relationship can also be complicated by precipitation variability, especially relating to processes such as wind drifting of snow.

The most complete way to describe mass balance on a glacier is to consider the *mean specific balance*, defined as the mass balance averaged over the glacier area. When the mean specific balance has a value of zero, the glacier is in a state of balance. In reality, interannual variability in the mean specific balance is large and in the case of a valley glacier, decadal trends in the mean specific balance are largely responsible for any pervasive changes in glacier length, although the glacier snout may respond to shorter climatic variations directly with a modest advance or retreat.

Since the 1970s, detailed analytical and observational work has been undertaken on the physical interactions between climate and mass balance on temperate glaciers (Björnsson, 1972, Munro and Davies, 1978, Kuhn, 1979, van de Wal *et al.*, 1991, Ishikawa *et al.*, 1992). These studies have provided a basic insight into the nature of the energy budget on glaciers. In more recent years, glacio-meteorological experiments have been undertaken where the components of the surface energy flux were measured simultaneously at a number of stations at different altitudes (Oerlemans and Vugts, 1993, Greuell *et al.*, 1994, Oerlemans *et al.*, 1999). These studies have allowed the physical interactions between glaciers and climate to be incorporated into energy balance models, which can allow the different roles of climatic variables on mean specific balance to be assessed (Greuell and Oerlemans, 1986, Oerlemans, 1992).

The response of a glacier to a temperature change depends on the climatic location of the glacier and the rate at which the mass balance on a glacier changes with altitude (Ahlmann, 1948). This latter quantity, known as the *mass balance gradient*, depends on the amount of local precipitation that a glacier receives. Glaciers in high precipitation areas have steep mass balance profiles, and undergo larger shifts in mean specific balance for similar climate changes. Recent studies undertaken with mass-balance models have confirmed this view. In a study of twelve different glaciers, Oerlemans and Fortuin (1992) carried out an analysis to see if a systematic relationship between input parameters to energy balance models (latitudinally dependent insolation, temperature, precipitation, humidity, cloudiness) and mass balance were evident.

The findings show that a significant relationship is found only with precipitation; glaciers in a wetter maritime climate undergo a larger change in mean specific balance for a uniform temperature change than glaciers in a drier continental climate. This can explain regional differences in glacier response. For example, the mass balance of maritime glaciers in the coastal mountains of Norway is much more sensitive to temperature changes than more continental glaciers further inland (Oerlemans, 1992). However, no simple relationship is found between the ELA and temperature, which indicates the importance of glacier geometry, and of using the mean specific balance as an indicator of the state of balance of a glacier. If we are to compare the response of mass balance to climatic changes on glaciers in different areas, then a broad concept is required; the *mass-balance sensitivity* can be defined as the change in mean specific balance for a uniform change in temperature on a particular glacier. In Chapter 3 (Equation 3.1) it is defined mathematically.

The climate of Iceland and mass-balance gradients

Iceland lies near the atmospheric polar front in the path of the storm tracks and the climate is dominated by the west to east movement of North Atlantic depressions (Figure 1.2). Temperature is mild for its latitude due to the moderating influence of North Atlantic Drift waters that extend around the south and western coast. The south coast is especially mild, and it is common for the low areas to be snow free for prolonged periods even during the winter months. In contrast, the north has a colder winter and a greater degree of seasonality. The most persistent winds blow from the south, especially the southeast, and precipitation is highest in the southern parts of the country and lowest in the northeast. Orographic effects are important due to the mountainous topography. Precipitation varies by an order of magnitude, reaching a maximum (in excess of 4 myr^{-1}) on the windward slopes of glaciers near the south coast, such as Myrdalsjökull and Vatnajökull and a minimum (<0.4 myr^{-1}) in the central highlands in the lee of Vatnajökull (Figure 1.3). Precipitation is locally higher on the glaciers than in surrounding areas. For example, Hofsjökull in central Iceland receives c. 2.5 myr^{-1} while the nearby plateau receives <0.8 myr^{-1}. Similarly, glaciers on the Trollaskagi Peninsula in northern Iceland receive c. 2 myr^{-1} while the coastal areas receive c. 0.8 myr^{-1}. The influence of the large gradients in precipitation on Icelandic glaciers is evident in the altitude of the snowline, which increases from about 1100 m on Vatnajökull to above 1700 m on mountains to the north of Vatnajökull, and down again to c. 1000 m on the Trollaskagi Peninsula (Ahlmann, 1948). Björnsson (1979) pointed out the different mass-balance characteristics and dynamic properties of glaciers in southern, central and northern Iceland. He showed that glaciers terminating on the south coast of Iceland have larger mass-balance gradients than the northern outlets of the ice caps. He also suggested that Icelandic glaciers should be very sensitive to changes in temperature.

Constructing hypotheses

Although several authors have pointed out the different mass-balance characteristics of glaciers in south, central and north Iceland, a difference in mass-balance sensitivity has not been quantified. Contemporary theory suggests that glacier mass balance in southern Iceland should be more sensitive to temperature change than glacier mass balance in drier regions of Iceland. The following hypothesis will be tested in this thesis:

- Glaciers located on the south coast of Iceland have a larger mass-balance sensitivity than glaciers in central and northern Iceland.

Topography and glacier dynamics: key concepts

The purpose of this section is to explain the theory of how bedrock topography can influence the dynamic response of a glacier to climatic change. This can be considered as the second major link in chain of events leading to the response of a glacier to climatic change (Figure 1.1). The response of a glacier to climatic change is relatively well understood in terms of glacier physics. In a series of papers that became a landmark in the field of glaciology, John Nye developed the theory on how a glacier responds to a small change in mass balance (Nye, 1960, 1961, 1963, 1965). It was shown that a glacier propagates a mass imbalance down glacier as a series of kinematic waves. As a result, a time delay is apparent between a mass-balance perturbation and a later time when the glacier has mostly adjusted to the new mass balance state. The response of a glacier to a climatic perturbation usually follows an exponential variation, hence it has become popular to define the period of this readjustment as a time constant that varies from glacier to glacier. This is known as the *response time*; it can be expressed in terms of glacier length (*length response time*) and glacier volume (*volume response time*). It is mathematically defined in equations 3.5 and 3.6, Chapter 3. The response time, usually several decades for large valley glaciers, is distinct from the time it takes for a glacier snout to react to a climatic change (*reaction time*); this is shorter, and it depends on the mass-balance history of the glacier in question. If the glacier is in equilibrium prior to a mass-balance perturbation then the reaction time will be zero.

One of the interesting properties of glaciers is that they do not respond uniformly to a climatic change. Large glaciers resting on small bed slopes tend to amplify the length response through mechanisms such as the *height mass balance feedback* (Oerlemans, 1980), where an increase in snowfall increases the surface elevation of a glacier resulting in a positive feedback loop which causes greater accumulation. The topography of the underlying trough also plays an important role, especially in the case of valley glaciers and small ice caps. Mercer (1961) first observed that the amplitude of glacier response can depend on the topography of a glacial trough. He identified the importance of bed slope on glacier response, and found that fjord glaciers undergo large changes in glacial extent because they advance and retreat along troughs with a small longitudinal bed profile. The seminal study of the influence of bed geometry on glacial response was that of Furbish and Andrews (1984). They used topographic hypsometry (the area-altitude profile of a glacier) to examine the stability and response of glaciers to changes in mass balance.

In recent decades, the physical theory of glacier flow has been incorporated into numerical models that are capable of simulating glacier dynamics. If models are well calibrated with empirical evidence, they provide a means of testing hypotheses on how topography influences the response of glaciers to climatic change. Several modelling and field studies have highlighted the influence of bed topography on glacier response (Oerlemans, 1989, Clapperton *et al.*, 1989, Payne and Sugden, 1992, Hubbard, 1997, Hulton and Sugden, 1997). Kerr (1993b) provides an authoritative review of the subject. The main findings are:

- The magnitude of change in glacial extent depends on the geometry of the glacier bed (Furbish and Andrews 1984, Oerlemans, 1989). For example, glaciers with small bed slopes undergo larger changes in glacier extent. Glaciers in widening troughs undergo smaller changes in glacial extent than glaciers in straight or narrowing troughs.
- Glaciers may become partially decoupled from climate if they are pinned at stable topographic positions (Mercer, 1961). This can result in two glaciers in close proximity to each other fluctuating differently to the same climatic change (Hubbard, 1997).
- Different steady state positions are possible for the same climate if the glacier bed is overdeepened. In this case, the glacial extent depends on the history of mass-balance changes (Oerlemans, 1989, Hulton and Sugden, 1997).

These findings can be used to differentiate topographic characteristics which enhance changes in glacier length from those that inhibit advance and retreat. In turn, this knowledge can be used to interpret the pattern of glacier fluctuations across a region (Table 1.1). It is useful to define a topographic feature that stabilises the position of a glacier snout as a *topographic threshold.* A topographic threshold in the glacier-climate-landscape system can be thought of as a type of geomorphic threshold (Schumm, 1979) that will cause a glacier (the landform) to halt an advance at a point of topographic change (the threshold), without a change in external forcing (climate). An example is a break in slope at the base of a mountain. Mountain glaciers often terminate at this boundary because further advance results in a lateral spreading and a large increase in the ablation area. Glaciation of these lowland areas is only sustainable in a very cold climate (Payne and Sugden, 1990). Glaciers are likely to remain stable at the topographic threshold for a range of climates (Hulton and Sugden, 1997).

Table 1.1: Topographic characteristics that enhance and inhibit glacial response to a climatic change.

Topographic feature	Reasoning
Enhance changes in glacier length	
A low-gradient longitudinal profile free of large undulations.	Altitude-mass-balance feedback is enhanced (Oerlemans, 1989). An initial increase in ice thickness leads to more snowfall and further increases in thickness. Thus the glacier may amplify a small change in climate.
A wide, flat accumulation area just above the ELA, and a confined ablation area descending to low altitude in a valley of uniform width.	Hypsometry effect - large change in accumulation area occurs for a modest shift in ELA (Furbish and Andrews, 1984). Confined ablation area channels ice to low altitude where melting rates are high. This enhances the mass-balance sensitivity.
Glacier terminates in a valley.	Hypsometry effect- snout must move up or down valley in order to respond to changing mass-balance.
Decreased sensitivity- topographic thresholds	
Widening channel due to changes in valley width near the snout.	Hypsometry effect- further advance greatly increases ablation rate as the glacier snout swells into a piedmont lobe (Furbish and Andrews, 1984, Hubbard, 1997).
Break in slope near the snout.	Bedrock hills, moraines or abrupt changes in slope may form a barrier to advance (Burbank and Fort, 1985). Overdeepening may induce hysteresis and lead to possibility of multiple steady states (Oerlemans, 1989).

Icelandic glaciers and topography

Icelandic topography is relatively simple in comparison to topography developed on continental crust. There are two main geological zones and each has a different suite of landforms (Figure 1.3) (Thorarinsson *et al.*, 1959). The older group consists of horizontal sheets of basalt. They form a plateau of 500 m to 1000 m dissected towards the coast by deep glacial toughs and fjords. The younger geological group contains an active volcanic zone of high relief with summits in excess of 1000 m. The landscape consists of stratovolcanoes such as Oerafajökull (2200 m), table mountains and shield volcanoes surrounded by lava and sandur plains. The topography is young and evolving dynamically. Deep glacial trough are lacking, although the older volcanoes are cut by fluvial gorges especially where relief and precipitation are high and fluvial erosion has occurred during jökullhlaups. The largest glaciers such as Hofsjökull, Myrdalsjökull and parts of Vatnajökull are located in the younger volcanic zone and form on volcanoes (Björnsson, 1988). Outlet glaciers descend from high plateau ice fields to expanded piedmont lobes on sandur plains or unconfined termini spread over mountain slopes. A few outlet glaciers are more confined by topography and have a similar form to valley glaciers. They occur when ice caps intersect the older geological areas and on dissected volcanoes where the glaciers breach caldera rims. Mostly they descend to the sandur plains. Only a few of the confined outlet glaciers terminate within their valleys. Over 150 cirque glaciers are found in the older geological areas in northern Iceland, and they form below the plateau surface extending a short distance down valley.

Constructing hypotheses

The relationship between topography and glacier response has not been studied in Iceland. Yet it has been shown that many glaciers in Iceland terminate on sandur plains, and are relatively unconfined by topography. In contrast, some glaciers are well confined by topography and terminate in valleys. Glaciological theory suggests that these glaciers will respond differently to climatic changes. Two hypotheses can be stated:

- Glaciers confined in valleys will undergo a larger change in length in response to climatic change than broad ice cap lobes.
- Glaciers terminating on sandur plains occupy stable topographic positions and will be insensitive to further advance.

Reconstructing the magnitude and timing of Holocene climatic changes

A detailed history of Holocene climatic changes is emerging in Iceland, mainly due to the success of tephrochronology in dating Holocene moraines. Periods of Holocene glacier expansion are now known from the periods c. 5000 and c. 3000 years BP, during the 6th, 10th, and 14th centuries AD, and between 1600 and 1900 AD (Dugmore, 1989, Gudmundsson, 1998, Stötter *et al.*, 1999). However, work is still required in order to understand how these glacier fluctuations relate to global changes. The main uncertainty lies in our understanding of the timing and magnitude of climatic changes in Iceland, and the relationship between wider changes. For example, the wider Holocene climate is believed to have experienced millennial-scale variability with alternating periods of warming and cooling. Cool intervals reccurred at 1500 to 2500 year intervals (Denton and Karlen, 1973, O'Brian *et al.*, 1995, Bond *et al.*, 1997, Bianchi and McCave, 1999). Yet little is known about how these changes influenced atmospheric circulation and ocean currents around Iceland. The most recent period of Holocene cooling, the *Little Ice Age* (defined as the period between AD 1600 and 1900; Grove, 1988), is known from documentary sources to have been climatically variable in Iceland and its existence has even been questioned (Ogilvie, 1992). Climatic variability seems to have occurred on a decadal-scale during the Little Ice Age. Critically, the coldest decades are known to have involved a large increase in sea ice incidence around Iceland (Ogilvie, 1984, 1991, 1992, 1996). This implies that the ice-carrying ocean currents have had a variable but important effect on the Icelandic climate for at least several centuries.

The 20th Century climate of Iceland has also been characterised by large decadal shifts in temperature. The cause of these temperature changes is complicated, but it is known that changing ocean surface temperatures have a role (Einarsson, 1991). Polar ocean waters and sea ice near Iceland have a direct cooling effect, but also influence temperature indirectly because all air masses arriving in Iceland pass over the sea. The cool period from 1965-1971 illustrates this relationship and the movement of ocean waters and resulting climate changes are well documented. The climate of Iceland is strongly influenced by the strength of the East Iceland Current, a branch of the East Greenland Current, which is the main southward flow of water from the Arctic Ocean (Figure 1.2). This flow is not always dominated by polar water. Icelandic sea surface temperature and salinity observations conducted in June each year indicate that between 1948 and 1961, the sea area between 67 and 69°N was dominated by warm saline North Atlantic Drift water (high salinity and temperatures of 0 to +2°C) (Lamb, 1979). This water mass was sourced from North Atlantic Drift water that had come around the north of Iceland via the Irminger Current (Figure 1.2). In contrast, after 1962 this area was penetrated to a varying extent by polar water (lower salinity and temperatures down to -1.8°C). This was known as the *Great Salinity Anomaly* and was characterised by an advection of polar water and sea ice beyond its normal bounds to the north and east coasts of Iceland (Mysak and Power, 1991). Sea ice was recorded off the east coast of Iceland in abundance during the late 1960s and early 1970s.

There is evidence to suggest that the Great Salinity Anomaly was a single event in decadal-scale climatic oscillation of the atmosphere sea ice ocean system that has been in existence for some time (Mysak and Power, 1991). Recently, a link has been made between sea-ice anomalies in the Greenland Sea and extreme phases of the North Atlantic Oscillation (Mysak and Venegas, 1998). The North Atlantic Oscillation is a large-scale alternation of atmospheric mass between the Icelandic low and the Azores High, which is an important source of seasonal to decadal climatic variability in the North Atlantic (Cook *et al.*, 1998). It appears that negative phases of North Atlantic Oscillation are associated with a strong East Greenland Current, an extensive ice cover in the Greenland Sea and weak Icelandic Low. During these phases, a sea-ice expansion occurs in the ocean around Iceland. Conversely, positive phases of the North Atlantic Oscillation are associated with a weak East Greenland Current, a low sea-ice cover in the Greenland Sea and a vigorous Icelandic Low (Mysak and Venegas, 1998). In the intervals between these two extremes, sea ice propagates in a clockwise direction around the arctic basin, reappearing in the Greenland Sea every 10 years.

There is evidence to suggest that there are periods when this 10 year climate cycle was more and less evident. For example, sea ice conditions were severe off the coast of Iceland during the first two decades of the 20th Century, and during the Great Salinity Anomaly in the 1960s and 1970s. In contrast, sea ice was rare in the 1930s (Mysak and Power, 1991). The Koch sea ice severity index, which is the number of weeks per year when sea ice affects the coast of Iceland, seems to be a good indicator of the periodicity and severity of sea-ice anomalies and there is a good correlation between this index and regional changes in temperature and pressure around Iceland (Kelly *et al.*, 1987).

Constructing hypotheses

Although the pattern of Holocene glacier fluctuations in Iceland is becoming well known, we still know little about how glacier fluctuations relate to climatic changes. The following hypotheses can be stated:

- Glacier expansion in Iceland occurs during cold periods when sea ice surrounds the Icelandic coastline.
- Climatic variability in Iceland has resulted from changes in ocean waters and atmospheric circulation, on a timescale of decades to centuries.

Predicting the response of Icelandic glaciers to climatic warming

There has recently been an interest in predicting the response of valley glaciers and small ice caps to possible anthropogenically induced climatic changes. This is because they respond more quickly than the large ice masses in Greenland and Antarctica, and despite locking up only a small amount of ice globally (c. 0.5 m sea level equivalent), they may make a significant contribution to sea level rise over timescales of decades to centuries (Warrick *et al.*, 1996). Many glaciers have retreated worldwide since the late 19th Century and glacier meltwater is believed to have contributed several centimeters toward an overall rise of 10-20 centimeters over the last 100 years (Meier, 1984, Oerlemans and Fortuin, 1992). One of the problems in calculating the contributions of valley glacier melt to sea level rise is that the climatic sensitivity of individual glaciers varies over at least one order of magnitude (Oerlemans *et al.* 1998). The climatic sensitivity of a glacier depends on its geometry and its local climatic regime, and it is possible for adjacent glaciers to respond differently. There is presently no consensus on how to include the individual dynamic characteristics of glaciers in a generalized scheme and it is agreed that more glaciers need to be studied in detail as basic information on the response of glaciers with different geometry and climatic settings is still lacking.

The response of Hofsjökull to a climatic warming scenario for Iceland has recently been calculated with coupled mass-balance ice-flow model (Johannesson, 1997). It was shown that the glacier responds strongly to a warming of c. 0.3°C per decade. Up to 40% of the initial ice volume is projected to melt by the year AD 2100, and the glacier essentially disappears by the year AD 2200.

Constructing hypotheses

The prediction of future changes in glacier length and volume is still at an early stage. Presently several approaches may be used and it is not certain which provides the most reliable estimates. This is because glacier response depends on the climatic and topographic setting of each individual glacier. The response of one Icelandic ice cap to climatic warming has been predicted. It is not known to what extent this finding can be considered representative in Iceland as a whole. It has already been argued that valley glaciers located on the south coast of Iceland are likely to experience a larger response to climatic changes than broad lobate outlet glaciers in central Iceland. Thus the following hypothesis will be tested in this thesis.

- Valley glaciers located on the south coast of Iceland will undergo larger changes in length and volume in response to climatic warming than ice cap lobes in central Iceland.

Understanding the spatial variability of Holocene glacier fluctuations in Iceland

One of the intriguing aspects of the Holocene moraine chronology in Iceland is that the maximum Holocene glacier extent occurred at different times in different regions, and that the scale of Holocene advances varied markedly in different areas. Until recently it was believed that most Icelandic glaciers reached their maximum extent during the 19^{th} Century (Thorarinsson, 1943, Björnsson, 1979). Dugmore (1987, 1989) showed that the Holocene extensions of Sólheimajökull displayed a very different pattern. The largest Holocene advances occurred during the mid-Holocene, and during this time the glacier was extended by 4-6 km beyond contemporary ice limits. Dugmore and Sugden (1991) suggested that the large fluctuations of Sólheimajökull resulted from ice-divide migration on Myrdalsjökull, where the mid-Holocene glacier drained a larger proportion of the ice cap than it did in recent times. This was explained conceptually as the behavior that might be expected in the growth cycle of a maritime ice cap with a specific sub-glacial topography.

In the last decade, the forelands of several other glaciers have been found to contain evidence of a mid-Holocene maximum glacier extent (Gudmundsson 1998, Stötter *et al.*, 1999). In these cases, the scale of the mid-Holocene advance was always less than at Sólheimajökull. The outlet glaciers of Oerafajökull were shown to have extended by 2-3 km during the mid-Holocene, while the cirque glaciers of northern Iceland advanced only slightly beyond their Little Ice Age moraines, typically 1 km from the present glaciers. At present it is unclear why Icelandic glaciers fluctuated in an asymmetric pattern, or why some glaciers do not show evidence of mid-Holocene fluctuations.

Constructing hypotheses

It is possible that the glaciological theory explained so far may explain the different response of Icelandic glaciers to Holocene climatic changes. Two hypotheses may be stated:

- The Holocene fluctuations of Sólheimajökull reflect climatic change rather than ice-divide migration. The large extent may be explained by an especially favourable climatic and topographic setting.
- The spatial variability of other Icelandic glacier fluctuations can be explained by local differences in climatic and topographic setting. This may also explain why some glaciers appear to have reached their maximum extent during the 19^{th} Century, while others display a more complete Holocene moraine record.

Approach

Numerical models are used as a hypothesis-testing tool in this thesis. The key to a successful modelling study is to use models that are of similar resolution to the field evidence in question, and to include the most important physical process. Many options are now available in terms of the type of model that can be used. Hughes (1995) and Paterson (1994) provide reviews of the subject. One approach is to use the most complex model possible. For example in a recent study of a valley glacier Hubbard *et al.* (1998) showed that it was possible to simulate the stress field evident in the pattern of crevasses. This simulation required the glacier stress field to be resolved in three dimensions at 30 individual levels. While this is of great use for studying dynamic processes in detail, on some glaciers it may not be possible to provide realistic initial boundary conditions and empirical data for testing the models. This is a particular problem in Iceland where few modelling studies have been undertaken, and little is known about Icelandic glaciers in general. Mass-balance programs have only recently been initiated on Vatnajökull, and at Hofsjökull they have been made for 12 years. Ice-radar studies have been carried out on Hofsjökull, Vatnajökull and Myrdalsjökull (Björnsson, 1988, Björnsson, 1996). At this stage, it is difficult to select a glacier for study with a complex model. The only possible candidate is Breidamerkurjökull, where the bed is well known (Björnsson, 1996) and a glacio-meteorological experiment has recently been undertaken (Oerlemans *et al.*, 1999). A problem with Breidamerkurjökull is that it shares an ice divide with the surging glaciers of northern Vatnajökull. Attempts to simulate the ice dynamics have so far been problematic (R. Hindmarsh pers. comm., 1996).

Another way of approaching the problem is to use models of intermediate complexity. In the case of glacier flow, one-dimensional flow-line models have been used to successfully simulate the 19th and 20th Century length fluctuations and ice surface profiles of glaciers in the European Alps and Norway in cases where little was know about ice velocity, basal flow conditions, or the strain distribution over the glaciers (Greuell, 1989, Oerlemans 1997a, Zuo and Oerlemans, 1997, Schmeits and Oerlemans, 1997). Such models are ideal for testing hypotheses relating to topography because they include valley parameterisations at high spatial resolution (generally grid spacing is 100 to 300 m). Another advantage is that a flow-line model has already been used to make projections of future glacier length for Hofsjökull (Johannesson, 1997). An energy-balance model of intermediate complexity is used in this study. The model treats the most important processes with rigour, specifying incoming and outgoing radiation, and turbulent exchange and allowance is made for altitudinal gradients in temperature and precipitation. However, complex problems such as the calculation of radiation terms for a three-dimensional ice surface, and turbulent fluxes from profiles of wind velocity at different altitudes and surface roughness are not attempted.

Sólheimajökull case study

The main themes of this study involve identifying contrasting behavior of glaciers in different climatic and topographic settings. It is also important to select glaciers for study which might be clear indicators of climatic change. Sólheimajökull is chosen for detailed study for several reasons;

- It is a valley glacier on the south coast of Iceland (Figure 1.4) and a comparison can be made with Hofsjökull in central Iceland.
- Sólheimajökull has a hypsometry that is typical of a glacier which undergoes large changes in glacier length in response to a climatic change (Figure 1.5).
- Much information is available on the past fluctuations of Sólheimajökull.
- A study of Sólheimajökull allows a test of the ice-divide migration hypothesis put forward by Dugmore and Sugden (1991).
- The 20th Century fluctuations of the glacier are believed to have followed climatic changes closely (Sigurdsson and Jonsson, 1995).

The starting point of this study was the collection of data on the sub-glacial topography of Sólheimajökull. This was carried out in a GPS and Ice Radar survey of the glacier in 1996 and 1997, the results of which are shown in the Appendix.

Sólheimajökull drains two domes (1500 m at highest point) of the Myrdalsjökull ice cap in southern Iceland through a valley 1 to 2 km wide (Figures 1.4 and 1.5). The glacier is up to 6 km wide in its accumulation zone and has a long narrow snout descending south to 100 m above sea level. In 1996, the glacier was 14.3 km long from ice divide to snout and covered an area of 44 km^2. Ice thickness reached a maximum of 430 m and volume was approximately 3 km^3. Sólheimajökull has once of the richest records of past fluctuations in glacier length in Iceland. Small length variations have been measured by the Icelandic Glaciological Society since 1932 (Sigurdsson, 1998). Glacier maps are available for 1904 (Thorarinsson, 1943), 1948 (AMS Map, 5717 I) c. 1970 (DMA Map, Myrdalsjökull 1812) and 1996 (Mackintosh *et al.*, 1999). Eight terminus positions are known from before 1932, the earliest dating from AD 1705 (Bardarsson, 1934, Ahlmann and Thorarinsson, 1940, Thorarinsson, 1943) (Figure 1.6). The most notable features of the 18^{th}, 19^{th} and 20^{th} Centuries are the large retreat and advance in the late 18^{th} Century, the relatively stable position during the 19^{th} Century, and the retreat between 1930 and 1970 followed by an advance during the last few decades.

Knowledge of earlier Holocene fluctuations of Sólheimajökull comes from geomorphological and tephrochronological evidence (Dugmore, 1989) (Figure 1.7). The oldest known Holocene fluctuation is marked by a till sheet dated between 4500 and 7000 ^{14}C years BP. At this time, the glacier was 20 km^2 larger than it is today and it was c. 19 km long from the present ice-divide to terminus. Another till sheet and moraine limit dated to c. 3430 years BP indicates a glacier length of 18 km. In addition, a suite of lateral moraines moraine dated to c. 1380 calibrated years BP (AD 570) indicates an expansion of c. 10 km^2 relative to the present, and a glacier length of 17.6 km. Two glacier positions have been attributed to early historic times in Iceland. The earliest indicates a glacier 17 km long at c. AD 920-930, while the later marks a glacier 16 km long at c. AD 1357. No information on glacier length is available between c. AD 1350 to 1705.

Thesis Structure

Each chapter of this thesis attempts to fulfill a specific objective:

- The development and testing of a coupled mass balance glacier-model for Sólheimajökull is considered in Chapter 2.
- Basic experiments are carried out with the model in order to identify the glacier sensitivity to climatic change and topography in Chapter 3.
- An inverse reconstruction of Holocene climatic changes with the ice flow model is made in Chapter 4.
- A projection of future glacier length to climatic warming scenarios with the ice-flow model is carried out in Chapter 5.
- A classification of glaciers in terms of their response to climatic change is developed in Chapter 6. The classification scheme is tested against the record of Holocene glacier fluctuations in Iceland.
- The pattern and forcing of Holocene climatic changes in Iceland is examined in Chapter 7.
- The contributions of the thesis and future research directions are identified in Chapter 8.

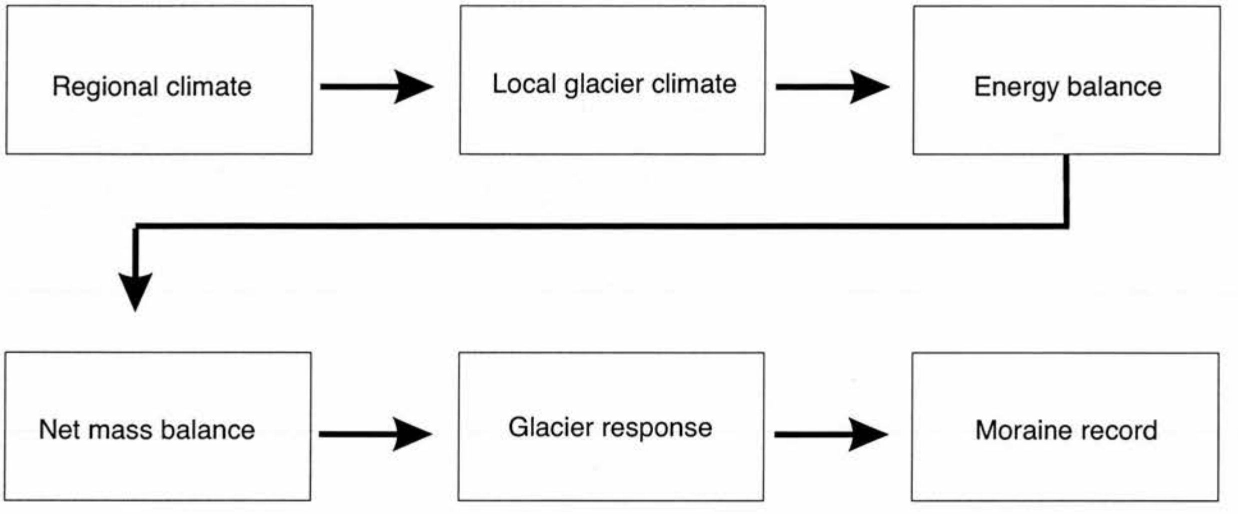


Figure 1.1

The links on an individual glacier between a change in the regional climate and the moraine record (After Meier, 1965). The two linkages considered in this study with models are between the local glacier climate and the mass balance, and between the mass balance and the glacier response.

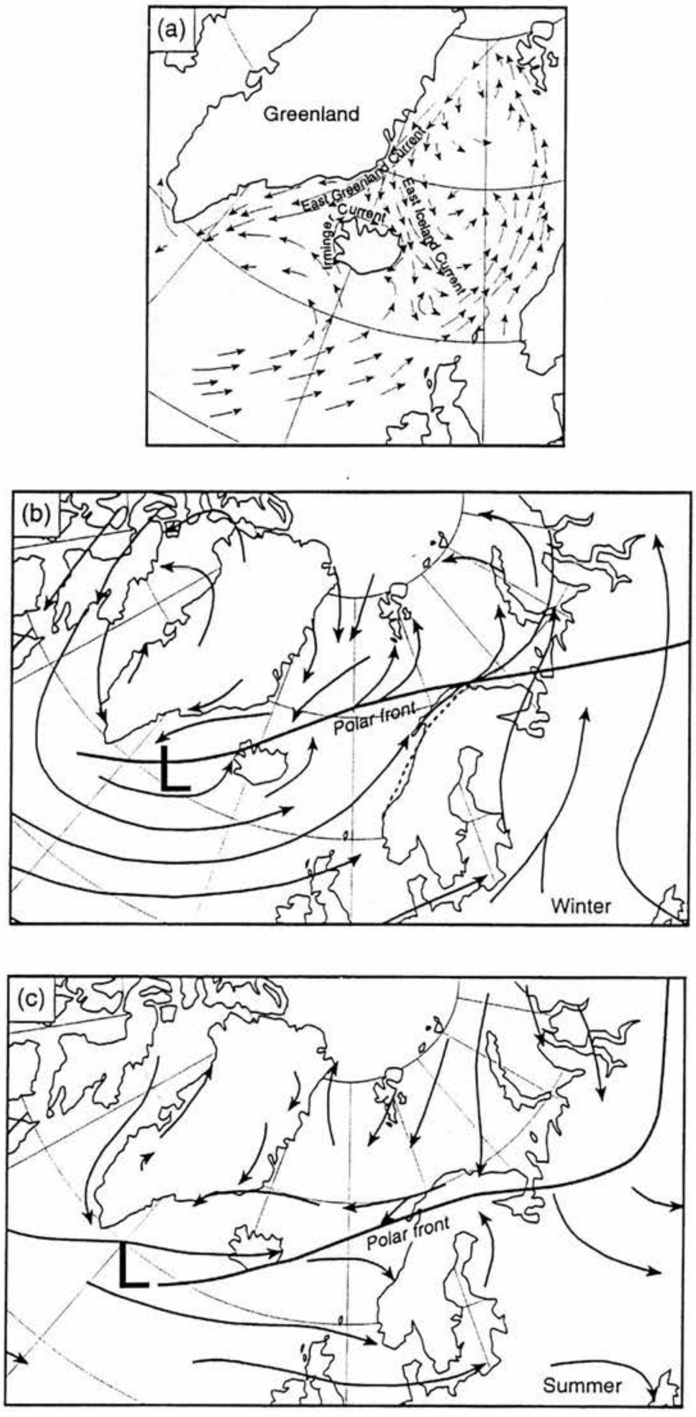


Figure 1.2

The upper chart (a) shows mean oceanic conditions around Iceland. The East Greenland and East Iceland currents are marked. North Atlantic Drift waters occur to the south of Iceland.

The lower two charts show mean atmospheric circulation conditions around Iceland in winter (b) and summer (c). These conditions prevail during the positive phase of the North Atlantic Oscillation when the climate of Iceland is dominated by a vigorous Icelandic Low (marked L).

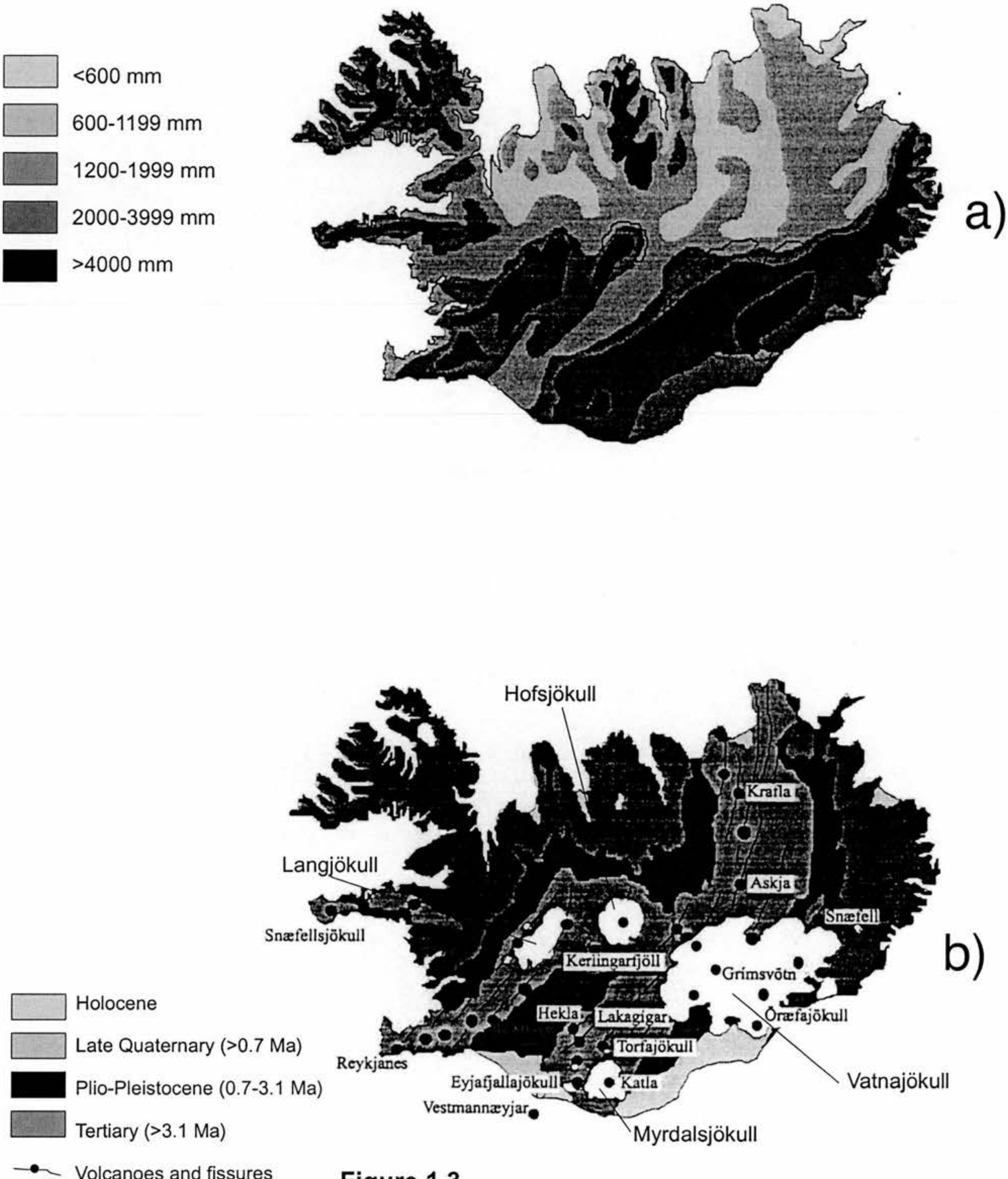


Figure 1.3

(a) Precipitation totals in Iceland showing the strong gradients. (b) The geology of Iceland. The youngest rocks occur in the active volcanic zone (Late Quaternary). Here the topography is young and evolving dynamically. The older group (Tertiary) consists mainly of a basalt plateau. The location of the main ice caps is also shown.

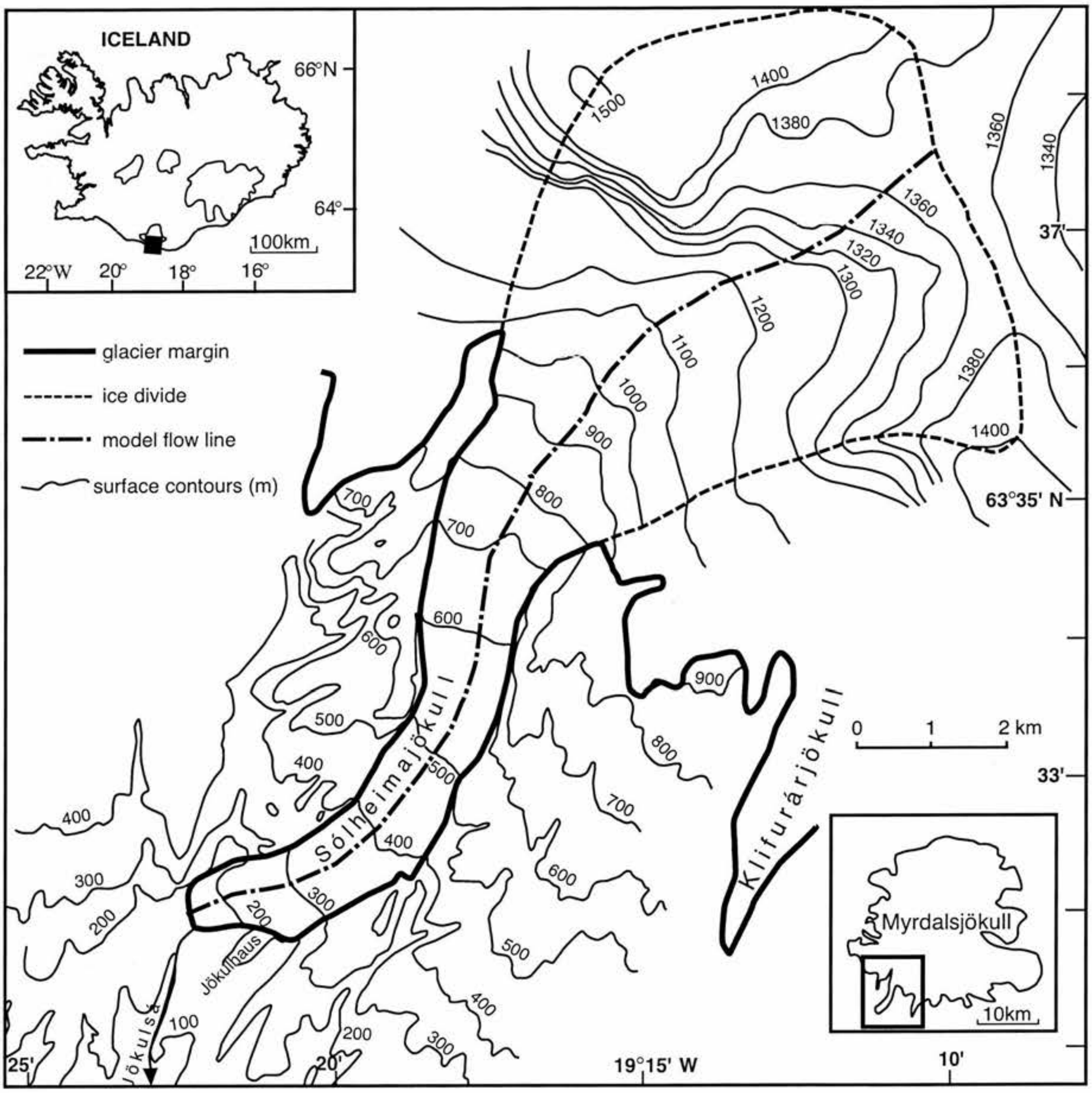


Figure 1.4

The topography of Sólheimajökull in 1996. The position of the flow line used in the modelling study is marked. The location of Sólheimajökull as an outlet of Myrdalsjökull in southern Iceland is also shown.

a

b

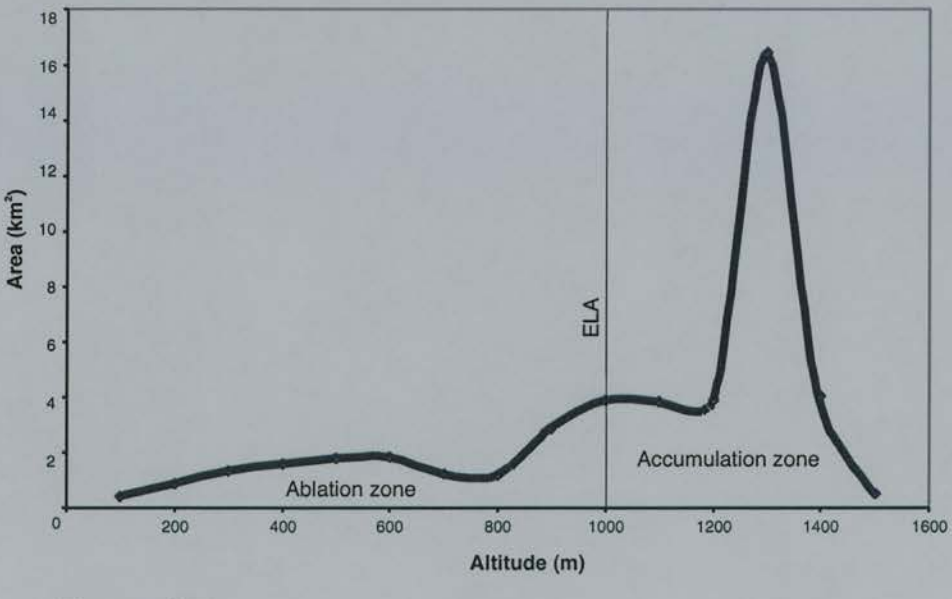


Figure 1.5

a. Oblique aerial photo of Sólheimajökull in 1985. The glacier flows from the ice cap Myrdalsjökull down a valley 1-2 km wide. Moraines from the Little Ice Age and earlier cool periods from the Holocene are circled.

b. The hypsometry of Sólheimajökull. During a typical year with an ELA of 1000 m, most of the glacier area is located in the accumulation zone.

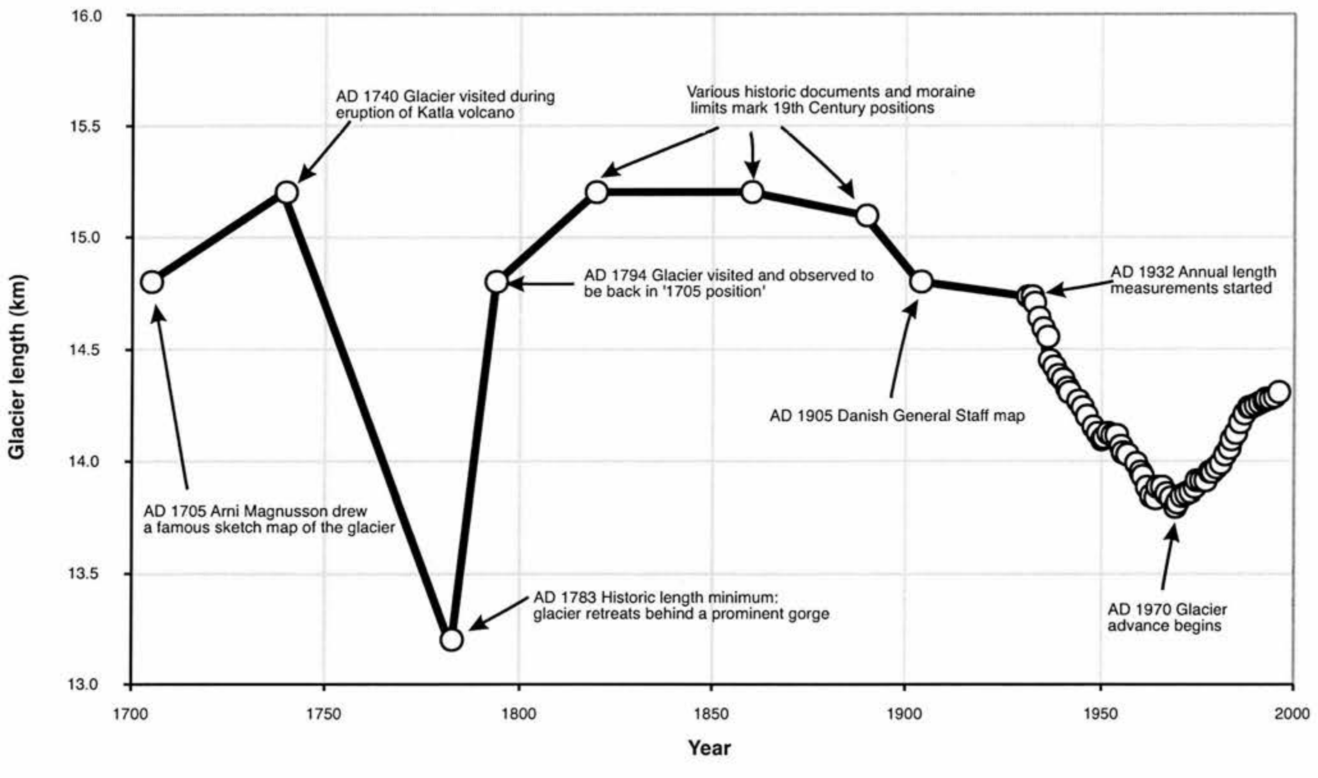


Figure 1.6

Length variations of Sólheimajökull from AD 1705-1998. Data from Bardarson (1934), Ahlmann and Thorarinsson (1940) and Thorarinsson (1943).

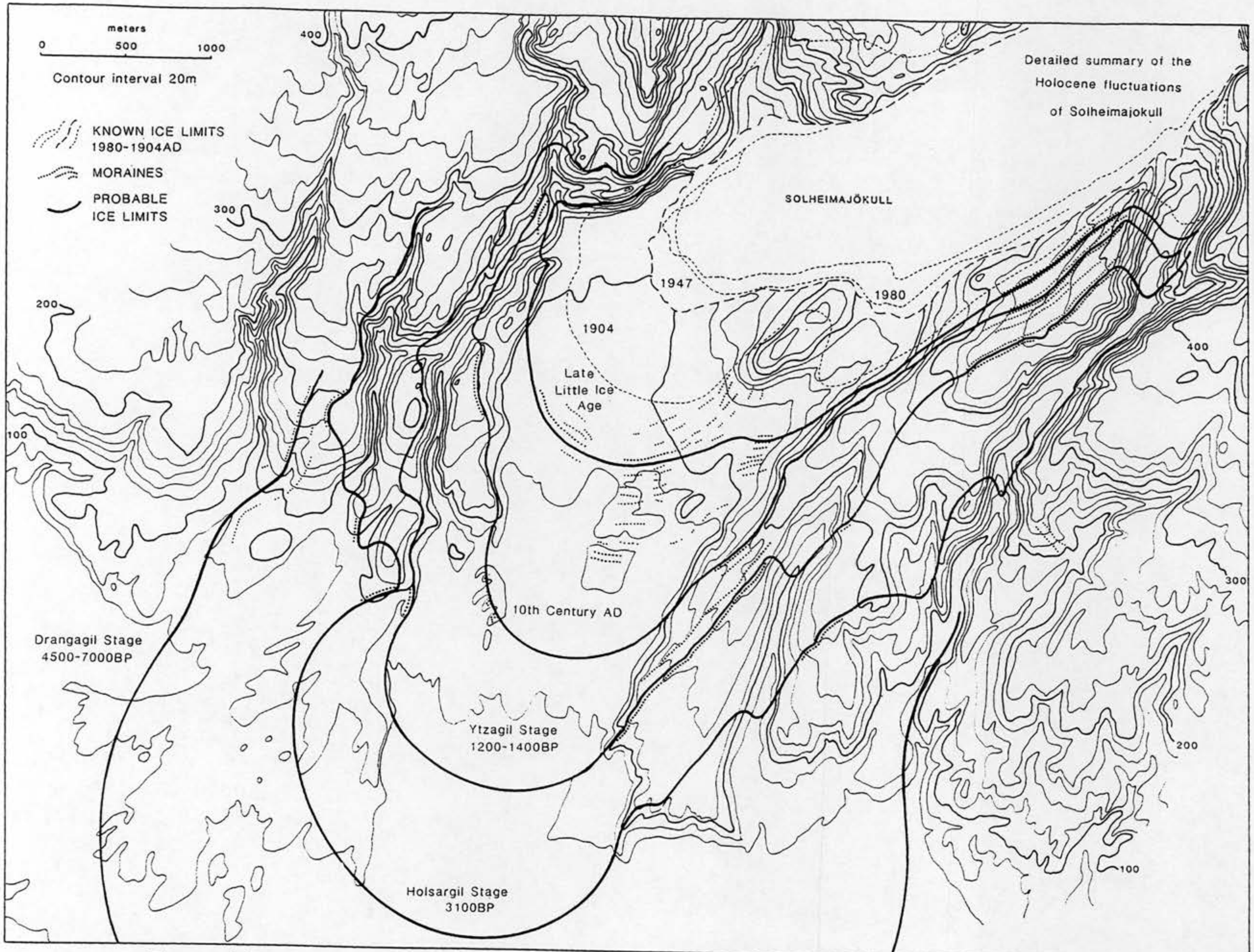


Figure 1.7

A summary of the Holocene fluctuations of Sólheimajökull. After Dugmore (1987).

Chapter 2: The Models

Introduction

The aim of this chapter is to develop a coupled mass-balance glacier-flow model, which is applied to Sólheimajökull. The emphasis is to describe the physics of the models and their tuning and testing. Additionally, one section deals with the method used to couple the mass-balance model to the flow model.

The first part of the chapter concerns the development of a mass-balance model. There are two approaches to simulating mass-balance in contemporary glaciology. The *degree-day method* is based around the assumption that from a statistical point of view, there is a high correlation between days with a positive temperature (in °C) and ablation (Braithwaite and Olesen, 1985). This relationship has allowed models to be developed where ablation is calculated from summer air temperature using an empirical regression equation (Eg. Reeh, 1991, Johannesson, 1997). A 'degree-day factor' is used to account for the different ablation rates over snow and ice, which in reality are due to variation in albedo. The advantage of such models is that they can be used in areas where little is known about meteorological conditions over the ice, and that they are relatively simple and involve short computational times.

A more rigorous approach is to calculate all the terms in the energy balance, including radiation, turbulent exchange, and albedo (Oerlemans, 1992). The advantages of this *energy-balance method* is that the physical processes are described, so it is possible to carry out sensitivity studies to show which factors are most important (Kerr, 1993a). There is also greater potential to tune the model with data derived from glacio-meteorological experiments. The disadvantage is that computation of energy-balance over a whole glacier is complicated by errors resulting from extrapolation of input data to sites at a distance from measurement locations. There are also limitations in simulating short-term variations in surface albedo, turbulent exchange and the radiation balance because in nature these processes have large spatial and temporal variability.

In this study, an energy-balance model is used in preference to a degree-day model for several reasons: Data on albedo variability in the ablation zone is available for Sólheimajökull, and this means that coefficients in an albedo scheme can be tuned to mimic reality. Secondly, use is made of data from a recent (1996) glacio-meteorological experiment undertaken on Breidamerkurjökull, an outlet glacier of Vatnajökull located approximately 200 km to the east of Sólheimajökull (Oerlemans *et al.*, 1999). The Breidamerkurjökull data is used to help tune the Sólheimajökull model. This is justified because both glaciers have a similar latitude, proximity to the sea, altitudinal distribution and aspect. Another advantage of using an energy-balance model is that there is potential for improvement in the future, when more data becomes available or when our understanding of the physical processes increases.

The second part of the chapter concerns the development of the ice-flow model, and the coupling of the mass-balance model to the ice flow model. The emphasis is on testing the ice-flow model using a method known as dynamic calibration. This is where simulated ice surface profiles are compared to glacier profiles at different times in a transient modelling run. This is possible at Sólheimajökull because we have a good record of changes in glacier length and know the ice surface profiles at four dates in the 20th Century.

The Mass-Balance Model

The mass-balance model is similar to that used in Oerlemans (1992). The aim is to calculate the mass-balance profile of Sólheimajökull for individual years by integrating the energy balance equation for the ice/snow surface over a mass-balance year. It takes meteorological data from a nearby climate station and the area distribution with altitude of the ice mass as input values. The model simulates the winter months first when there is net accumulation of snow and the summer ablation season second when the transient snowline migrates up the glacier. This allows the three main properties of glacier mass balance to be calculated at the end of the ablation season; the distribution of mass balance with altitude, the mean specific balance and the equilibrium-line altitude (ELA). The governing equations are:

Mass balance $$B = \int_{year} (-F_{\downarrow} + P)\partial t \qquad (2.1)$$

Energy balance $$-F_{\downarrow} = (1-\alpha)G + L_{\uparrow} + L_{\downarrow} + H_{se} + H_{la} \qquad (2.2)$$

Where $F_{\downarrow}$ is the energy flux at the surface, P is the rate at which solid precipitation is added to the surface, α is the surface albedo, G the global radiation, $L_{\uparrow} + L_{\downarrow}$ the up-welling and down-welling longwave radiation and $H_{se} + H_{la}$ the turbulent fluxes of sensible and latent heat respectively. B is expressed in m.w.e.yr^{-1}. The refreezing of meltwater is not included in the model and meltwater is assumed to run off. This is reasonable given the temperate nature of the glacier. The different components of the energy budget in Equation 2.2 are calculated using simple schemes developed in boundary layer meteorology. The following paragraphs describe how they are calculated.

The schemes used to calculate incoming short and long wave radiation are unchanged from Oerlemans (1992) as they can simulate an altitudinal distribution of net radiation that is similar to measured net radiation on Breidamerkurjökull (Figure 2.1). Incoming short wave radiation at the top of the atmosphere is found using the method of Walraven (1978) and allowance is made for the solar zenith angle and surface elevation of the glacier. The geometry of the ice surface is not included because it is considered too complicated in comparison with the quality of the data available for testing the model. The partitioning between direct and diffuse radiation depends linearly on daily cloudiness as this is an important process in southern Iceland. It is not uncommon for cloud cover to be in excess of 80% around Sólheimajökull which can result in most radiation arriving in diffuse form. The parameterisations of cloud transmissivity are taken from Oerlemans (1992) and are based on experimental results from the Alps conducted by Sauberer (1955). The long-wave energy component is calculated as the balance between the outgoing radiation from the glacier surface and that incoming from the atmosphere. Outgoing long-wave radiation is set to 316 Wm^{-2}, a figure equivalent to that emitted by a black body at the melting point. Incoming long wave radiation is derived from two sources, the lowest tens of metres of the clear sky and from the base of clouds

(Oerlemans, 1992). In the absence of measurements, the cloud base is assumed to lie at an altitude of 900 m, a typical value in Iceland (Eythorsson and Sigtryggsson, 1971).

Albedo is generated internally by the model as a function of altitude and time since the last snowfall event (Oerlemans, 1992). This allows changes in albedo associated with the seasonal migration of the snow line to be simulated. A background albedo profile (α_b) is defined by the assuming that the albedo profile is linked to the ELA. This is the albedo profile to be expected at the end of the melt season, and represents the transition from snow and clean ice at higher altitudes to dirty ice and debris near the terminus. Brock (written comm., 1996) has measured albedo on the glacier and the measurements are used to set the coefficients in the background albedo equation in order to simulate realistic albedo values in the ablation zone:

$$\alpha_b = 0.115\text{ATAN}[(\text{altitude-ELA}+300)/200)] + 0.43 \qquad (2.3)$$

The actual albedo is calculated by perturbing the background albedo profile depending on the presence of unmelted snow. Figure 2.2 shows how the albedo profile of Sólheimajökull is simulated through a mass-balance year. At day 260 when most melting is complete, the albedo profile approaches the background albedo profile. In early winter at day 360 there is a transition between 500 and 700 m from an ice albedo to a snow albedo, while the late winter albedo profile at day 80 is dominated by the albedo of fresh snow (0.75). The calculation of turbulent flux is a source of uncertainty in mass balance modelling and a field of ongoing research (e.g. Denby, 1996). The bulk transfer method is used in this study where turbulent flux is proportional to the difference in temperature/humidity at a standard measuring height (2 m) and the ice surface. Surface roughness, atmospheric stratification and wind speed, all of which influence turbulent exchange and change significantly in space and time, are not considered explicitly. In light of this uncertainty and the recent finding that the microclimate of Vatnajökull (in summer) is strongly shaped by katabatic flow, the exchange coefficient is tuned with data from the Vatnajökull experiment (Oerlemans *et al.*, 1999). This is described in a following section.

The model runs on a time step of 15 minutes so that the daily temperature cycle is captured; mass balance is calculated on a vertical profile at 100 m intervals. The yearly cycle begins at day 300. This ensures that the simulation is started when ablation season has more or less ended. The integration is extended over several years until a stable solution arises. This is because the calculation of albedo and turbulent flux is linked to the ELA, which can only be calculated at the end of one year's integration. It is also to allow transient effects associated with the initial conditions of the albedo profile to dampen out. The meteorological data needed to run the model include the air temperature (°C), precipitation (mm), relative humidity (%) and cloud cover (%). These values are taken for the period 1966 to 1996 from the weather station Loftsalir/Vatnskardshólar, located 7 km from the glacier (T. Jonsson, written comm. 1998). This is the closest weather station and the temperature series lacks any major heterogeneous features, having a high correlation with annual temperature data from the nearby Vik weather station (0.76) and even with the temperature series at Stykkisholmur on the west coast of Iceland (0.66).

Tuning the mass-balance model

The three most sensitive parameters in the model describe the altitudinal dependence of temperature, turbulent flux and precipitation. The temperature lapse rate is an important parameter because many processes in the mass-balance model are temperature dependent (Oerlemans, 1992, Kerr, 1993a). For example, air temperature is used to calculate the altitude of the snow/rain threshold (2°C), the sensible heat flux and the contribution of long-wave radiation emitted by clouds and the clear sky. The free-air temperature lapse rate was extensively studied during the Vatnajökull experiment with over 100 radiosonde balloon soundings, and was found to be low ($-0.0049Km^{-1}$) (Oerlemans *et al.* 1999) compared to typical mid latitude values ($-0.0065Km^{-1}$) (Barry, 1981) due to the extremely maritime environment. This value is used in all the energy balance modelling experiments.

The exchange coefficient for the turbulent fluxes is tuned with the Breidamerkurjökull data. The aim is to achieve a good match with the altitudinal distribution of turbulent exchange on Breidamerkurjökull, but also to keep the range of exchange coefficient (which represents the mean wind speed and surface roughness) within acceptable physical limits (Braithwaite, 1995). The best match is found by making the exchange coefficient ($Wm^{-2}\ K^{-1}$) depend on altitude (m) relative to the ELA (m) following Oerlemans (1992):

$$\text{Exch} = 10 + 0.003(\text{ELA-altitude}) \qquad (2.4)$$

The justification is that surface roughness and mean wind speed tend to increase down-glacier, especially below the ELA. The function used (Equation 2.4) underestimates turbulent exchange at low altitudes (72 Wm^{-2} at 100 m on Sólheimajökull compared with 89 Wm^{-2} on Breidamerkurjökull at 165 m) and slightly overestimate turbulent flux at higher altitudes (25 Wm^{-2} at 700 m on Sólheimajökull compared to 18 Wm^{-2} at 715 m on Breidamerkurjökull) (Figure 2.1). However further modification may lead to 'over-tuning' and unrealistically high values of the exchange coefficient. It is also possible that turbulent exchange on Sólheimajökull at low altitudes might be less important than on Breidamerkurjökull. It has already been shown that katabatic flow over Breidamerkurjökull is persistent and intense (Oerlemans *et al.*, 1999) and it may be possible that turbulent exchange plays a more important role because the fetch of the glacier wind is larger on Breidamerkurjökull than on Sólheimajökull.

The altitudinal distribution of precipitation is tuned last. The aim is to modify the gradient until modelled ELAs fall into the range found for the glacier by remote sensing of late-summer snowline altitudes by Brown (1998). A linear dependence of precipitation on altitude is used:

$$P(h) = PC_0 + C_1 h \qquad \{P=0, P(h)=0\} \qquad (2.5)$$

Where $P(h)$ is the distribution of precipitation with altitude in myr^{-1}, P is the annual precipitation in myr^{-1} and C_0 and C_1 are coefficients. The coefficients determined during the tuning process ($C_0=1.5$, $C_1=0.002$) result in a mean annual precipitation of c. 4 m over Sólheimajökull. This is within the range of values to be expected as glacial runoff measurements indicate 4 to 6 m of annual precipitation (Lawler *et al.*, 1996). It soon became apparent that it would not be possible to achieve an exact match between modelled ELA and the snowline data. This is to be expected, because it is very difficult to compare a simulated ELA with snowline data, for two reasons. There are problems associated with the methodology. For example an error would result if an early snowfall occurred just prior to estimating the height of the snowline. There are also processes involved in the formation of snowlines that are not considered in the model. Brown's (1998) data illustrates this point; there are differences of up to 150 m in snowline altitudes between southern Myrdalsjökull and Sólheimajökull at any one time. This probably reflects local processes such as the redistribution of snow by wind, shading from valley sides and glacier slope effects. In light of the above, the match between modelled and measured data determined by the tuning procedure is good (Table 2.1).

Table 2.1: Comparison between modelled ELA and late summer snowline data for Sólheimajökull and Myrdalsjökull (Brown, 1998).

The good match between modelled ELA and the measured snowline reflects the tuning procedure. The match is better during 1980, 1984, and 1994 than during 1992. The large difference between the snowline on Sólheimajökull and Myrdalsjökull in Brown's data shows that the error bars on the snowline reconstruction are large.

Year	Modelled ELA (m.a.s.l.)	Snowline, Sólheimajökull (m.a.s.l.)	Snowline, southern Myrdalsjökull (m.a.s.l.)
1980	982	960	900
1984	1068	1100	960
1992	868	1080	1020
1994	991	960	960

Testing the mass-balance model

It is important to find an independent way of testing the tuned mass-balance model. Usually this is achieved by comparing model output to mass-balance profiles that have been measured in the field. Because this is not possible at Sólheimajökull, another method has to be found. One way is to assess the model's ability to capture interannual variability by comparing the results with mass-balance measurements from other Icelandic glaciers. Mass balance has been measured on the Hofsjökull ice cap in central Iceland since 1988 by the Natural Energy Authority (Sigurdsson, 1998). One would not necessarily expect that absolute values of mean specific balance at Sólheimajökull and Hofsjökull to be identical during any one year because precipitation and cloudiness are significantly different in southern and central Iceland. However, because temperature in southern and central Iceland is usually well correlated (Einarsson, 1991), we would expect interannual *trends* in mean specific balance to be similar on both glaciers. The results show the match to be better than expected: Measured mean specific balance and ELAs for Hofsjökull are compared to modelled values on Sólheimajökull in Figure 2.3. There is a striking match between modelled net balance on Sólheimajökull and measured net balance on Hofsjökull between 1988 and 1994 (r^2=0.90). The interannual variability appears to be captured. For example, the model predicts the positive mass-balance years in 1989, 1992 and 1993, and the see-saw change to a negative balance year in 1991. There is also an excellent match between modelled ELA on Sólheimajökull and measured ELA on Hofsjökull. As might be expected, the ELA during any one year is lower on Sólheimajökull than on Hofsjökull. This is because Hofsjökull lies inland of Sólheimajökull and there is less precipitation. Overall, the model performs beyond expectations. Although the numbers generated by the model reflect the tuning procedure and may differ slightly from reality, the independent testing suggests that the most important physical processes are included.

The Ice-Flow Model

The ice-flow model is a simplified version of Hubbard (1997). It is a one-dimensional model where time-dependent calculations are made on a centered finite difference array at a fixed interval of 100 m along a glacier flow line (Figure 1.4). Similar models have been used to simulate recent glacier fluctuations in the European Alps and Norway (Huybrechts *et al.*, 1989, Greuell, 1989, Oerlemans, 1997a, Zuo and Oerlemans, 1997, Schmeits and Oerlemans, 1997, Wallinga and van de Wal, 1998). Ice flow is driven by shear stresses, and longitudinal stresses are not included. This simplification is also made in the above mentioned glacier modelling studies and seems justified unless the bed is especially bumpy or if very detailed field data are available to set the boundary conditions of the model. Furthermore model output calculated in sensitivity experiments with an ice flow model for Sólheimajökull including a full stress field show a negligible difference from the output of the model described here (A. Hubbard pers. comm., 1997). This is because the longitudinal bed profile of Sólheimajökull is smooth, at least at the scale of the horizontal resolution of the model (100 m) (Appendix 1). Due to the temperate nature of the glacier, ice is assumed to be warm-based and isothermal. Variability in sub-glacial hydrology is not taken into account because little is known about the sub-glacial drainage system at Sólheimajökull. Instead, sliding at the bed is regulated with an appropriate choice of flow parameters.

Valley parameterisation

The three-dimensional geometry is taken into account by parameterisation of the cross-sectional geometry at each grid point. The glacier width distribution is calculated on the assumption that the valley profile can be approximated with a parabola following Huybrechts *et al.* (1989):

$$B_s = B_{ref} (H/H_{ref})^{1/2} \tag{2.6}$$

Where B_{ref} and H_{ref} are the reference width and height. These values are obtained from topographic maps of the glacier surface (1:50000, Myrdalsjökull DMA sheet 1812) and an ice-radar survey of bed (Appendix 1). The valley beyond the present snout is parameterised using a 1:50000 topographic map and the pattern of Holocene lateral moraines on the valley side (Dugmore, 1987).

Driving stress and ice velocity

The depth-averaged ice velocity is calculated entirely from the local driving stress (τ) a function of surface slope and ice thickness (H):

$$\tau = -\rho g H \, \partial h/\partial x \tag{2.7}$$

A Weertman-type sliding law is used assuming that basal water pressure is a constant fraction of the ice overburden. This leads to the following formulation for the vertical mean ice velocity in a cross section:

$$U = U_d + U_s = F_d H \tau^3 + F_s \tau^3/H \qquad (2.8)$$

Where U_d and U_s are the contributions from ice deformation and sliding, and F_d and F_s are the corresponding flow parameters. The flow parameters determine how ice flow is partitioned between the sliding at the bed and internal deformation within the glacier. In this study, the experimentally derived flow parameters of Budd *et al.* (1979) are used. They result in a relatively larger contribution from sliding when the glacier is thinner, as might be expected. In reality, flow parameters can only be approximately known and vary from glacier to glacier depending on the individual bed conditions, structure of the basal ice layers, debris and water content, and crystal structure of the englacial ice (Paterson, 1994). In a following section, sensitivity experiments are carried out where the flow parameters are modified in order to check that Budd's flow parameters are a good choice for Sólheimajökull.

Calculating ice thickness

The model is based on mass conservation principles, i.e. divergence of the mass flux must be balanced by a change in the cross-sectional area of the glacier at that point because ice is assumed to be incompressible (Paterson, 1994). The three-dimensional geometry of the valley is implicitly accounted for within the continuity equation through the changing width distribution (B_s). This equation can be used to solve for ice thickness (H) at each grid point along the flow line over time (t):

$$\partial H/\partial t = B(h) - 1/B_s\, \partial(UHB_s)/\partial x \qquad (2.9)$$

The mass balance at each grid point B(h) is derived from the mass-balance model. Equations 2.6, 2.7 and 2.8 are substituted into Equation 2.9, and ice thickness is solved over 200 grid points with a forward explicit finite difference scheme. This is found to be stable for small time steps (<0.01 years) where the truncation errors are negligible. The numerical model is coded in Fortran-77 and the model is run on a Sun Workstation.

Defining mass balance in the ice-flow model

The aim of coupling the two models is to allow climatic changes (changes in temperature and precipitation) to be expressed in terms of a change in glacier length, or to allow a change in glacier length to be inverted to estimate a change in climate. The first step involves defining a relationship between net mass balance and altitude that can be read by the model. For example, the mass-balance distribution with altitude at any one time is expressed as:

$$B(h) = B_{ref}(h) + \partial B_t \qquad (2.10)$$

B_{ref} (h) is a reference mass-balance profile. It represents the mean state of mass balance on the glacier and is independent of time. ∂B_t is a mass-balance perturbation which reflects a deviation from mean mass-balance conditions that occur during any one year. The reference mass-balance profile is determined with the aid of the energy-balance model. The model is run for a number of individual years (in this case, using climate data for the period 1966 to 1996). The resultant mass-balance profiles are shown in Figure 2.4. The reference mass-balance profile is then derived by running a second order polynomial regression line through the mean profile for this period. The mean profile for 1966 to 1996 seems reasonable for Iceland (e.g. Björnsson, 1979), with 3 m of yearly accumulation at 1400 m and 9.5 m of yearly ablation at 100 m. The reference mass-balance profile used in this study is:

$$B_{ref}(h) = 0.0127h - 0.00000171h^2 - 11.0 \qquad (2.11)$$

Where h is the height above sea level in metres. This function describes a curve that fits the model output with a squared correlation coefficient (r^2) of 0.998.

The next step is to determine how to best represent a balance perturbation (∂Bt). The simplest method is to assume that perturbations in mass balance are independent of altitude. This means that the shape of the mass-balance profile remains constant in a changing climate, but the entire profile is shifted vertically. In this case, the model has two input functions describing mass balance, the reference mass-balance profile (Equation 2.11) and the ELA, which has the effect of shifting the curve along the y-axis and forcing changes in mass balance over time (∂Bt). This is a reasonable assumption in the ablation zone of a temperate glacier such as Sólheimajökull. For example, Figure 2.4 shows that mass-balance profiles for different years are essentially parallel for the period from 1966 to 1996, despite considerable changes in mean net balance and ELA. This behavior has also been observed on other temperate glaciers and indicates that interannual variability in mass balance in the ablation zone results mainly from temperature changes (Lliboutry, 1974). There is a slightly different pattern in the accumulation zone, where there is less variability (standard deviation = 0.6 m) than in the ablation zone (standard deviation = 1.2 m). This means that shifting the mass-balance profile vertically is likely to result in errors when particularly low ELAs are modelled. For example, in 1989 when the ELA was low (850 m), ablation rates were very low (only c. -6 ma^{-1} at 100 m) but accumulation rates were normal (c. +3 ma^{-1} at 1400 m). If this year is simulated by perturbing the reference mass-balance profile with an ELA of 850 m, ablation is well simulated, but the accumulation rate is overestimated. This problem is overcome by setting an upper limit to the accumulation rate. A nominal figure of 4.5 ma^{-1} is used because this was the maximum modelled accumulation rate between 1966 and 1996.

The final stage is to derive a relation between changes in the ELA and annual temperature and precipitation changes (relative to the 1966 to 1996 period). This is achieved by calculating the mean change in ELA for a +/- 1 K temperature change (St) and a +/- 10% precipitation change (Sp) with the mass-balance model following Oerlemans (1996):

S_t = [ELA (+1K) -ELA (-1K)]/2 (2.12)

S_p = [ELA (+10%) -ELA (-10%)]/20 (2.13)

The above temperature and precipitation perturbations yield S_t=160 $mK^{-1}a^{-1}$, and S_p=2.2m$\%^{-1}a^{-1}$. In other words, a 1 K change in temperature results in a 160 m change in ELA, and a 10 % change in precipitation results in a 22 m change in ELA. These values are in the range to be expected and are similar to the values calculated for Hofsjökull (S_t =110 $mK^{-1}a^{-1}$, S_p=4.0 m$\%^{-1}a^{-1}$) with a degree-day model (Johannesson *et al.* 1995). The ELA can now be expressed in terms of temperature (°C) and precipitation (%):

ΔELA = 160ΔT + 2.2ΔP (2.14)

Equation 2.14 can be used to translate changes in ELA to changes in temperature or precipitation (relative to the 1966 to 1996 mean) on the assumption that one term is equal to zero and that changes in other meteorological variables have no effect.

Testing the ice-flow model

This section describes the 'dynamic calibration' methodology that is used to test the coupled mass-balance ice-flow model (Oerlemans, 1997a). Ice-flow models are often tested by comparing a computed equilibrium ice-surface elevation with a known glacier profile for a given sub-glacial geometry and mass balance profile. This is not a good way to judge the performance of a flow model if the glacier is out of equilibrium during the comparison, and the problem is exaggerated if the mass-balance profile and sub-glacial geometry are not well known. The advantage of dynamic calibration is that transient behavior is taken into account and therefore no assumption about a state of equilibrium is required when model performance is assessed. The procedure involves reconstructing the mass-balance history for the 19th and 20th Centuries so that the changes in glacier length are modelled accurately. Snapshots of modelled ice-surface profiles are then compared with glacier maps. The procedure is repeated with different flow parameters until the root mean square (rms) difference between modelled and measured profiles is minimized. At Sólheimajökull, the sub-glacial geometry is well known and the calibration is concerned with the selection of flow parameters.

Initially, the performance of the model is compared with three sets of flow parameters. The starting point is Budd *et al.* (1979) (Fd = 1.9×10^{-24} $Pa^{-3}s^{-1}$, Fs = 5.7×10^{-20} Pa^{-3} m^2s^{-1}). The simulation is started in 1820 because the glacier snout was known to be stable at 15.2 km in length between 1820 and 1860. An initial ELA is chosen so that the modelled glacier reached a steady state at 15.2 km. Then values of the ELA are determined experimentally in order to simulate glacier fluctuations that match the glacier length record from 1860 to 1996. Seven different ELAs are required and they are imposed for time periods of between 10 and 30 years (Figure 2.5). The experiment is then repeated with different sets of flow parameters (a doubling and halving of Budd's parameters). Figure 2.6 shows the results of the simulations. Budd's initial flow parameters are most successful at reproducing the ice surface profiles in every snapshot, and the mean rms error (20.9 m) is almost exactly half that obtained with the other sets of flow parameters (Table 2.2).

This suggests that the rms error is effectively minimised and that Budd's initial flow parameters are the correct choice. Indeed, the rms error is of similar magnitude to the vertical uncertainty in the measured ice surface profiles, which is about contour level on the maps (20 m) and 15 m in the 1996 GPS survey of the ice surface (Appendix 1). This excellent result indicates that the model is appropriate for simulating the transient fluctuations of the glacier.

Table 2.2: Model performance with different flow parameters.

Budd' refers to the initial flow parameters suggested by Budd *et al.* (1979). 'Budd*2' and 'Budd/2' refer to a doubling and halving of the initial flow parameters. The table shows the rms errors between ice-surface profiles derived from the model and glacier maps during four different snapshots in time (in metres). The average rms error is also shown for each set of flow parameters.

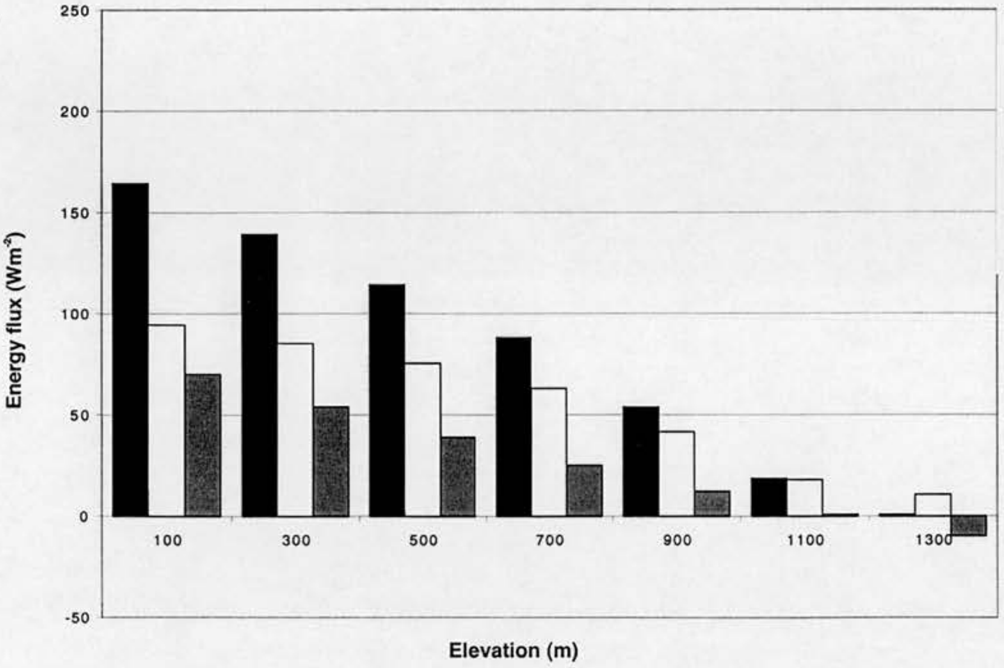
	Budd	**Budd*2**	**Budd/2**
average rms	20.9	41.5	41.6
rms 1904	17.5	30.4	34.1
rms 1948	14.4	46.2	35.3
rms 1970	28.1	30.1	50.7
rms 1996	23.7	59.3	46.4

Figure 2.7 shows the simulated 1996 ice-surface profile, and the driving stress and velocity distribution generated using Budd's initial flow parameters. It should be emphasized that the glacier is not in a steady state. The driving stress increases rapidly from a value of zero near the ice divide (where longitudinal stresses would be important in reality) to over 2 KPa at 4.5 km. This value is maintained between 4.5 and 10 km, after which the driving stress drops of slowly at first to about 1.5 KPa and then rapidly to zero at the snout. Deformation dominates over sliding between 7.5 km and 13 km, where the glacier flows through a deep trough. Sliding is important when ice thickness is smaller, around the breach in the caldera (3.5 to 6 km) and towards the snout (beyond 13 km). Total velocity peaks at 130 ma^{-1}, and is generally greater than 100 ma^{-1} between 3.5 km and 10 km, where driving stress is high. Simulated velocity between 7.5 and 12.5 km (mean =114 ma^{-1}) compares well to a mean empirical estimate derived from tracing the movement of tephra layers on the ice surface with aerial photographs over the last 50 years (120 +/- 25 ma^{-1}). Although both the modelled and measured ice surface velocities contain uncertainties, the match is encouraging and suggests that the glacier dynamics are appropriately simulated with the model.

Summary

A coupled mass-balance glacier-flow model has been constructed for Sólheimajökull. When tested against the independent mass-balance record at Hofsjökull, the model is shown to simulate interannual variations in net mass balance well. The flow model is tested in a time-dependent simulation from 1820 to 1996. The model successfully calculates ice-surface profiles for four snapshots in time. The success of the time-dependent simulation means that we can use the models to perform experiments in order to test the hypotheses presented in Chapter 1. We now have the tools to assess the response of the glacier to climatic change. The good fit suggests that model can also be used to invert a climate signal or to make projections of future glacier length for imposed climatic changes.

a



b

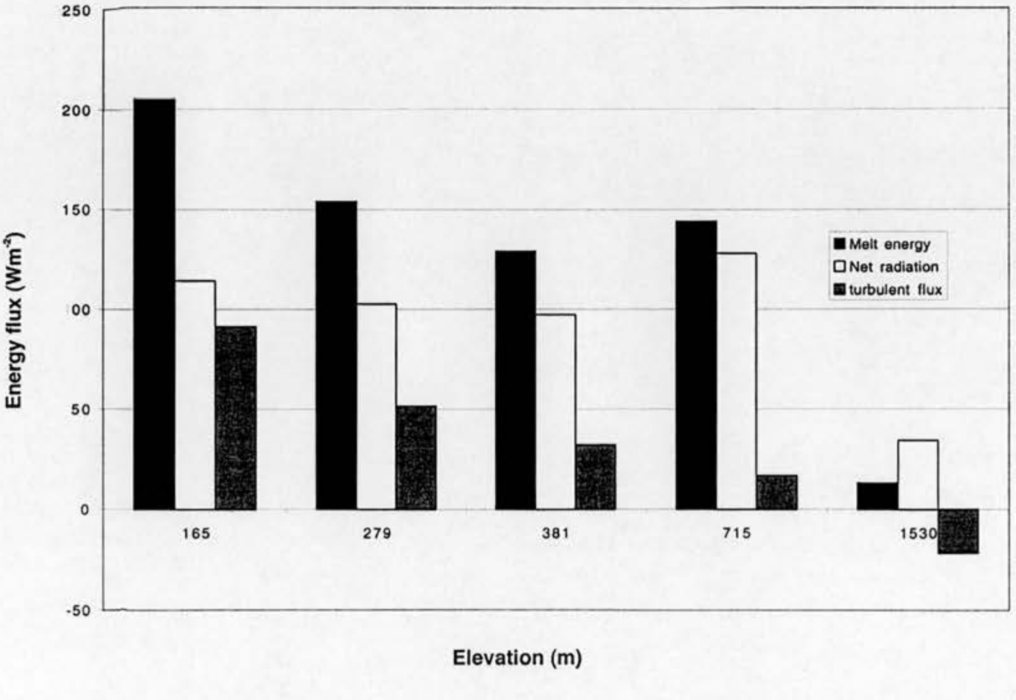


Figure 2.1

Modelled summer energy fluxes on Sólheimajökull (a) for the same 100 day period when energy fluxes were measured on Breidamerkujökull (b). The measurements are at slightly different altitudes. The main difference between the modelled and measured energy fluxes is the large amount of melt energy available at 715 m on Breidamerkujökull, which results from a lower surface albedo.

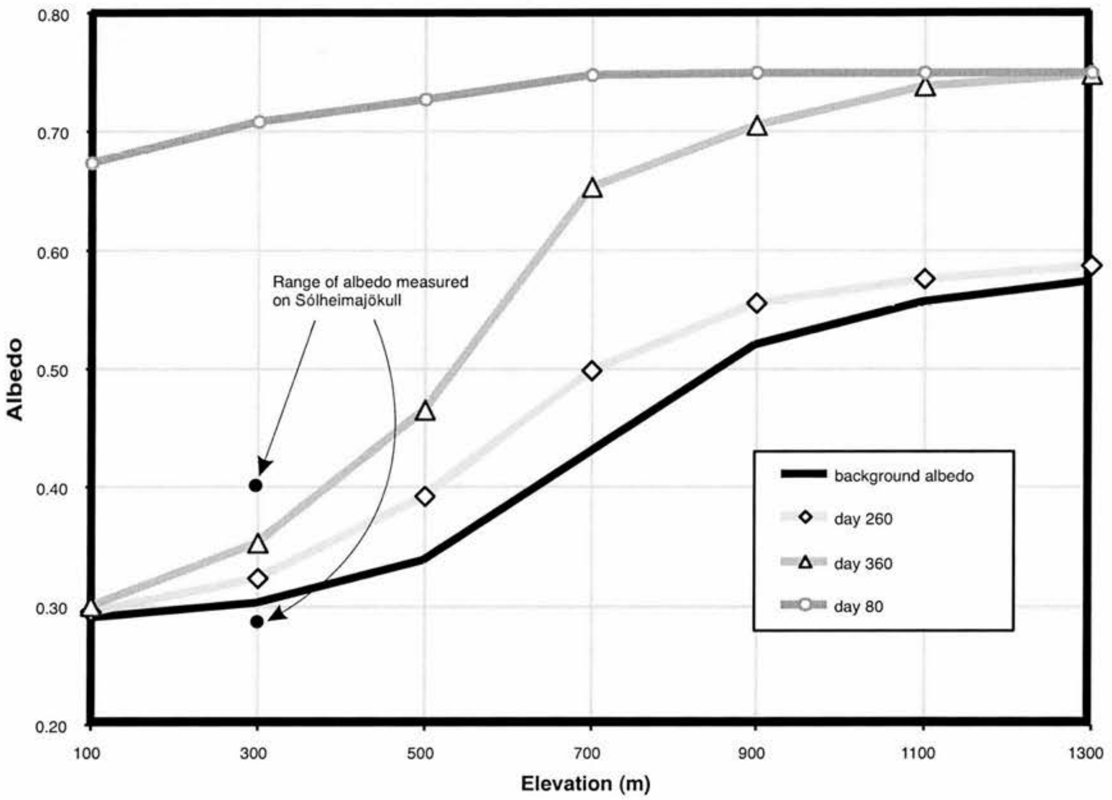


Figure 2.2

Simulated albedo profiles on Sólheimajökull for a mass balance year. The simulated profiles on the ablation zone fall inside the range measured by Brock (1996).

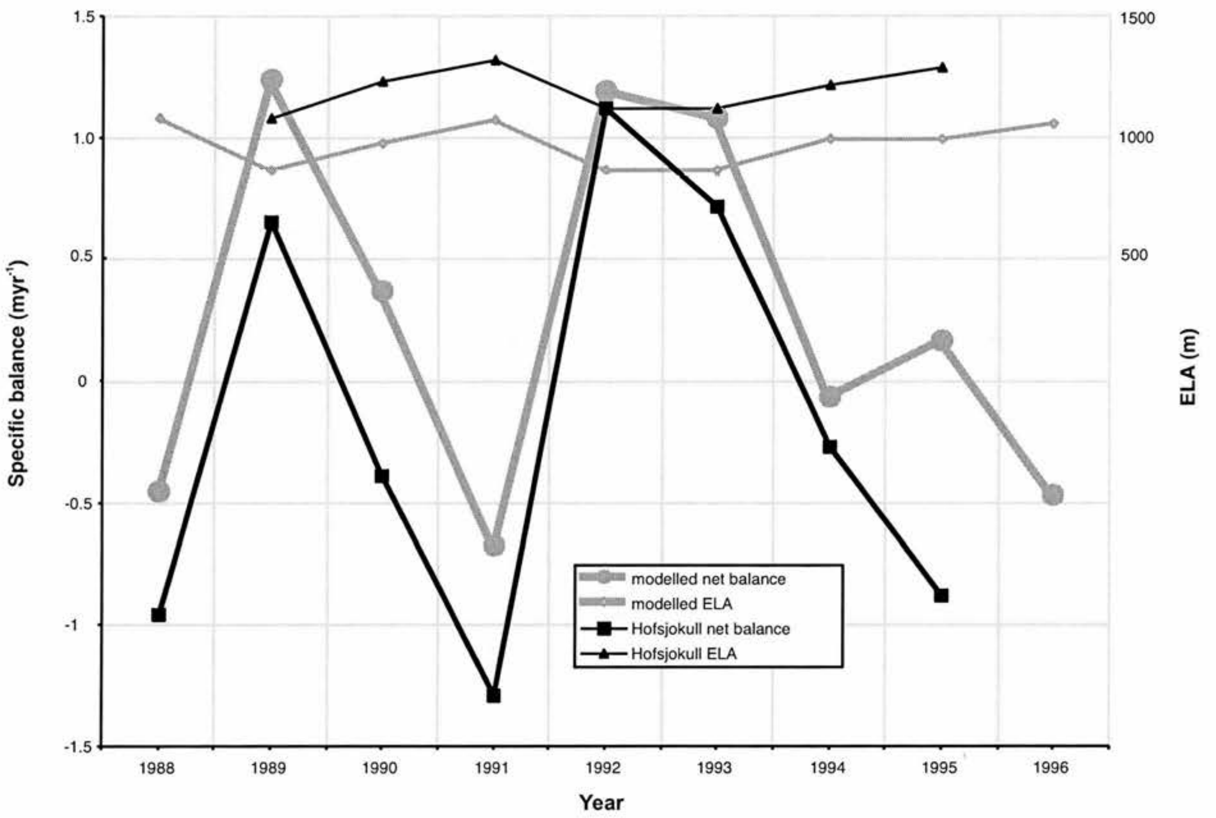


Figure 2.3

A comparison between simulated ELA and specific balance on Sólheimajökull and measured ELA and specific balance on Hofsjökull from 1988-1996. The interannual trends in mass balance appear to be well captured by the model.

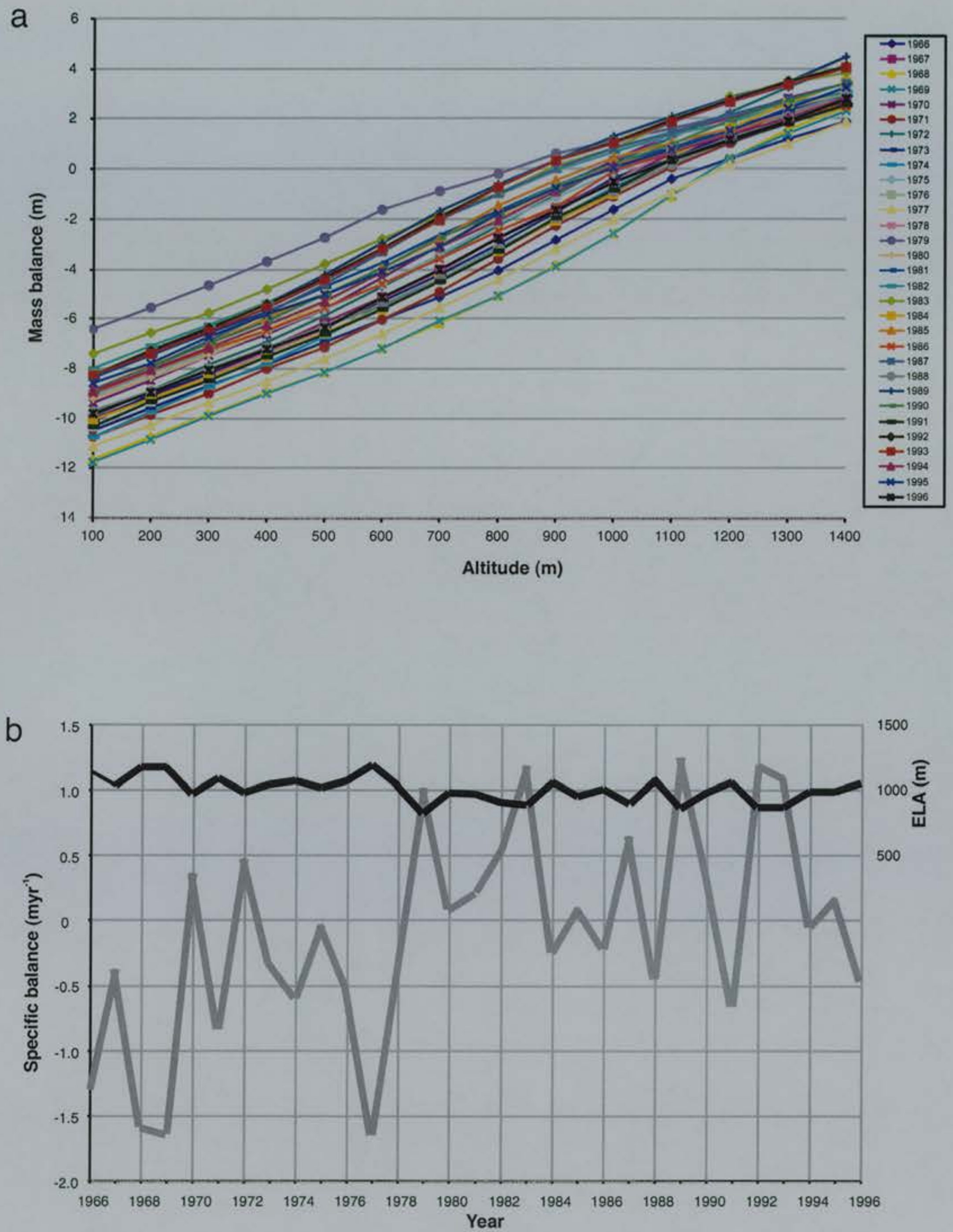


Figure 2.4

a. Modelled mass-balance profiles from 1966-1996 for Sólheimajökull. The gradients are essentially parallel in the ablation zone

b. Modelled specific balance and ELA for the same period. Note the large interannual variability in specific balance.

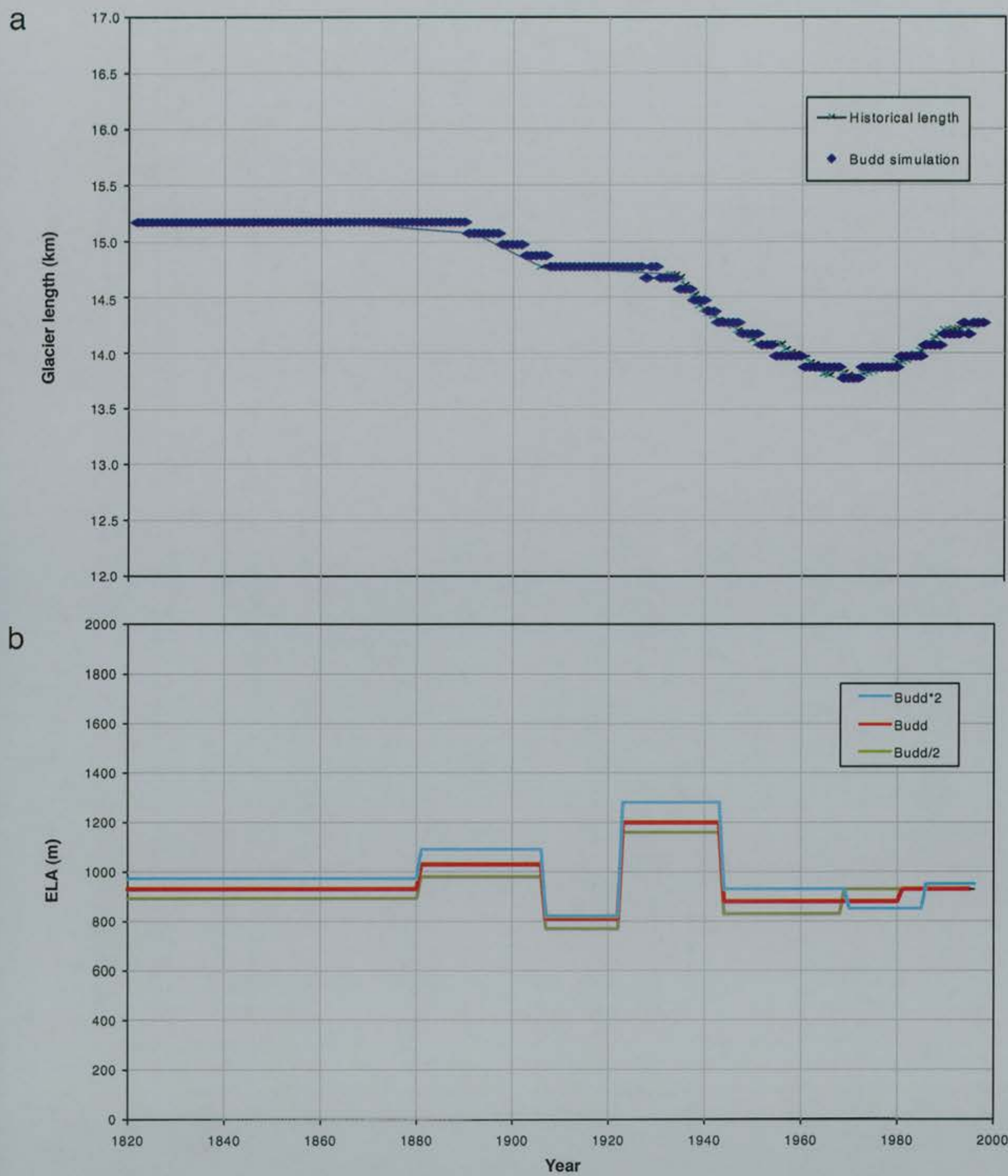


Figure 2.5

a. Simulated glacier length and measured glacier length fluctuations. The simulation started at 1820 at an equilibrium length of 15.2 km.

b. The simulation was forced by shifting the ELA vertically. The magnitude of change differs slightly for the different sets of flow parameters but the pattern is similar.

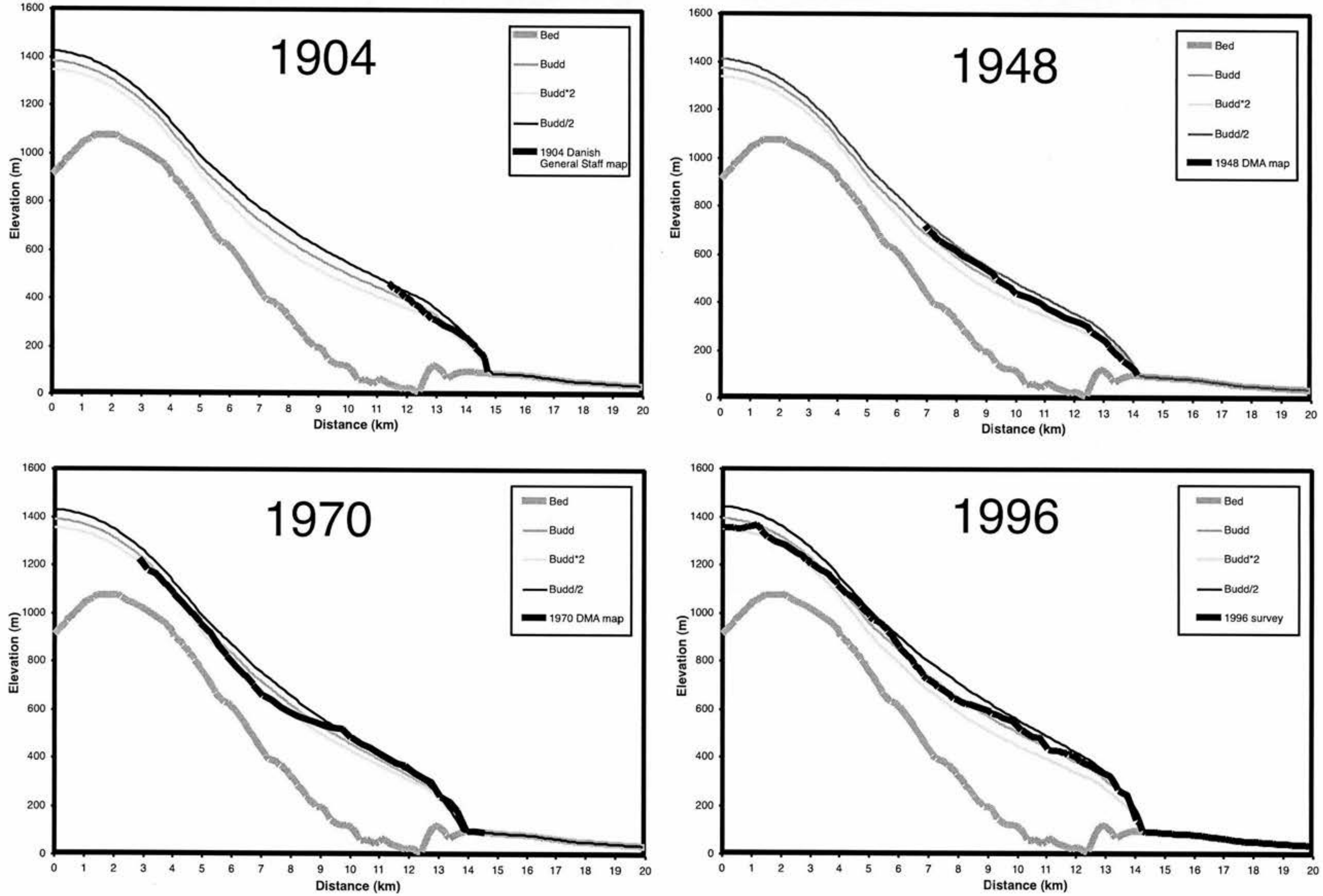


Figure 2.6

Comparisons between modelled ice surface profiles and measured glacier profiles for four snapshots in time during the 20th Century. Budd's initial flow parameters perform best in each case.

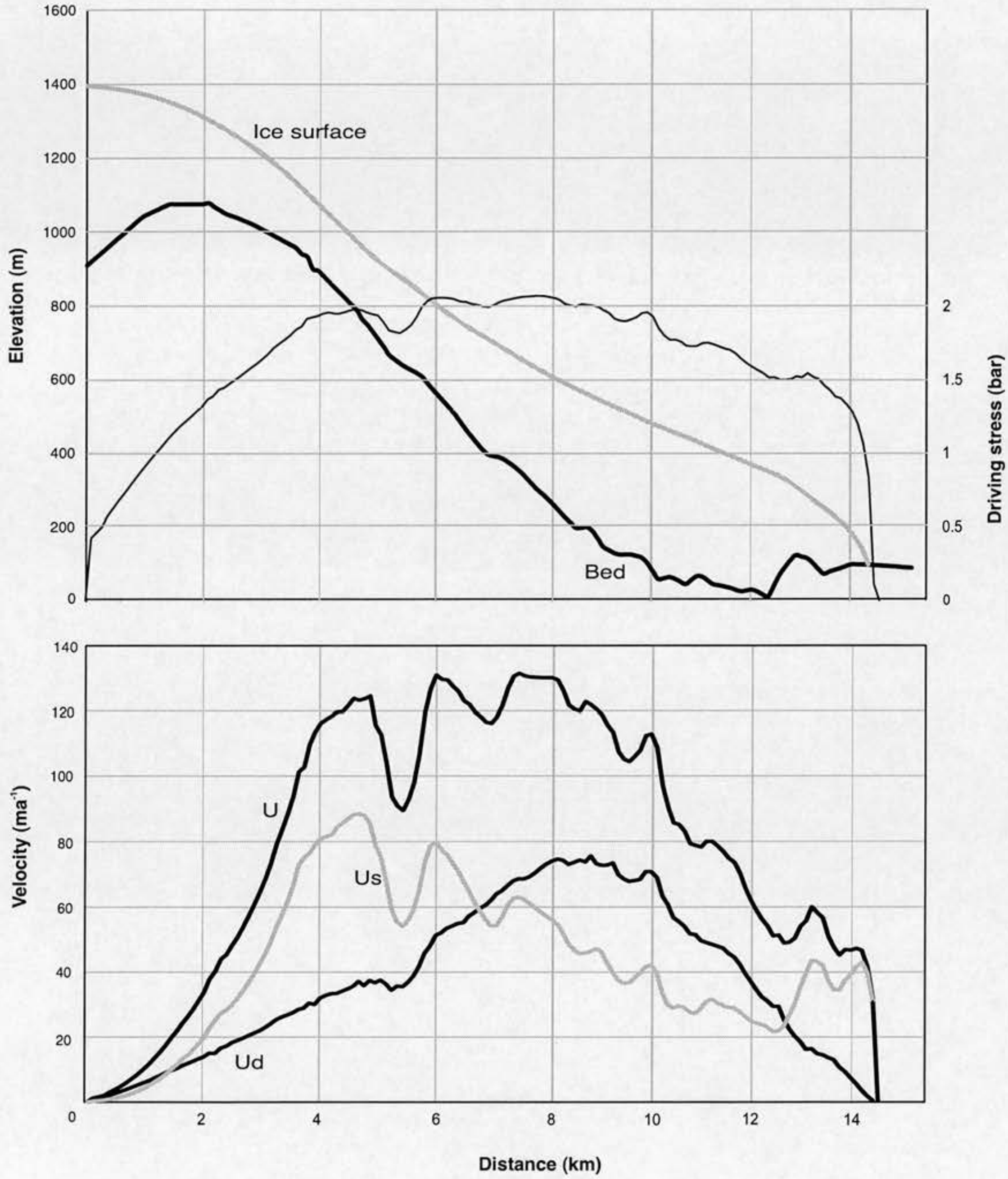


Figure 2.7

Modelled glacier dynamics for 1996. The ice velocity (U) peaks between 4 and 10 km between the ELA (at c. 4 km) and lower parts of the glacier where flow remains well channeled. The sliding velocity (U_s) is largest when ice thickness is small where the glacier crosses the caldera rim (from 4-6 km), while the deformation velocity (U_d) is highest where the glacier reaches its maximum thickness (at 8-9 km). The ratio between sliding and deformation appears to be realistic, but no data is available to critically test this result.

Chapter 3: Modelling the Response of Sólheimajökull to Climatic Change

Introduction

The aim of this chapter is to use the models developed in Chapter 2 to assess the response of Sólheimajökull to climatic change. Little quantitative information is available on how Icelandic glaciers respond to climatic change, and modelling studies have been restricted to Hofsjökull (Johannesson, 1991, Johannesson, 1997). Glacier mass balance is believed to be very sensitive to climatic changes in high precipitation areas in temperate regions (Oerlemans, 1992). Glaciers on the south coast of Iceland are located in a region where storm activity is intense and annual precipitation is greater than 4 m. As a result, they have steep mass-balance gradients, large ice fluxes and many extend to near sea level (Björnsson, 1979). It is expected that the mass balance of Sólheimajökull is very sensitive to climatic change. One aim of this chapter is calculate the mass-balance sensitivity for Sólheimajökull. The model is used to isolate the meteorological controls on the mass-balance sensitivity by determining the relative importance of temperature and precipitation. It is also used to determine how glacier mass balance responds to changes in temperature and precipitation during different seasons.

The second aim of this chapter is to assess how bedrock geometry influences the response of longer-term terminus fluctuations of Sólheimajökull. The response of a glacier snout to a climatic change occurs in several steps involving feedback loops between climate, glacier dynamics and topography (Meier, 1965, Furbish and Andrews, 1984, Oerlemans, 1989). These processes can dampen or enhance the response of a glacier to climatic change. There is reason to believe that Sólheimajökull has topographic characteristics that promote large amplitude changes in glacier length in response to a climatic change; a wide flat accumulation area, a long narrow snout and a gently inclined bed (Mackintosh *et al.,* 1999). In the second part of the chapter this hypothesis is tested in a series of experiments that are designed to isolate the influence of valley geometry on the glacier's response to climatic change.

Sensitivity Experiments with the Mass-Balance Model

The aim of this section is to explore the relationship between climatic changes and mass balance at Sólheimajökull. In the last chapter, mass-balance profiles were reconstructed for the period from 1966 to 1996 in order to define a reference mass-balance profile. The reconstruction shows that there is greater interannual variability in ablation than in accumulation (Figure 2.4). This implies a high sensitivity to temperature change (the first order influence on ablation variability) and a lower sensitivity to precipitation change (the first order influence on accumulation variability). This inference does not of course take glacier hypsometry into account and there is a possibility that precipitation might be more important than one might expect because the glacier has a large proportion of its area in the accumulation zone (Figure 1.5). The most realistic way of calculating the sensitivity of the mass balance to changes in temperature and precipitation is to calculate the change in mass balance averaged over the glacier area for a uniform temperature or precipitation perturbation. Following Oerlemans (1996) the mass-balance sensitivity with regard to changes in temperature and precipitation (Ct, Cp) are defined as the change in mean specific balance for a uniform change in temperature or precipitation:

$$C_t = \partial B/\partial T = [B\,(+1K) - B(-1K)]/2 \qquad (3.1)$$

$$C_p = \partial B/\partial P = [B\,(+10\%) - B(-10\%)]/20 \qquad (3.2)$$

The mass-balance sensitivity to temperature and precipitation change is calculated with the energy balance model. C_t is calculated as 1.4 $ma^{-1}\,K^{-1}$ and C_p at 0.025 $ma^{-1}\,\%^{-1}$. This means that a 70% increase in annual precipitation is required to balance a 1°C rise in temperature. This again confirms that in temperate regions glacier mass balance is very sensitive to temperature change and somewhat less sensitive to precipitation change. One way to put these findings into a realistic context is multiply the values of the mass-balance sensitivity by the standard deviations of an annual temperature and precipitation series from an Icelandic climate station. This is done for the standard deviation of temperature from the series 1890 to 1990 at Stykkisholmur (σ=0.72°C) and the precipitation series from Vik over the same time period (σ=0.21 m or 14.3 % of the 1890 to 1990 mean). The results show that the mass-balance is indeed more sensitive to temperature changes; the variation in mean specific balance resulting from one standard deviation of temperature change from 1890 to 1990 is 1.0 ma^{-1}. Mass-balance variation resulting from one standard deviation of precipitation change over the same time period result in a much smaller mass-balance variation of 0.36 ma^{-1}. This indicates that during the 20th Century, temperature changes were probably responsible for the fluctuations of the glacier. It also suggests that large precipitation changes would be required to slow or halt a glacier retreat resulting from a future climate warming.

The next set of modelling experiments is designed to assess how climatic changes in different seasons influences the mass-balance profile. In temperate regions it is often assumed that the glacier mass balance is most sensitive to changes in summer temperature and winter precipitation. Yet significant changes in temperature or precipitation occur during other seasons, and it is important to know whether or not these changes would influence mass balance. For example, in Chapter 6 we will see that when sea ice surrounds the Icelandic coastline, it has the largest effect on spring and winter temperature. Furthermore Johannesson *et al.* (1995) have predicted that future changes in temperature associated with enhanced global warming in Nordic countries will be larger in the winter months.

Seasonal changes are investigated in a series of experiments in which spring, summer, autumn and winter temperature and precipitation are systematically varied and the changes in mass balance with altitude calculated (Figure 3.1). Changes in summer temperature are most important as expected, especially at lower altitudes where melting rates are large and most of the changes in mass balance occur. However the change in mass balance for perturbations in spring, autumn and even winter temperatures is also significant. In comparison, temperature changes on the highest part of the glacier during winter have little effect, because air temperatures rarely rise above the freezing point. The mass-balance profile undergoes large changes in response to perturbations in winter precipitation as might be expected, but again, changes in spring and autumn precipitation are also significant. On the other hand, changes in summer precipitation have a negligible effect on mass balance. The experiments also reveal the important role of the surface albedo. During summer, the largest changes in mass balance occur below the ELA, where the exposed ice has a low albedo. During spring and autumn, the change in mass balance is relatively similar at all altitudes. This is because a well-distributed snow cover dominates the albedo. Another key factor is the altitude of the snow/rain threshold, which approximates the freezing level. In summer the sensitivity to precipitation change approaches zero below c. 1000 m, (the lower limit at which snow will fall), while in winter, changes in mass balance are still significant at 100 m because snowfall is common down to sea level.

Sensitivity Experiments with the Ice-Flow Model

The interactions between the glacier and its underlying topography are important because they can modify glacier response to climatic change. The aim of this section is to explore the relationship between climatic changes, bedrock topography and glacier dynamics at Sólheimajökull. The time scale and magnitude of glacier response to climatic change are investigated in a series of experiments. First, several concepts are defined mathematically.

Equilibrium changes in glacier length and volume: concepts

For a given climatic change, the magnitude of response in terms of glacier extent and volume is known to vary from glacier to glacier, depending on the topography of the sub-glacial trough (Oerlemans, 1994). One way to quantify the different response between glaciers is to calculate the derivative of glacier length ($\partial L/\partial T$) and volume ($\partial V/\partial T$) with regard to temperature. This can be done with the glacier flow model using a simple experiment; a glacier is in an equilibrium state with a length of L_1 and volume of V_1 corresponding to a climate state described by the reference mass-balance profile (Equation 2.11) and an ELA_1. The ELA is shifted in a step from ELA_1 to ELA_2, and the model is allowed to run forward in time until the glacier reaches a new equilibrium length L_2 and volume V_2. The temperature change corresponding to ELA_1-ELA_2 can be calculated from Equation 2.14 assuming that precipitation remains constant. Thus, the derivatives ($\partial L/\partial T$ and $\partial V/\partial T$) can be calculated:

$$\partial L/\partial T \cong 160(L_2-L_1) / (ELA_2-ELA_1) \tag{3.3}$$

$$\partial V/\partial T \cong 160(V_2-V_1) / (ELA_2-ELA_1) \tag{3.4}$$

Response time: concepts

The response time is broadly defined as the time taken for a glacier to complete its adjustment to a climatic change from one steady state length or volume to another. Three different response times are defined following Oerlemans (1996):

Growth time

V_i is the equilibrium volume of a glacier that is in balance with the prevailing constant climatic state C_i. The growth time is the time a glacier needs to attain a volume of $(1\text{-}e^{-1})\ V_i$ starting from zero ice volume in the climatic state C_i.

Volume response time

The climatic state is changed stepwise from C_1 to C_2. The corresponding equilibrium glacier volumes are V_1 and V_2. The volume response time is the time a glacier needs to attain a volume $V_{(t)}$ of:

$$V_{(t)} = V_1 + (1\text{-}e^{-1})\,(V_2\text{-}V_1) \cong V_o + 0.632(V_2\text{-}V_1) \tag{3.5}$$

Length response time

The climatic state is changed stepwise from C_1 to C_2. The corresponding equilibrium glacier lengths are L_1 and L_2. The length response time is the time a glacier needs to attain a length $L_{(t)}$ of:

$$L_{(t)} = L_1 + (1\text{-}e^{-1})\,(L_2\text{-}\ L_1) \cong L_o + 0.632(L_2\text{-}\ L_1) \tag{3.6}$$

The relationship between topography, glacier extent, volume and response time.

The first experiment is designed to investigate the relationship between valley geometry, $\partial L/\partial T$, $\partial V/\partial T$ and the length and volume response times. The ELA is lowered in uniform steps of 25 m, from an initial ELA of 1075 m to final ELA of 700 m. This corresponded to 15 equilibrium glacier profiles ranging from zero ice volume to a glacier extending to the trough limits (Figure 3.2). After each step ELA lowering, the ice flow model is allowed to run forward in time until the glacier reaches a new equilibrium position. The derivatives $\partial L/\partial T$ and $\partial V/\partial T$, and the length/volume response times are then calculated. This allows the dependence of glacier response on topography to be assessed.

In Figure 3.3, $\partial L/\partial T$, $\partial V/\partial T$, and the length and volume response times are plotted against glacier length. The variation of valley width and bed altitude with glacier length is also plotted, so that the dependence of the above-mentioned properties on the valley geometry can be assessed. The dependence of $\partial L/\partial T$ on topography is examined first. In summary:

- $\partial L/\partial T$ is largest when the glacier advances between 11.5 and 14.0 km; in this section of the valley, an advance of up to 900 m occurs for each 25 m lowering of the ELA.
- At 15 km, $\partial L/\partial T$ is smaller; an advance of only 400 m occurs for a 25 m lowering of ELA. Beyond a glacier length of 15 km, $\partial L/\partial T$ increases again; an advance of c. 0.8 km occurs for each 25 m lowering of ELA.

- After 18 km $\partial L/\partial T$ decreases at its minimum value, an advance of only 0.2 km occurs for every 25 m lowering of ELA.

$\partial L/\partial T$ appears to depend on bed topography (Figure 3.3); a local minimum in $\partial L/\partial T$ coincides with an increase in valley width (Threshold 1), and the rapid decrease in $\partial L/\partial T$ beyond a glacier length of 18 km occurs when the glacier spreads laterally as it advances out of the valley, and onto the sandur plain (Threshold 2). Changes in bed altitude seem to have a smaller influence on $\partial L/\partial T$; the hill at a glacier length of 13 km seems to cause only a minor decrease in $\partial L/\partial T$.

We now turn to the relationship between changes in glacier volume and topography. It appears that $\partial V/\partial T$ is less influenced by topography than $\partial L/\partial T$. In Figure 3.3 it is evident that $\partial V/\partial T$ increases as the glacier becomes larger. This is because the glacier thickens as it advances over a gentle bed slope. A threshold is again crossed beyond a critical glacier length of 18 km (Threshold 2) and $\partial V/\partial T$ decreases again. This is because the glacier spreads laterally as it advances onto the sandur plain. This is interesting because it indicates that the sandur plain forms a barrier to glacier advance in terms of both length and volume.

The relationship between glacier response time and topography is now considered. In Figure 3.3 it can be seen that length and volume response times vary between 20 and 90 years. The response times for a contemporary glacier length of c. 14.3 km vary between 60 and 70 years, although it must be kept in mind that the response time also depends on the size of the mass-balance perturbation (Wallinga and van de Wal, 1998). For example, a step change in ELA of 50 m at a glacier length of c. 14 km results in a shorter response time of c. 40 years. As expected, the length response time is mostly larger than the volume response time (Oerlemans, 1996). However, beyond a glacier length of 18 km (Threshold 2), both the length and volume response times decrease and the volume response time becomes larger than the length response time. This is probably an artifact of the coarse horizontal resolution of the model where the snout halts at a grid point but in reality, small changes in glacier length would continue to occur until the glacier volume approached equilibrium with the new climatic state. In other words, if a glacier is located at a point of valley widening, a fall in ELA causes the glacier to undergo a very small change in length. After the advance nears completion, glacier volume continues to increase for a longer period as the glacier continues to thicken without advancing much further.

Growth time

In a second experiment, the growth time is calculated for three different steady state ice volumes corresponding to three climatic states with ELAs of 1070 m, 968 m, and 700 m. These ELA values are chosen because they represent the steady state lengths of three characteristics; the first glacier to form after the ELA intersects with the caldera rim (Figure 3.2), a glacier with the 1996 length of 14.3 km, and a glacier that extends to the end of the valley. The experiment reveals that the growth time is strongly dependent on the size of the ELA lowering. For example, the growth time for an ELA of 700 m is 92 years, 229 years for an ELA of 968 m and 741 years for an ELA of 1070 m (Figure 3.2). This is an interesting result, showing that glacier initiation can be rapid in a maritime environment if the climatic conditions are favourable. On the other hand, it takes a much longer for a small glacier to grow from zero ice volume. The rate of growth for an ELA of 1070 m is exceedingly slow and non linear; the glacier grows slowly from 2 to 5 km (700 years) and then rapidly to 10 km (100 years). Rapid growth is only initiated once the glacier moves off the plateau and ice thickness increases over a larger area. This indicates that the elevation mass balance feedback (Oerlemans, 1980) is important in initiating glacier growth at Sólheimajökull. If the glacier gains enough altitude initially, then a strong positive feedback results and the glacier grows quickly.

Discussion

The calculated mass-balance sensitivity of Sólheimajökull to temperature change (1.4 $ma^{-1}K^{-1}$) is large on a global scale. For example, it is 1.5 times greater than the computed sensitivity of Nigardsbreen in Norway to temperature change (0.9 ma^{-1} K^{-1}), which is amongst the most sensitive in a study of 12 glaciers worldwide (Oerlemans *et al.*, 1998). Presumably the even greater mass-balance sensitivity of Sólheimajökull reflects the extreme maritime influence and the location of Iceland in the path of the North Atlantic storm tracks. On the other hand, the mass-balance sensitivity of Sólheimajökull to precipitation changes is lower than at Nigardsbreen (0.035$ma^{-1}\%^{-1}$). This gives support to the notion that glaciers in temperate regions are more sensitive to temperature changes in high precipitation areas but less sensitive to precipitation changes (Kerr and Sugden, 1994).

The experiments relating altitudinal changes in mass balance on Sólheimajökull to seasonal changes in temperature and precipitation show that mass balance responds to changes in temperature and precipitation all year round, rather than just to summer temperature and winter precipitation changes. The mass-balance is especially sensitive to changes in spring and autumn temperature. Again, this reflects the location of Sólheimajökull in a maritime environment. Temperatures are mild for this latitude due to the moderating effect of the nearby ocean. Furthermore, the glacier has a favourable hypsometry that allows it to descend to low elevations. These two factors mean that ablation occurs year round and annual ablation rates are large. The mass-balance profile of Sólheimajökull is much more sensitive to spring and autumn temperature changes than the mass balance of the Unterer Grindelwaldgletscher in the European Alps where a similar study has been undertaken (Schmeits and Oerlemans, 1997).

The experiments with the ice-flow model indicate that fluctuations of Sólheimajökull are affected by threshold behavior induced by the topography of the glacier trough. Glacial inception only occurs after the ELA intersects the caldera rim, and at this point the glacier length is controlled by a positive feedback loop; as the glacier thickens, it increases in altitude. As a result, the area of the accumulation zone increases and more accumulation is added.

The derivative $\partial L/\partial T$, representing the scale of the glacier length response to climatic change, does not remain constant as the glacier advances through the valley. In general $\partial L/\partial T$ is high, especially between a glacier length of 10 and 15 km and between 16 and 18 km. However, $\partial L/\partial T$ becomes smaller when the glacier has a length of 15 km and when the glacier extends beyond a length of 18 km. The decrease of $\partial L/\partial T$ is related to the geometry of the bedrock topography. In Figure, 3.3, two topographic thresholds are identified. The topographic thresholds are superficially similar to 'topographic pinning points' which are known from studies of fjord glaciers (e.g. Mercer, 1961, Warren, 1992) where glaciers stabilize at points of valley widening or at changes in the bed slope. In both cases in this study, the reduction of sensitivity can be ascribed to trough widening. The topographic threshold is located at a glacier length of 15 km at a point of valley widening. This is just beyond the present glacier snout where many Little Ice Age moraines are found (Figure 1.5, Figure 1.7). The second topographic threshold is located at the end of the valley at a glacier length of 18 km, where unconfined spreading of the glacier over its sandur plain causes the ablation zone to increase in area, rendering the ice unsustainable. $\partial L/\partial T$ becomes smaller by a factor of 3 to 4 beyond this critical glacier length. This means that a glacier terminating at 18 km will only advance 2 km if the ELA is lowered by 250 m (equivalent to a 1.6°C lowering of temperature) yet a glacier terminating at 10 km will advance by up to 6 km for the same change in ELA.

Topographic thresholds identified in this study are distinct from topographic pinning points (fjord glaciers) because they do not involve a hysterisis, as is witnessed in the case of some fjord glaciers. In the case of a hysterisis, an unrecoverable change in glacier length can occur if a glacier becomes decoupled from a pinning point, and multiple steady state positions are possible for the same climatic condition (Warren, 1992). In the case of the topographic threshold identified in this study, the threshold simply induces a transient change in behavior while the glacier remains at a length corresponding to the position of the threshold.

The time scale of the glacier response to a climatic perturbation is influenced by topography. A substantial glacier could grow from ice-free conditions between c. 100 and 800 years, depending on the prevailing climate. This large range of growth times results from a non-linearity induced by the topography of the glacial trough. The long growth time relates to a glacier forming under marginal climatic conditions, where ice build up is slow until the glacier flows off the plateau surface. At this point, fast growth results from a positive feedback where an ice thickness increase leads to an increase in glacier altitude over a large area.

The volume and length response time of the glacier can also be related to topography. The response time increases in areas where the ice thickness increases and the glacier has a small ice surface slope. This behavior is apparent for the larger glacier profiles in Figure 3.2, where the glacier has a length greater than 15 km (Figure 3.3). This is to be expected because the response time of a glacier is known to be directly proportional to the ice thickness and inversely proportional to the balance gradient at the snout (Johannesson *et al.,* 1989). The response time analysis reveals another interesting relationship. The length response time tends to be larger than volume response time when $\partial L/\partial T$ is large, and the reverse is true when $\partial L/\partial T$ is small. This is evident at the topographic threshold at a glacier length of 18 km, indicating that the glacier must continue to thicken after a certain point without a significant advance of the snout.

It would be interesting to repeat this experiment with a more sophisticated ice-flow model to test the robustness of this finding. A two dimensional model including longitudinal deviatoric stresses would improve the simulation of the glacier once it expanded onto the sandur plain, since back-stresses would become more important and the one-dimensional valley parameterisation used in this study becomes less valid.

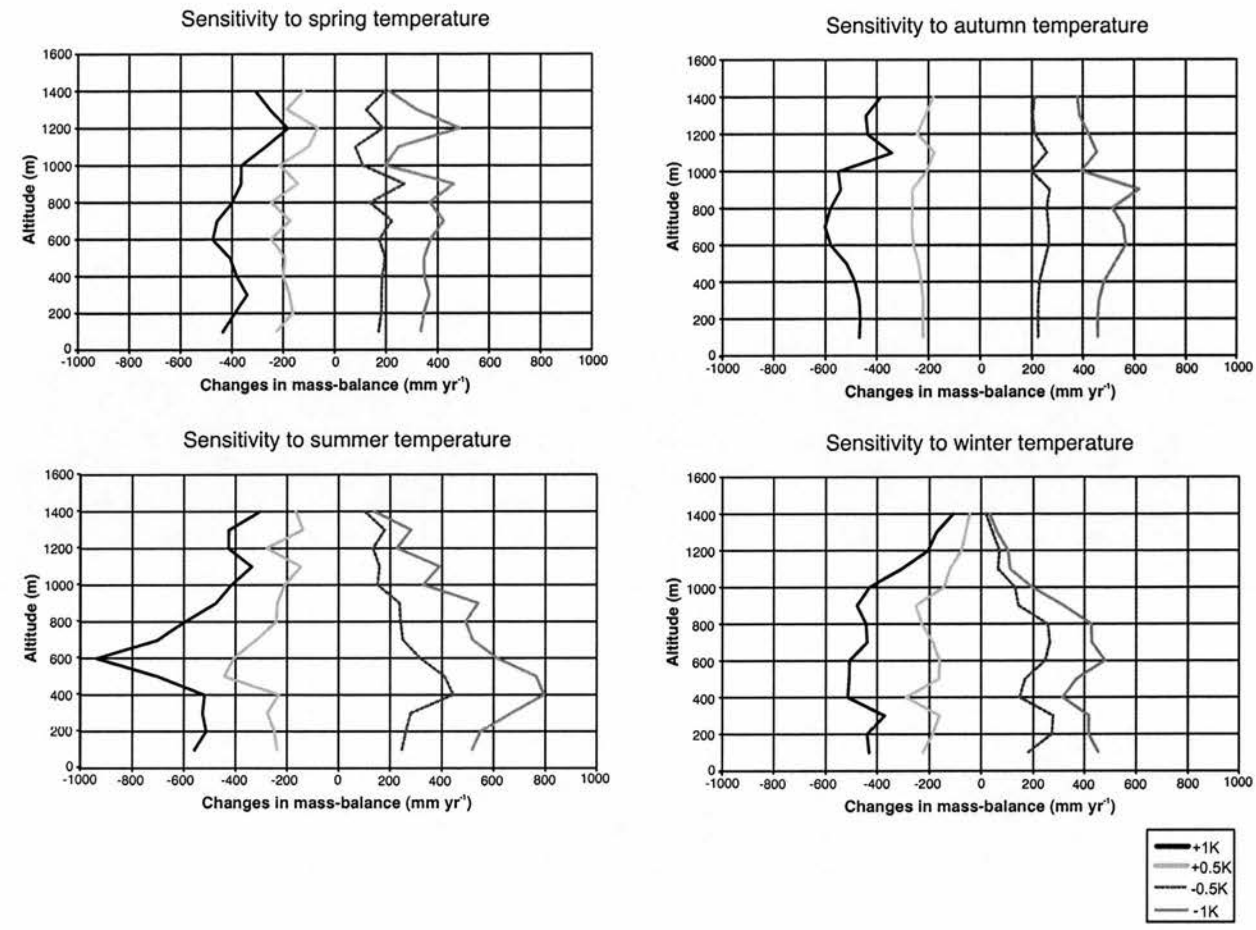
The finding that topography influences the response of Sólheimajökull to climatic change may help in understanding the pattern of past glacier fluctuations. For example, the Little Ice Age fluctuations of Sólheimajökull were small in comparison with other Icelandic glaciers (Dugmore, 1989). Figure 1.6 shows that the glacier spent much of the 19^{th} Century at a length of 15 km. Topography may have been partly responsible for the small scale of the Little Ice Age glacier advance at Sólheimajökull.

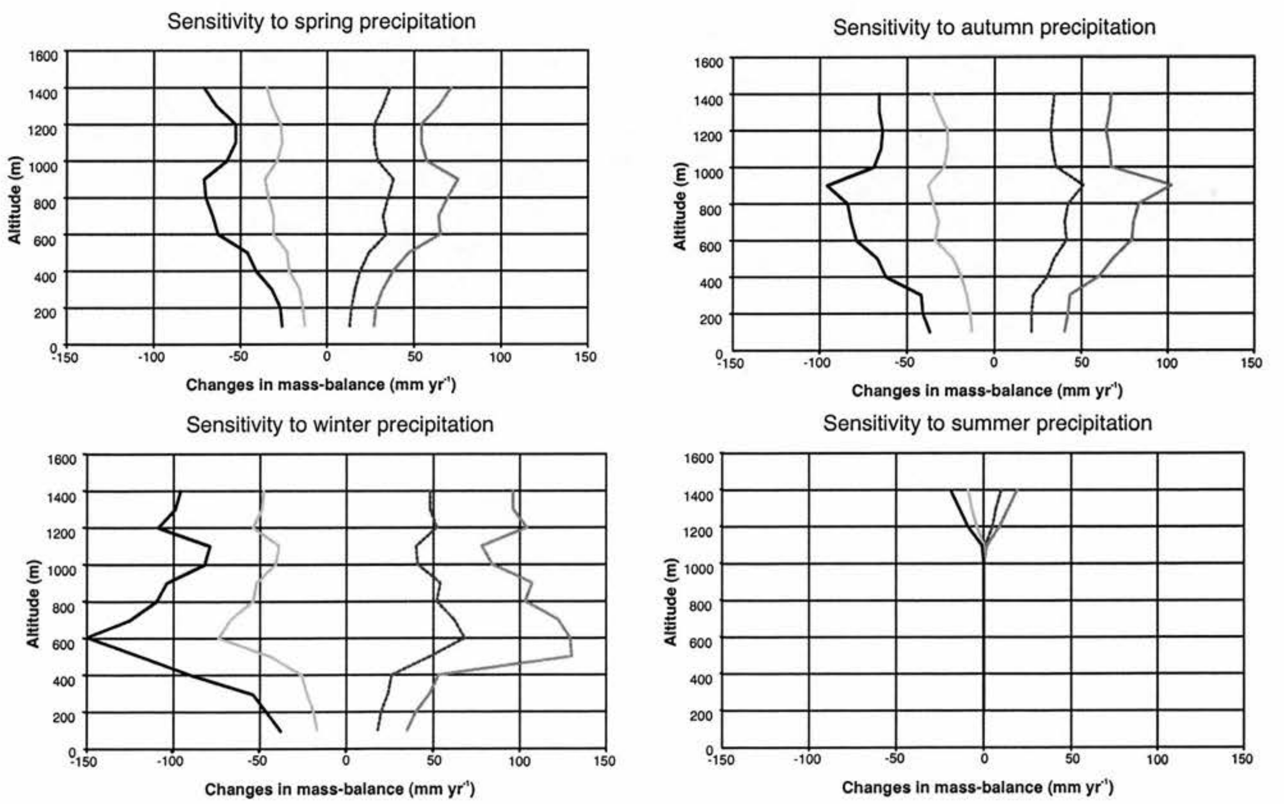
Figure 3.3 demonstrates that changes in valley width have the largest effect on the local sensitivity of the glacier snout to climatic changes. When the glacier reaches the topographic threshold at 15 km it temporarily ceases to advance, but ice volume continues to increase until eventually the point is overrun. This indicates that the topographic threshold simply induces a geometric effect. In contrast, the topographic threshold at 18 km inhibits further changes in glacier length and volume. This suggests that the glacier essentially decouples from further climate deterioration beyond this point, and gives support to the notion that the boundary between mountain slopes and lowland plains on isolated massifs is a major barrier to glacier advance (Payne and Sugden, 1990, Kerr, 1993a and 1993b). This explains why this threshold is usually only crossed by glaciers in polar climates.

Summary

The mass balance of Sólheimajökull is more sensitive to changes in temperature than to changes in precipitation. In detail, the mass balance is most sensitive to changes in summer temperature and winter precipitation but a high sensitivity to changes in temperature during autumn and spring is also evident. The mass-balance sensitivity of Sólheimajökull (1.4 $ma^{-1}\ K^{-1}$) is larger than that calculated for most glaciers in continental Europe and Scandinavia (0.6-0.9 $ma^{-1}\ K^{-1}$) for which similar studies have been undertaken (Oerlemans and Fortuin, 1992).

Fluctuations in glacier length at Sólheimajökull are influenced by topography. Initial formation of the glacier occurs when the ELA intersects the caldera rim, and is controlled by the altitude-mass-balance feedback. The glacier grows from ice-free conditions to a length of 10 km without further lowering of the ELA. Fluctuations of the glacier front are reduced at points of trough widening, and enhanced in areas where flow is well constrained. The computed response times range from 30 to 90 years and can also be related to variations in topography; the length response time is shorter than the volume response beyond a glacier length of 18 km, where the glacier becomes locally stable at a topographic threshold.





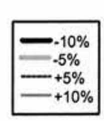
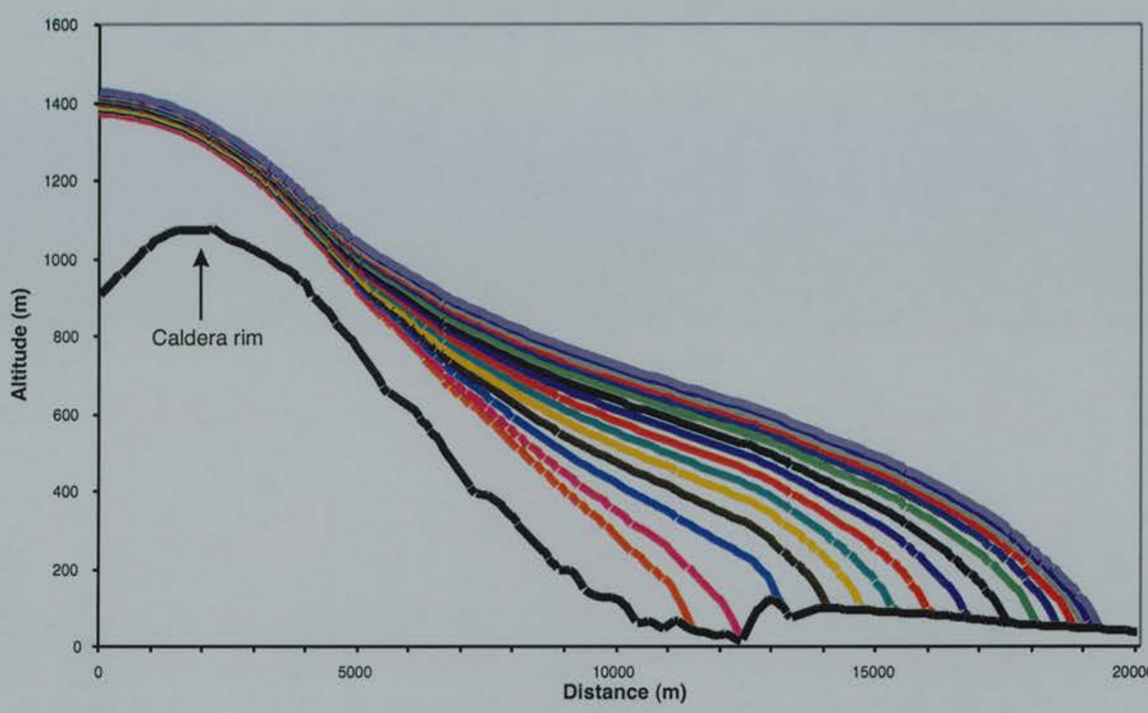


Figure 3.1

Changes in mass-balance with altitude calculated for seasonal perturbations in temperature and precipitation. Mass balance is sensitive to temperature changes all year around. In comparison, mass balance is only sensitive to precipitation changes in winter, spring and autumn.

a



b

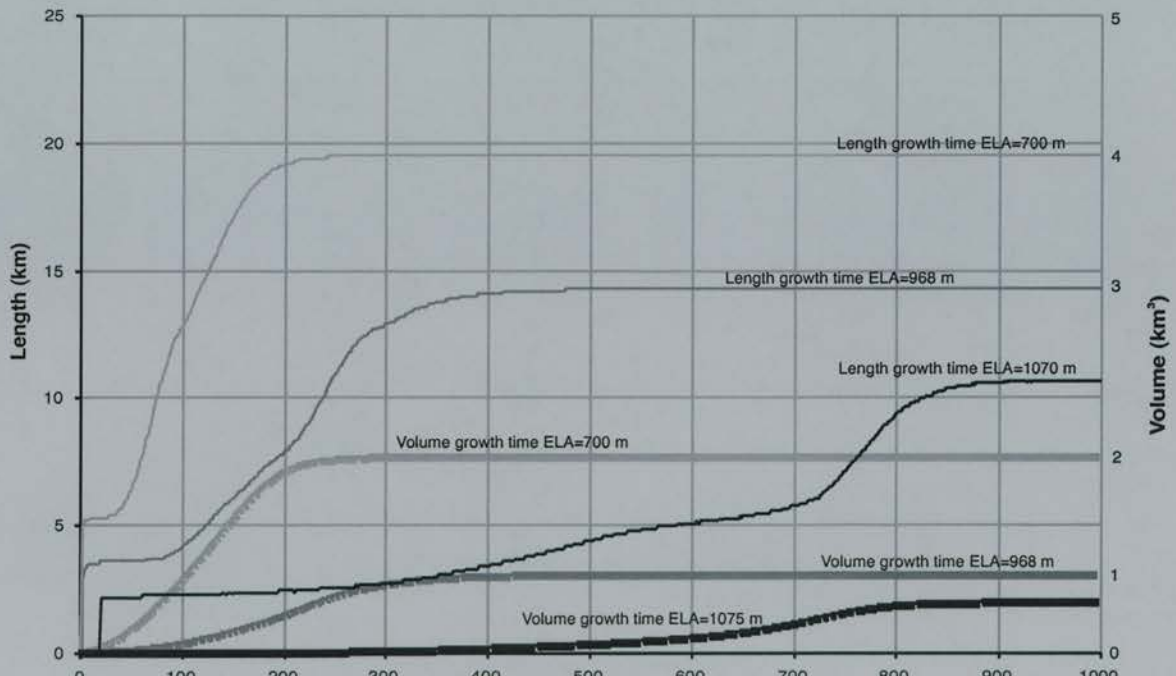


Figure 3.2

a. Simulated equilibrium surface profiles for Sólheimajökull. Each glacier profile represents the equilibrium position resulting from a step ELA lowering of 25 m from 1100 m to 700 m.

b. The time is shown for a glacier to grow to three different lengths/volumes from an initial state of zero length. The three different lengths/volumes are for a glacier extending to the end of the trough (ELA=700 m), to the 1996 length of 14.3 km (ELA=968 m) and to the first equilibrium profile in part a of this figure (ELA=1070 m). The growth time depends on the size of the mass balance perturbation, and is faster when a lower ELA is imposed.

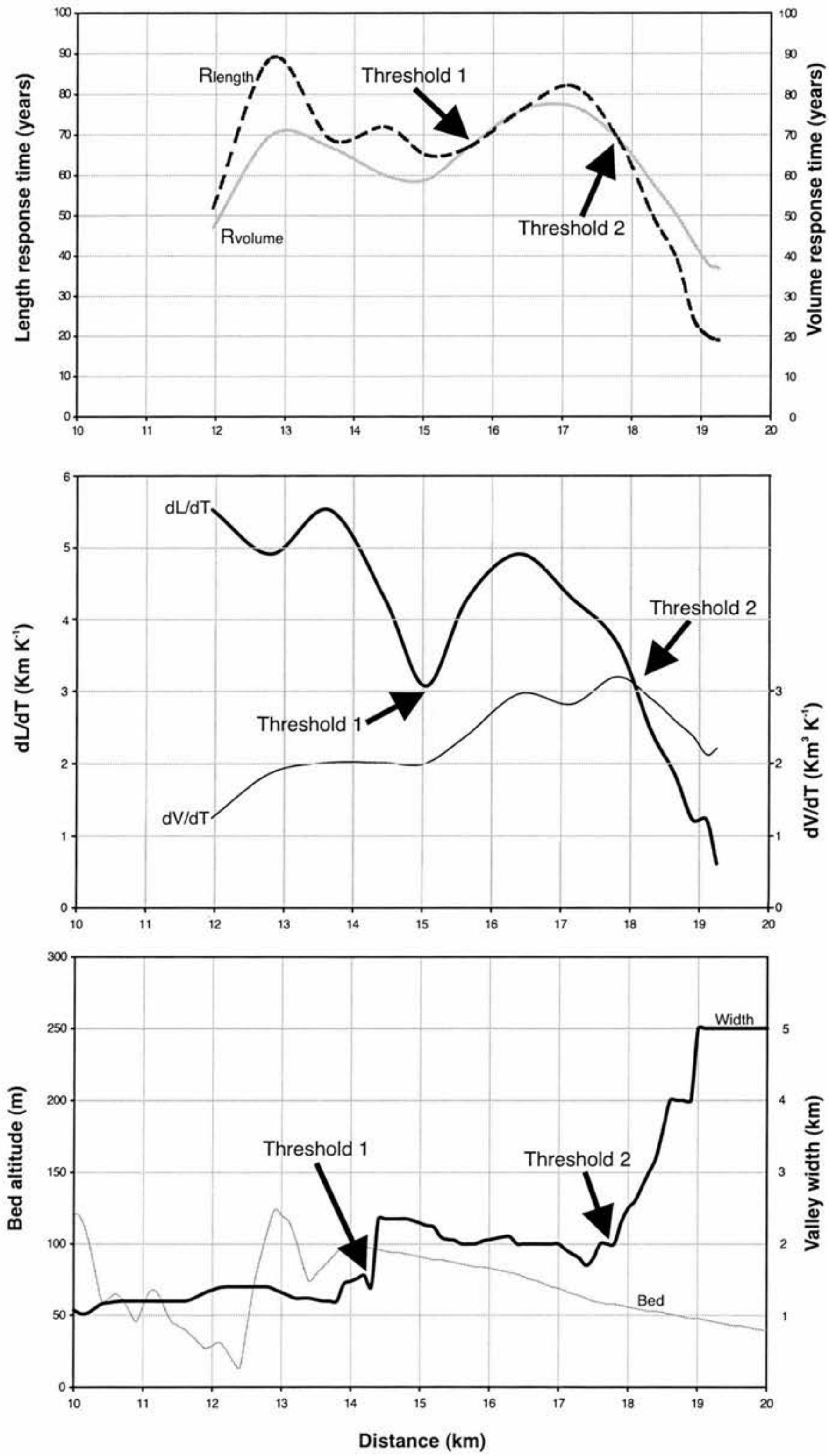


Figure 3.3

This composite figure shows how the topography of the glacier valley influences the evolution of glacier length, volume and response time. Two topographic thresholds are apparent. The first threshold occurs at the increase in valley width at a distance of 14.3 km. At this point dL/dT reaches a local minimum meaning that changes in glacier length are smaller for a uniform climatic change. The threshold has the greatest influence at a glacier length of 15 km, some distance after the valley becomes wider. The second topographic threshold occurs at a distance of 18 km. The influence of Threshold 2 is evident in dL/dT, dV/dT, and the response times. dL/dT and dV/dT decrease indicating a reduction in changes in length and volume for each uniform change in climate. Both the length and volume response times become smaller, but the length response time decreases most rapidly. This is because dL/dT approaches a value of zero, at which point the length response time would also reach a value of zero. In reality this means that most of the change in volume is achieved through thickening rather than a glacier advance.

Chapter 4: Reconstructing Past Climate from Past Changes in Glacier Length: Strengths and Limitations

Introduction

The aim of this chapter is to attempt a climatic reconstruction by modelling the record of glacier fluctuations spanning the past 5000 years at Sólheimajökull. In theory, a record of past glacier fluctuations can contain rich palaeoclimatic information. It is well known that yearly variations in glacier mass balance correlate with annual temperature and precipitation (Paterson, 1994). Global records of glacier fluctuations over time periods of about 100 years show broadly synchronous fluctuations (Oerlemans, 1994) and large valley glaciers register a smoothed climate signal, recording interdecadal trends in climate with a time delay of years to decades (Haeberli, 1995). The advantage of using glaciers for climatic reconstruction is that quantitative or semi-quantitative information can be derived about temperature and precipitation. This allows the relative magnitude of climatic changes during different periods of time to be compared, in a similar vein to oxygen isotope ratios in foraminifera or tree ring thicknesses.

A new climate proxy reconstruction is important for Iceland because the palaeoclimatic record is less constrained than in the European Alps or Scandinavia. Available climate proxy evidence indicates that the Holocene climate was variable on a millennial and centennial scale (Gudmundsson, 1997). For the period after decay of the inland ice sheet at around 8000-7000 years BP and prior to human occupation at around AD 900, information comes from studies of glacier fluctuations, pollen analysis, soil formation, permafrost and rock avalanche activity. 5-6 periods of climatic cooling are inferred from these studies, with Neoglaciation occurring some time after 5000 years BP (Gudmundsson, 1997). Temperature changes are poorly constrained and there is a need for a quantitative climate reconstruction. The most plausible explanation for the climatic changes are shifts in surface ocean conditions such as those reported by Koc *et al.* (1993), although as yet no attempt has been made to directly correlate climatic changes in Iceland with climate proxy records derived from offshore marine sediments.

The climatic picture during the last 1000 years is more complete, but a quantitative reconstruction is still lacking and there is still uncertainty regarding the timing and magnitude of cooling events during the Little Ice Age (Grove, 1988, Ogilvie, 1992, Gudmundsson, 1997). Historic documents have been used to reconstruct climatic conditions and sea ice extent at a annual to decadal resolution (Ogilvie, 1984, 1991, 1992, 1996). In this chapter, extensive use is made of the literature so that changes in ELA at Sólheimajökull can be linked to the wider Icelandic climate. In Chapter 7, this is developed further with regard to changes in sea ice extent around Iceland, and the climatic reconstruction is compared to Greenland where several well-known climate proxy records have been recovered from ice cores.

Glacier length variations cannot be used to directly infer climatic changes because a time lag due to the glacier response time occurs between a change in climate and the eventual response of a glacier. In addition, the sub-glacial topography and basal flow characteristics of glaciers can dampen or enhance a terminus response (Mercer, 1961, Furbish and Andrews, 1984, Oerlemans, 1989). The advantage of reconstructing climate from glacier fluctuations with a numerical model is that the response time and glacier dynamics associated with valley geometry are taken into consideration. However, a numerical modelling study still presents several challenges. Possible sources of uncertainty are listed below:

- During the majority of the last 5000 years, a climatic reconstruction can only be made by assuming that the glacier reached equilibrium with the climate. This is a valid assumption if climatic changes persisted for longer than the glacier response time, which has a range from 40-100 years. However in reality, climatic changes of large magnitude and short duration may have occurred. Climatic changes of this type will be reconstructed incorrectly by the glacier model.
- There is some uncertainty in the record of historic glacier length positions during the 18th Century.
- There is a possibility that the Holocene fluctuations of Sólheimajökull have been influenced non-climatic factors such as ice-divide migration on Myrdalsjökull (Dugmore and Sugden, 1991).
- There is also a possibility that sub-glacial eruptions of the Katla volcano underlying Myrdalsjökull have influenced glacier dynamics during the Holocene.

The importance of these factors will be assessed by comparing the climatic reconstruction at Sólheimajökull with other climate proxy records. This will allow the relative importance of temperature, precipitation or non-climatic factors in the Sólheimajökull ELA record to be assessed. In this way, we can determine the strengths and limitations of inverse modelling as means of reconstructing a climate proxy record.

Methodology

The theory of inferring a mass balance history from variations in glacier length was first presented by Nye (1965). Although this inverse method is effective for small changes in glacier extent, larger changes in glacier geometry require many adjustments of an initial reference state. It is more efficient to work directly with a numerical model such as the one described in Chapter 2 if large changes in glacier length are to be simulated. Numerical models have previously been used to reconstruct climatic trends from glacier length variations by Allison and Krus (1977) and Smith and Budd (1981). Several approaches can be taken in the climatic reconstruction. If accurate regular measurements of glacier length exist, then an inverse method can be used where a series of step functions (∂B_t in Equation 2.10) are fitted in order to reconstruct a mass-balance record. This was the approach taken by Nye (1965) and recently applied by Oerlemans (1997a). It is identical to the 'dynamic calibration' method presented in Chapter 2, except that here the emphasis is on the climatic information present in the reconstructions of ELA. The advantage of this approach is that glacier equilibrium does not have to be assumed. In other words, the time-dependent behavior of the glacier is included in the reconstruction. This is appropriate for periods when much historic information is available about glacier length (after AD 1705 at Sólheimajökull), or for periods when yearly measurements of glacier length are available (after AD 1932 at Sólheimajökull).

In cases where information is available on the extent and timing of a glacier fluctuation, but the time gaps between measurements are much larger than the response time of the glacier (40-100 years, Chapter 3), then a different approach must be taken. The only option is to assume that the glacier reached equilibrium. It is common to find examples in the literature where equilibrium is assumed during climatic reconstruction (Porter, 1975, Colhoun, 1985, Sugden, 1977, Dugmore and Sugden, 1991). With the exception of Sugden (1977) the concept of equilibrium and the potential errors of its assumption are rarely discussed. The problem is that if equilibrium is assumed then climatic changes can be greatly underestimated. This is most likely if the climate change is of large magnitude and of shorter duration than the response time of the glacier. It is sometimes possible to tell from geomorphic evidence how realistic an assumption of equilibrium might be. For example, if a large moraine is present and the debris is deposited at a normal rate then it can be assumed that the glacier snout was stationary for decades during deposition. In this case, an equilibrium assumption can be made with more assurance. The errors involved in assuming equilibrium can be minimised in climate reconstruction with an ice-flow model if the reference profile (Equation 2.11) and model flow parameters (Equation 2.8) are known accurately. Even so, the quality of the reconstruction will be lower than that of a transient model run.

Once an ELA trend has been reconstructed, it is possible to translate this to possible changes in temperature or precipitation. One of the limitations of climatic reconstruction using glacier length variations is that is not always possible to disentangle this signal. In practice, it may be possible to make some conclusions regarding the past climate because other climate proxy indicators reveal the overall climate. For example, in Iceland there is an excellent documentary record of climate detailed in books and journals.

The final point worth considering is the quality of the input data used and how this might influence the climatic reconstruction. If there are errors in the timing or extent of former glacier positions, the errors will follow through to the climatic reconstruction. The types of errors likely to exist depend on how the data was collected. Geomorphic evidence is usually very accurate in terms of glacier extent but the timing of fluctuations is not always well constrained. In contrast, evidence in historical documents is usually accurate in terms of the date of the sighting, but interpretation of former glacier length from descriptions or drawings can sometimes be subjective. It is likely that the uncertainty of the reconstruction will vary through time and this will be covered in more detail in the discussion.

Results

A test of the methodology: the 20th Century

The 20th Century fluctuations are used as a test case because glacier length records and climate statistics are almost complete. The test indicates the potential of the model for reconstructing climatic changes accurately. The ELA reconstruction is taken from the dynamic calibration run using Budd's initial flow parameters (Chapter 2). Figure 4.1 shows the ELA plotted against the instrumental record of temperature and precipitation. All three variables show the 20th Century to be a period of climatic variability. The century begins with a low ELA from 1905-1920. A dramatic rise in ELA occurs in 1920, with high ELAs persisting until the end of the 1930s. The period from 1940 to 1985 is characterised by lower ELAs, but not as low as at the beginning of the century. The ELA is again higher after 1985, but it does not reach the altitudes attained in the 1920s and 1930s. This pattern has a striking qualitative resemblance to the instrumental temperature record. For example, the 1920s and 1930s were the warmest of the century and the period from 1960-1990 was colder than the 1920s and 1930s, however it was not as cold as the years from 1900-1920 (Einarsson, 1991).

Now we take a closer look at the relationship between the reconstructed ELAs and climatic trends. With Equation 2.9 we can calculate envelopes of temperature and precipitation changes for the reconstructed ELAs. For example, if the change in ELA between 1920 to 1940 is assumed to represent temperature changes only, then a warming of 2.4°C is reconstructed with the model. Alternatively, a decrease in precipitation of 170% is required if temperatures remained constant. In reality, the Stykkisholmur climate station registered a 2.6°C increase in temperature between 1918 and 1932. Precipitation increased by up to 60% during some years and remained at pre-1918 levels during the majority of others (T. Jonsson, written comm., 1998). In this case, it is clear that changes in mass balance have been more influenced by temperature variability than by changes in precipitation. It is the opposite pattern of what would be expected if the ELA fluctuations were being driven primarily by precipitation. Krüger (1995) and Sigurdsson and Jonsson (1995) also consider temperature to be more important than precipitation in forcing 20th Century glacier fluctuations in Iceland. It is important to note that precipitation may have played a secondary role. For example, there has been a slight trend of increasing precipitation since the late 1960s, evident in the 20-year running mean (Figure 4.1). This may have been partly responsible for the positive mass balance in recent decades, but precipitation alone cannot explain the low ELA during this period.

It is encouraging to see that the timing of temperature change is captured. This is best illustrated by comparing the ELA with the 20-year running mean of temperature. In each case the change in ELA occurred within 5 years of the change in the sign of the trend line. The 20th Century climatic reconstruction picks up a smoothed trend and does not detect temperature changes at sub-decadal resolution. For example, the instrumental record shows a step-like lowering of temperature in the mid 1940s, the late 1960s, and the early 1980s (Figure 4.1), while the ELA reconstruction shows this period to be uniformly cool. It is also interesting to see that the model picks up the climatic warming that has been evident in Iceland since the mid 1980's.

Reconstructing earlier Holocene climatic changes

Now that the model is shown to be capable of reconstructing climatic changes, an attempt will be made to reconstruct Holocene climatic trends. The aim is to highlight the magnitude and duration of the climatic changes during different periods. ELA reconstructions for the 18th and 19th Centuries are presented first. This is followed by a ELA reconstruction for earlier Holocene limits back to 5000 years BP. ELA trends are presented as curves and they are then translated to possible changes in temperature or precipitation.

AD 1705-1900

Glacier length records from the 18th Century indicate impressive glacier fluctuations (Figure 4.2); in the early decades, the glacier was advancing. A rapid retreat dominated in the middle part of the century culminating in the glacier reaching its historic minimum position in 1783. In the late 1780s and 1790s the glacier began to advance rapidly and by the beginning of the 19th Century, the glacier had returned to its early 18th Century maximum position. Glacier length fluctuations during the 19th Century were less spectacular (Figure 4.2); the glacier remained at a length of 15.2 km until 1880, after which it began to retreat slowly.

The glacier length simulation for this period is started in AD 1670 with the glacier at an length of 14 km. The timing and position of this limit in 1670 is chosen on the inference that the glacier was known to 'have been advancing for some time' when it was visited in 1705 (Ahlmann and Thorarinsson, 1940). From 1705 onwards, the glacier length simulation represents the best fit for the known glacier length fluctuations (Figure 4.2). For this period, the climate reconstruction method follows Oerlemans (1997a) where the vertical changes in ELA over time are determined experimentally by repeating the simulations until a good match between known and simulated glacier length is made. After a number of attempts, the 'simulated and 'known' glacier length records begin to converge. In the final run (Figure 4.2), the difference between 'simulated and 'known' glacier fluctuations is minimised. The ELA forcing of the final modelling run is then presented as the climatic reconstruction. It must be emphasized that this means that it is not necessary to assume that Sólheimajökull reaches equilibrium with the climate.

The ELA reconstruction from 1705-1900 begins with a stepped ELA lowering from 1670 until 1705 (Figure 4.2). This is needed in order to start the glacier advance described in 1705. From 1705 until 1750 a stepped rise in ELA is reconstructed. This rise reached its maximum between 1730 and 1750, when the highest ELA of the entire Holocene reconstruction is attained. This is needed in order to cause a retreat of 2 km between 1730 and 1780. From 1760 to 1795 a very low ELA is reconstructed, the lowest ELA reconstructed for the Holocene. It allows for a fast advance between 1780 and 1794.

The ELA reconstruction for the 19th Century shows less variability (Figure 4.2). A high ELA is reconstructed initially. This is required to slow down the late 18th Century advance. A ELA of 950 m is reconstructed from 1820 until 1880. This corresponds to the period when the glacier snout was stationary at length of 15.2 km, and appears to indicate climatic stability. After 1880, a higher ELA is again reconstructed. This allows a slow glacier retreat to begin.

5000 years BP to AD 1670

The ELA trend between 5000 years BP and AD 1670 is reconstructed in a different manner. Equilibrium has to be assumed because the time break between each limit (hundreds or thousands of years) exceeds the response time of the glacier (40-100 years) and no information is available on glacier length during these time breaks. The simulation is started from ice-free conditions by lowering the ELA until the glacier reached equilibrium at the c. 5000 year BP limit (Figure 4.3). The ELA is then determined for each former ice limit so that there is a good match between known glacier positions and modelled equilibrium glacier length. Checks are also made to ensure that the areal extent of the glaciers at each limit have similar dimensions to the various glacier limits mapped by Dugmore (1987). Retreat is assumed to occur to a length of 16 km between each still-stand. This nominal length is chosen because field evidence indicates that the glacier remained larger than 16 km during most of the period between 5000 years BP and AD 1670 (Dugmore, 1987).

Figure 4.4 shows the simulated surface profiles and widths for the larger Holocene fluctuations of Sólheimajökull. A comparison of Figure 1.7 and Figure 4.4 shows that the aerial extent of the glacier fluctuations are well simulated. The simulated and mapped glacier expansion dated to 5000 years BP are in especially good agreement, both involving an increase in glacier area of 20 km^2 relative to the 1985 glacier position. The shape of the expanded glacier is also well simulated, having the form of a piedmont lobe beyond a length of 18 km as it spreads out onto the sandur plain. This is in agreement with field evidence (Dugmore, 1987).

The ELA reconstruction for the main Holocene still-stands is shown in Figure 4.3. The reconstructed ELA values are of similar magnitude and become slightly higher during each still-stand. This is the result of assuming that the glacier reached equilibrium with the climate for a series of glacier positions of decreasing extent. Perhaps surprisingly, the reconstructed ELAs for the large Holocene fluctuations are of similar magnitude to the ELAs reconstructed for the last three centuries. This is because it is assumed that the larger Holocene fluctuations are in equilibrium with climate. In contrast, during the last three centuries the glacier rarely approaches equilibrium with the climate, and the climatic changes are of large magnitude.

As a final point, it is worth considering whether the ELA reconstruction presented here is realistic in light of wider field evidence. One way of constraining the maximum amount of ELA lowering is to consider the highest level at which soil accumulated during the Holocene. An advantages of studying soil development in Iceland is that a well known stratigraphy of tephra layers is present in the soil profiles (Dugmore, 1989). It is possible to use tephrochronology to accurately date the timing and rate of soil formation. This may allow the maximum ELA lowering to be constrained if it is assumed that soil would not accumulate above the level of the ELA, because the ground at these altitudes would be covered by glaciers or permanent snow. This reasoning contains a large uncertainty, because the ELA on a valley glacier can sometimes be several hundred metres below the regional snowline (Meierding, 1982). However, the comparison does provide an interesting result; modelling indicates that the ELA remained above 800 m for most of the Holocene, and on several occasions the ELA dropped to minimum value of c. 700 m for a short period. Soil pit evidence at 5 sites in within 40 km of Sólheimajökull indicate that soil development throughout the last 6000 years occurred at a maximum altitude of 710 m. (A. Dugmore, written comm. 1999). Above this altitude, soil pits show discontinuous development and contain many layers dominated by locally derived lithic grains, as might be expected if the ground was periodically covered by a year-round snow cover. This suggests that soil pit data provide support for the magnitude of ELA lowering reconstructed in this modelling study.

Reconstruction of climatic envelopes

The reconstructed ELA trend is translated to an envelope of temperature and precipitation changes relative to the 1966-1996 mean by assuming each variable remains constant in turn. This gives an indication of absolute changes in temperature that may have occurred during the Holocene. The findings show that only small temperature changes are sufficient to simulate the Holocene fluctuations. Absolute Holocene temperatures were never higher than 2.1°C or lower than 1.6°C relative to the 1966-1996 mean (Figure 4.5). Alternatively, precipitation was c. 110% above or c. 145% below the 1966-1996 mean. The largest relative changes in temperature or precipitation may have occurred around 5000 years BP when a cooling of 2°C or an increase in precipitation of c. 140% is reconstructed, and between AD 1750 and 1800 when the model reconstructs a cumulative cooling of 3.6°C or an increase in precipitation of c. 250%. The largest reconstructed increase in temperature is 2°C or decrease in precipitation is 70% relative to the 1966-1996 mean. This is for the first half of the 17^{th} Century.

Discussion

Climatic reconstructions

One way of assessing the quality of the climate reconstruction at Sólheimajökull is to compare the reconstructed ELA record with the uniquely rich documentary record of Icelandic climate (Ogilvie, 1984, 1991, 1992, 1996). For earlier parts of the Holocene where a documentary record is absent, the reconstructed ELA record is compared to wider climate proxy evidence. This may also allow temperature or precipitation signals to be disentangled in the ELA reconstruction. It may also allow the possibility of discovering matches and mismatches between the climatic reconstruction at Sólheimajökull and other climate proxy records.

AD 1700 to 1900

The Sólheimajökull ELA reconstruction shows the 18^{th} Century experienced great climatic variability (Figure 4.5). This is also evident in the record from historic documents (Ogilvie, 1992). At Sólheimajökull, a cold or wet phases is identified between AD 1670 and 1700. In the documentary record, these decades are identified as being cold and the 1670s and 1680s saw a return to colder regime. The last decade of the 17^{th} Century was especially cold. Sea ice was present along the Icelandic coastline during most years, and in 1694 it is said to have penetrated as far as the Westman Islands, adjacent to Sólheimajökull (Ogilvie, 1992). This is an excellent match between the ELA reconstruction and the historic record. It illustrates that the glacier was responding to temperature rather than precipitation changes during these decades. If precipitation changes are not considered, the ELA reconstruction indicates that temperature during these decades was 0.4°C lower than the 1966-1996 mean (Figure 4.5).

The Sólheimajökull ELA reconstruction shows that the years from 1700 to 1750 saw a return to warmer or drier conditions. Again, this matches well with the historic record, at least from 1700 until 1730. In Iceland, these years were the mildest of the century. The decade from 1701 to 1710 was especially mild (Ogilvie, 1992). According to the documentary record, the 1730s saw a return to colder conditions on the whole although there was considerable year to year variability. The 1740s were undoubtedly cold, similar to the 1690s. The return to cold conditions does not seem to be picked up by the ELA reconstruction until 1750. Some possible reasons for this mismatch are discussed later in this chapter. On the whole, the ELA reconstruction does a good job, clearly indicating the mild decades at the beginning of the century. Again, it seems that the ELA reconstruction is picking up temperature trends rather than trends in precipitation. If precipitation changes are not considered, these decades were 1-2˚C warmer than the 1966-1996 mean temperature (Figure 4.5).

The most impressive match between the ELA reconstruction and the documentary record is evident in the last part of the 18th Century. The ELA reconstruction shows a deteriorating climate, reaching a peak of cold in the 1780s and 1790s. The documentary record indicates the 1750s were cold and invoked considerable hardship for the Icelanders (Ogilvie, 1992). This is because there were a large number of unfavourable summers in the 1750s. In Chapter 3 it was shown that the glacier mass balance at Sólheimajökull responds most to changes in summer temperature, so it is not difficult to understand why this decade is strongly highlighted in the ELA reconstruction. The model reconstructs a large drop in temperature (1.9˚C) at the beginning of the 1750s (Figure 4.5). The 1760s saw a return to slightly milder conditions in the documentary record. In the ELA reconstruction this is evident in a step in the cooling. Interestingly, both the Sólheimajökull ELA reconstruction and the documentary record indicate that the latter part of the 18th Century was a time of maximum climatic deterioration. The Sólheimajökull ELA reached its lowest point of the Holocene reconstruction. The documentary record indicates that the 1780s were the coldest decade in southern Iceland between 1500 and 1800 (Ogilvie, 1992). During the 1780s, sea ice was present every year off the Icelandic coastline. During 1782, the sea ice reached the south coast adjacent to Sólheimajökull and remained until August (Ogilvie, 1992). This excellent match shows that it is possible to reconstruct the timing and magnitude of climatic changes with an inverse modelling approach. Again, the glacier is clearly picking up temperature trends. According to the model, temperature was 1.6˚C lower than the 1966-1996 mean if precipitation changes are not considered (Figure 4.5).

The reconstructed ELA for the 19th Century shows the climate to be less variable. Climate proxy sources for the 19th Century are of high quality and instrumental records became available from 1845 at Stykkisholmer in west Iceland. In agreement with the Sólheimajökull ELA record, they indicate that the 19th Century was cold, but not as cold as the late 18th Century (T. Jonsson, written comm. 1998). Nevertheless, there are questions concerning some of the details. The ELA reconstruction indicates that the first two decades of the century were warmer or wetter than the 1966-1996 mean, while the documentary record indicates cold conditions, although not as severe as the 1780s and 1790s (Ogilvie, 1996). Cold phases in the middle decades of the century do not appear to be picked up by the model. Possible reasons for the mismatch will be presented in the next section. The match is better in later part of the century when the ELA reconstruction and the instrumental record are in agreement and show a slight improvement in the climate. If precipitation changes are not considered, the ELA reconstruction indicates that most of the 19th Century was only 0.1˚C colder than the 1966-1996 mean temperature (Figure 4.5).

In Figure 4.6 the Sólheimajökull ELA record is compared to the climate proxy record in Iceland as is presently known from evidence of glacier fluctuations, permafrost activity, rock avalanche activity, a temperature index based on sea ice extent, and a temperature index for the Northern Hemisphere based on long instrumental temperature records (Gudmundsson, 1997). It is clear that the new ELA/temperature reconstruction is considerably more detailed, showing decadal trends and large fluctuations that do not show up in other climate proxy evidence. This is the first time that geomorphic evidence has been used to reconstruct a climate signal that is of a similar resolution to documentary evidence in Iceland. It is good news because previously it was thought that the two different forms of climate proxy evidence were at odds, where now it is clear that previous attempts at climate reconstruction were restricted by the temporal resolution of the geomorphic evidence.

There is general agreement between the Sólheimajökull record and the northern hemisphere temperature index (Figure 4.6), although it is clear that temperature changes in Iceland are larger than mean hemispheric changes. This is to be expected because of Iceland's location near the atmospheric and oceanic polar fronts. It is also apparent that the cooling event in the late 18th Century in Iceland was not as significant more widely in the Northern Hemisphere. This point is considered in more depth in Chapter 7.

c. *AD 900 to 1700*

Several authors have suggested that the climate during the settlement period of Iceland (AD 874-930) and for several centuries thereafter was stable and possibly warmer than at present in the North Atlantic (Dansgaard *et al.*, 1975, Lamb, 1977, Ogilvie, 1991, Bjornsson, 1979). This has recently been supported by evidence from Greenland which shows a stable, warm climate existed between c. AD 730 to 1100 (Jennings and Wiener, 1996, Meese *et al.*, 1994). Geomorphic evidence indicates that Sólheimajökull retreated from a length of c. 16 km to a length of around 14 km during this period (Dugmore written. comm. 1999). However, the retreat was broken by a glacier still-stand or advance at c. AD 920-930. An advance or still-stand is also known to have occurred around c. AD 1000 in northern Iceland (Stötter *et al.*, 1999). The ELA reconstruction indicates that the glacier still-stand dated c. 920-930 could have resulted from conditions only 0.24°C colder or 16% wetter than present. This suggests that the period may not have been uniformly favourable, and that a short period of climatic deterioration occurred within an overall period of climate amelioration. Without further evidence, it is not possible to indicate whether this climatic change resulted from a decrease in temperature or an increase in precipitation.

Little direct evidence is available from the model for the period from c. AD 930 to AD 1350. A glacier still-stand occurred at AD 1350. The ELA reconstruction indicates a climate that was marginally colder or wetter than at present during this time. This pattern is consistent with the documentary record of Icelandic climate which gives circumstantial evidence for a mild climate until the end of the 12th Century and a variable climate thereafter through the 13th and 14th Centuries, including decades where sea ice was present around Iceland (Ogilvie, 1991). Sólheimajökull and Oerafajökull are the only glaciers known to have advanced during this time (Dugmore, 1987, Gudmundsson, 1998). Nonetheless, the marginal evidence indicates that temperatures were probably low during this period. If precipitation changes are not considered, the ELA reconstruction corresponds to a temperature lowering of 0.1°C relative to the 1966-1996 mean.

Evidence is again poor between AD 1350 and c. AD 1705. Glaciers are not known to have expanded during this period, and although the documentary record is limited by a lack of contemporary sources between AD 1430 and 1650, circumstantial evidence indicates that a mild period occurred between AD 1395 and 1470, and more reliable evidence indicates another mild period from AD 1640 to 1690 (Ogilvie, 1991). The intervening periods are believed to have been climatically variable but not exceptionally cold (Ogilvie, 1991). It must be emphasised that the ELA reconstruction is poorly constrained, but is in agreement with the historic record. If precipitation changes are not considered, the temperature was 0.4°C higher than the 1966-1996 mean.

5000 to 1400 years BP

The Sólheimajökull ELA reconstruction indicates that the Holocene was punctuated by several periods where temperatures were at least 1°C colder or 70% wetter than the 1966-1996 mean. Because the glacier was assumed to reach equilibrium with the climate during the simulation of these ice limits, it must be emphasised that these are minimum estimates. Very little can be said about the intervals between dated ice limits, except that warming or decrease in precipitation must have occurred in order to explain the overall pattern of retreat.

The overall timing of Holocene glacier fluctuations at Sólheimajökull coincides with fluctuations of other glaciers in Iceland (Figure 4.7). The ELA reconstruction also correlates with a qualitative temperature reconstruction based on a combination of climate proxy indicators including pollen analysis, soil formation, permafrost, and rock avalanche activity (Figure 4.7) (Gudmundsson, 1997). This suggests that the ELA fluctuations resulted mainly from temperature changes. Each 'cold' stage seems to have persisted for c. 1500 years. Geomorphic evidence in Iceland indicates glacier expansions with similar timing to that of Sólheimajökull at Oerafajökull and Trollaskagi. The first Holocene cold stage is bracketed by ^{14}C dates to between c. 6000 and 4600 years BP (Stötter *et al.*, 1999, Gudmundsson, 1998, Dugmore, 1989). Elsewhere, mountain glaciers expanded in both hemispheres between 5800 and 4900 calendar years BP (Denton and Karlen, 1973). The second cold stage occurred between 3470 and c. 2900 years BP in northern Iceland and at c. 3300 years BP at Oerafajökull (Stötter *et al.*, 1999, Gudmundsson, 1998). Glacier fluctuations are known between 3300 and 2400 BP from both hemispheres and in some areas this was the most severe Holocene cooling event (Denton and Karlen, 1973). There was also another period of climatic cooling at around 1400 years BP (c. AD 570). Glacier expansions are known from other parts of Iceland during the earlier part of this period, and were not necessarily synchronous across the island. For example, moraines dated to c. 1555 ^{14}C years BP and c. 1700 years BP are known from Trollaskagi and Oerafajökull (Stötter *et al.*, 1999, Gudmundsson, 1998). Glacier expansions from other mountainous parts of the world indicate a shorter duration or smaller magnitude climatic deterioration than during earlier Holocene climate events (Denton and Karlen, 1973).

The ELA reconstruction is not a wholly reliable climatic indicator for the period from 5000 years BP because Sólheimajökull is assumed to have reached equilibrium with the climate. The most important finding is that the ELA reconstruction indicates that only small changes in temperature or precipitation are required in order to explain large changes in glacial extent. This result not expected, considering the large scale of the glacier fluctuations. The dramatic changes in glacial extent can be explained by the large mass-balance sensitivity and sub-glacial geometry of Sólheimajökull, which has been discussed in Chapter 3. With the limited evidence available, it is not possible to disentangle this climatic signal precisely as the climatic changes might have resulted from any number of combinations of temperature and precipitation changes. Nonetheless, because the fluctuations of Sólheimajökull occurred synchronously with other glaciers in Iceland and worldwide, it is more likely that the ELA reconstruction reflects changes in temperature in the order of 1°C.

Early Holocene

One unresolved question in the Holocene study of glacier fluctuations in Iceland is whether or not Icelandic glaciers persisted through the Holocene or melted entirely during a mid-Holocene climatic optimum (Gudmundsson, 1997). It was once believed that Icelandic glaciers disappeared during the early Holocene and did not reform until c. 2500 years BP (Bjornsson, 1979). Recent findings of mid-Holocene moraines at Sólheimajökull (Dugmore, 1987, 1989), Trollaskagi (Stötter *et al.*, 1999) and Oerafajökull (Gudmundsson, 1998) indicate that this is most likely incorrect. The period in Iceland when glaciers may have been absent is now bracketed to c. 3000 years, between 8000 years BP and 5000 years BP (Dugmore, 1989, Gudmundsson, 1998). This time also corresponds to the part of the Greenland Ice Cores where many melt layers can be found, indicating an insolation maximum and possibly higher temperatures than at present (O'Brian *et al.*, 1995). The modelling provides some constraint on the question of whether it is possible for Sólheimajökull to melt and reform during this short period. The results presented in Chapters 3 and 5 show that it takes less than 300 years for Sólheimajökull to melt entirely from a length of 14.3 km if conditions are slightly warmer than present, and about 300 years are required for the glacier to grow from ice-free conditions if a similar climate to that of the last few centuries is sustained. There is enough time for Sólheimajökull to melt and reform during the period between 8000 and 5000 years BP. Because this result relates directly to Sólheimajökull, is uncertain to what extent this finding can be generalized to all Icelandic glaciers. In Chapter 5 we will see that Sólheimajökull responds more quickly to climatic changes than Hofsjökull in central Iceland. It is possible that the interior ice caps might take longer to melt and reform.

What was the primary forcing mechanism of Holocene glacier fluctuations?

The aim of this final section is to look at the relationship between climate and glacier extent in more detail and to consider the role (if any) that non-climatic factors have played in the past 5000 years at Sólheimajökull. Uncertainties in the ELA reconstruction will also be considered in more detail.

During the 18th Century the main climatic trends appear to be captured by the model reconstruction. However, the mismatch in the middle of the century with the documentary record needs to be addressed because it suggests that the timing of the climatic change might have been reconstructed incorrectly. It is also possible that the magnitude of the favorable period between AD 1700 and 1750 has been overestimated, because the ELA reconstruction suggests it might have been warmer than the decades from AD 1920 to 1940, which is widely thought to be the warmest in recent history (Briffa and Jones, 1993). Two possible sources of uncertainty may have contributed toward the mismatch. The first and most likely problem is that the glacier length position in 1783 is in error. The description of the glacier position in the literature is reasonably clear relative to a named gorge, but the precise glacier position could be in error by up to +/- 250 m. The mismatch between the reconstructed ELA and documentary climate record could be easily explained if the amount of retreat is slightly overestimated. Another possibility is that the volcanic eruption of Katla in 1755 influenced the glacier.

Volcanic activity may influence glacier dynamics in several ways. Volcanic eruptions might potentially cause a glacier retreat if either a large amount of ice is melted in the accumulation area, or if a fine-grained thin blanket of dark tephra increases ablation by lowering the albedo of the glacier. As far as the author is aware, there is no case in the literature where this has been illustrated. Indeed, volcanic eruptions under glaciers seem either more likely to have minimal impact on ice dynamics or might be responsible for promoting an advance. For example the recent eruption of the Gjalp volcano under Vatnajökull had a limited impact on glacier dynamics, despite causing a large sub-glacial outburst flood (Björnsson *et al.*, 1998). In another documented example, tephra emitted from a volcanic eruption blanketed the ablation area of an Alaskan glacier triggering an advance (Sturm *et al.*, 1986). It is also possible that volcanic eruptions might influence glacier dynamics by changing sliding conditions at the glacier bed. In its most serious form this might result in an anomalous advance if increased meltwater causes hydraulic jacking (Kamb, 1987). Volcanic activity is believed to have caused the advance of a different glacier resting on a volcano in Alaska by altering basal sliding conditions (Sturm *et al.*, 1991).

The 19th Century ELA reconstruction is flat in comparison with the 18th and 20th Century ELA reconstructions. Although this might reflect a real climatic phenomenon, it seems possible that something is amiss in the reconstruction. In Iceland, many glaciers reached their Holocene maximum during this time. It has already been pointed out that the Little Ice Age maximum extent of Sólheimajökull was smaller than expected (Dugmore and Sugden, 1991). In 1850 the glacier was visited and described as thickening visibly against the hill Jökulhaus, but the snout was not advancing. One possible reason for this as indicated in Chapter 3 is that during nearly all of the 19th Century, the glacier was at a length of between 15 and 15.2 km (Figure 1.6). This is precisely the glacier length at which the glacier is influenced by a topographic threshold (Figure 3.3). The effect on the climatic reconstruction would reduce the amplitude of the ELA reconstruction. Indeed, small climatic fluctuations might not register as a change in glacier length at all and a climatic reconstruction might result in a flat signal. In summary, although the 19th Century climatic reconstruction appears to capture the main climatic trends, some climatic changes, for example the cooling in the 1850s and 1860s are not picked up by the model. This is because the glacier was located at a topographic threshold during this time.

The pattern of earlier Holocene fluctuations is considered in detail by Dugmore and Sugden (1991). They suggest that the large mid-Holocene extent of Sólheimajökull resulted from ice-catchment changes on the Myrdalsjökull ice cap rather than from a direct climatic response (Dugmore and Sugden, 1991). The modelling study provides new insight into this issue. The central argument of the ice-divide migration hypothesis presented by Dugmore and Sugden (1991) is that the ELA lowering required to simulate the large mid-Holocene extent of Sólheimajökull is large (-600 m) when compared to that inferred for nearby glaciers (<400 m) for the same period. In contrast, the ELA lowering required to simulate the Little Ice Age extent of Sólheimajökull is small (-50 m) when compared to nearby glaciers (-400 m). The apparent anomalies in ELAs on Sólheimajökull are accounted for by inferring that the ice-divide migrated southwards during the Holocene, progressively reducing the catchment area of Sólheimajökull.

Although this hypothesis is glaciologically plausible and provides easy explanation for the Holocene pattern of glacier fluctuations around Myrdalsjökull and Eyjafjallajökull (Dugmore, 1989), the modelling results presented here show that these ELA estimates are in error. The ELA lowering (relative to the 1966-1996 mean) reconstructed for the mid-Holocene lobe of Sólheimajökull during this study is lower (-165 m) than that estimated by Dugmore and Sugden (1991), while the maximum ELA lowering during the Little Ice Age is higher (-250 m). These errors result because the accumulation area ratios used by Dugmore and Sugden (1991) do not take the large sensitivity of the glacier into account and this results in an overestimation of ELA lowering. The transient behavior is not taken into consideration, resulting in an underestimation of ELA when the glacier is not in equilibrium. It is possible to have similar ELA lowering but a different glacier extent if Sólheimajökull does not reach equilibrium. For example if the ELA from 1766-1795 (690 m) is sustained for a period of c. 100 years the glacier continues expanding to a length of approximately 19 km (the Holocene maximum). Some uncertainty exists because equilibrium is still assumed for the larger ice limits in this study, and the real ELA lowering might be lower than 165 m. But while these new findings do not disprove the ice-divide migration hypothesis outright, they show that it is not the only way to explain the large Holocene glacier extent. In light of the recent finding that other Icelandic glaciers experienced their largest Holocene extent during the mid-Holocene (Gudmundsson, 1998, Stötter *et al.*, 1999), it seems more likely that Sólheimajökull exhibited a direct climatic response during the mid Holocene.

Summary

The inverse model achieves considerable success in reconstructing the climate of the 18^{th}, 19^{th} and 20^{th} Centuries. The magnitude and timing of climatic changes is well reconstructed. In addition, the modelling results provide quantitative information on climatic changes in Iceland. This has so far been lacking from climatic reconstructions during the 18^{th} and early 19^{th} Century in Iceland. The ELA reconstruction is best explained by changes in temperature rather than changes in precipitation. In agreement with the climate record derived from historic documents, the late 18^{th} Century was the coldest part of the climatic reconstruction. Temperatures were reconstructed as 1.6°C lower than the 1966-1996 mean during this time. The warmest part of the climatic reconstruction was the period from 1700 to 1750. Although this is in agreement with the documentary record from 1700 to 1730, there is a mismatch from 1730 to 1750. It also appears that the magnitude of warming at 2°C higher than the 1966-1996 mean may have been overestimated. The most likely cause of this error is that the glacier length position from 1783 is slightly incorrect.

To conclude, it is worth re-emphasising that very small changes in the ELA of Solheimajökull are required to simulate large changes in glacier length. This means that it is difficult to make quantitative evaluations of climatic change if former glacier positions are slightly in error, or if it is assumed that the glacier reached equilibrium with the climate. This makes it difficult to reconstruct climatic changes accurately for the period from 5000 years BP to AD 1700. Nonetheless, the ELA reconstruction shows that the glacier fluctuations during this period might have resulted from modest temperature changes of 1°C. More catastrophic explanations for glacier fluctuations such as volcanic eruptions or changes in the ice-divide of the Myrdalsjökull seem unnecessary in order to explain the Holocene behavior. The only case where local glacier dynamics seem to have influenced the climatic reconstruction is during the 19^{th} Century. This means that the climatic reconstruction for the 19^{th} Century is lacking in some detail. The bulk of evidence suggests that climate is the main forcing mechanism for changes in the length of Sólheimajökull and therefore the ELA record presented here can be viewed as a climate proxy record with few possible exceptions.

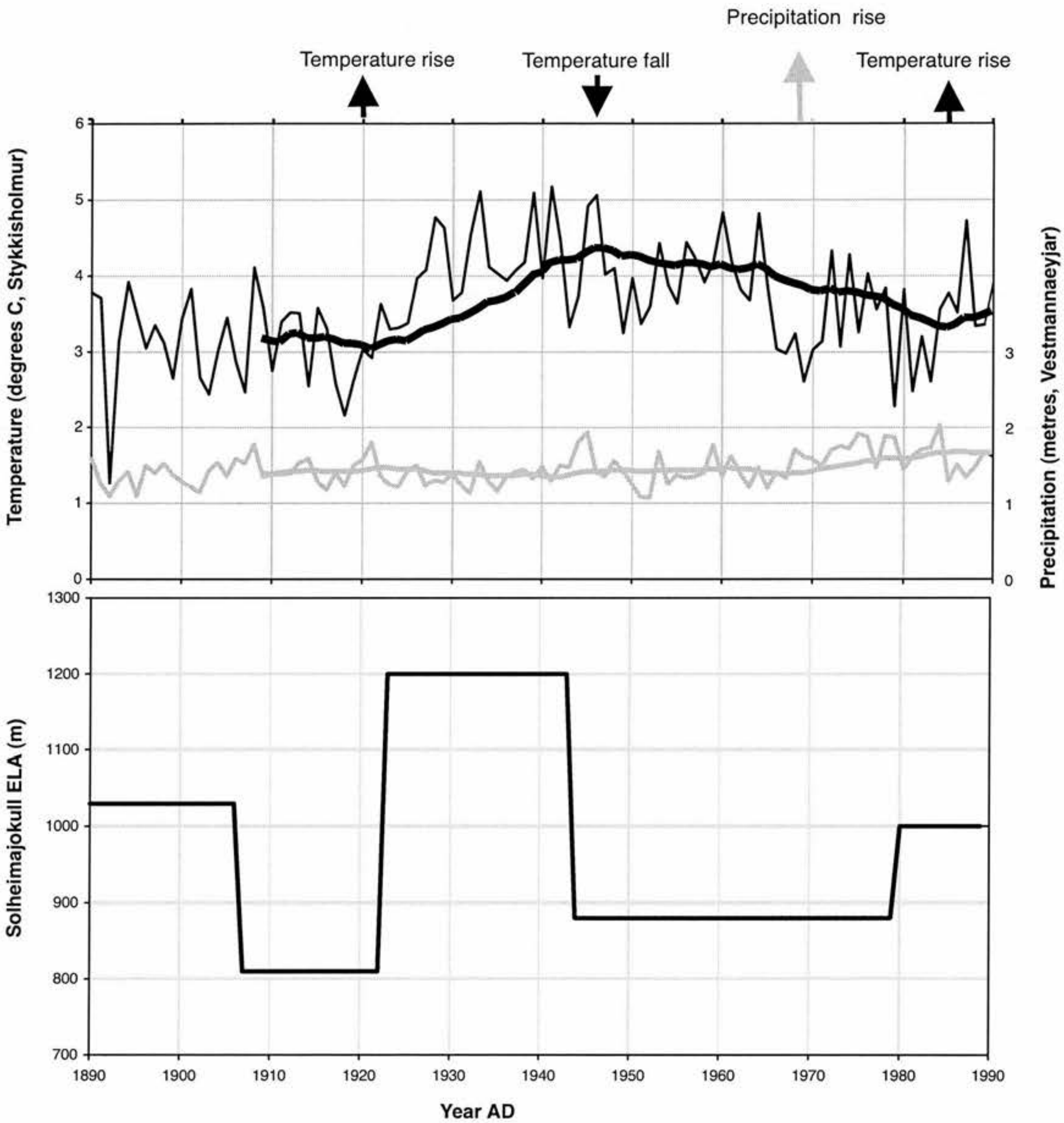


Figure 4.1

Reconstructed ELA and instrumental climate data 1890-1990 AD.

The top chart shows instrumental climate data from the Stykkisholmer climate station in western Iceland (Temperature, black line) and the Vestmannaeyjar climate station in southern Iceland (Precipitation, grey line). The data is presented as yearly values and a 20-year running mean (bold line).

The lower chart shows the reconstructed ELA trend over the same period for Sólheimajökull. The ELA reconstruction captures the main features of the instrumental temperature trend; the temperature rise from 1920-1940, the temperature decline from 1940-1985 and the temperature rise from 1985-1990.

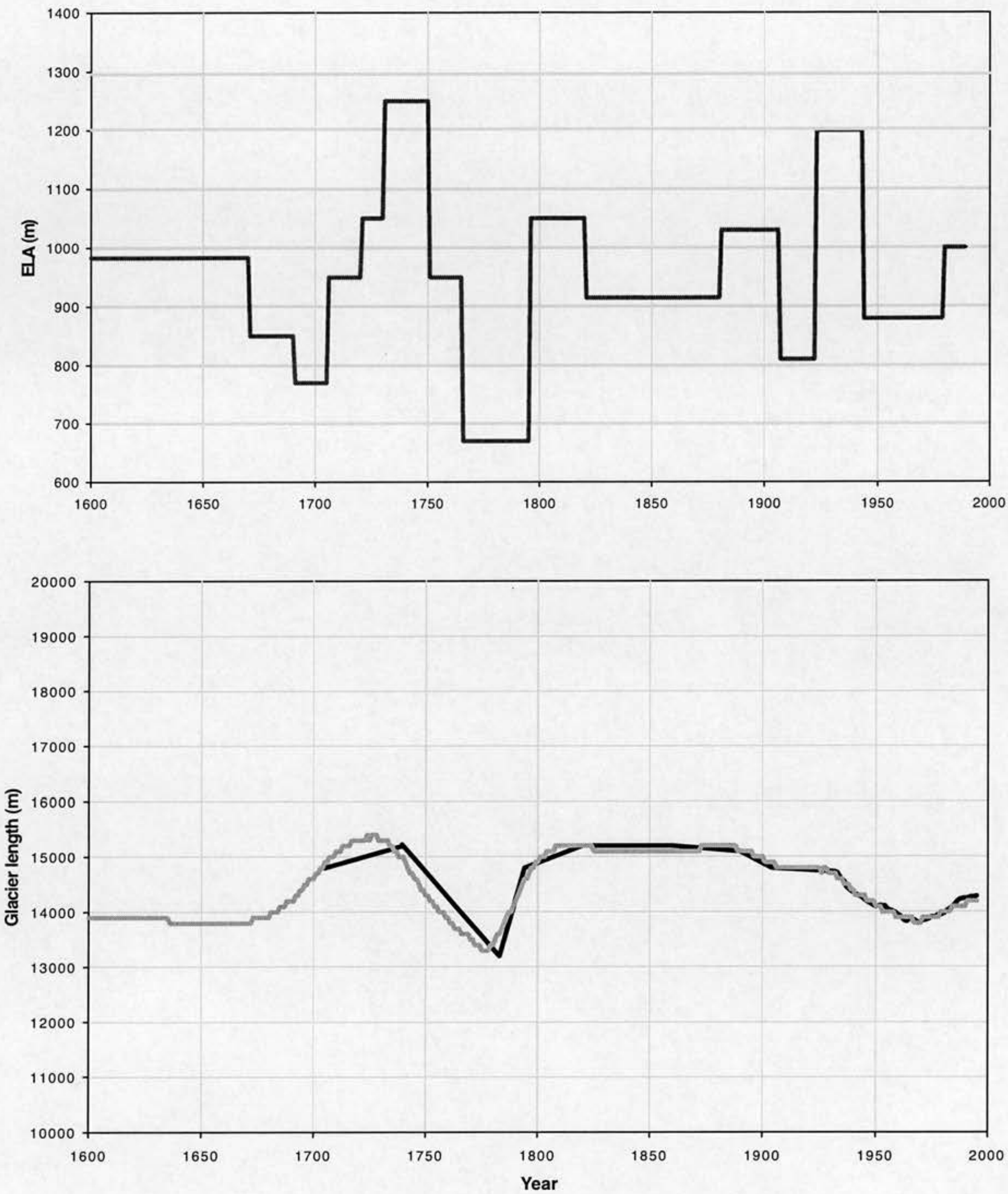


Figure 4.2

The lower chart shows the known glacier fluctuations (black line) as derived from a variety of historical sources (see Figure 1.5). The grey line represents the best-fit simulation to the known changes in glacier length as simulated with the ice flow model.

The upper chart shows the ELA reconstruction for AD1600-1990 as derived from the best-fit simulation with the ice flow model.

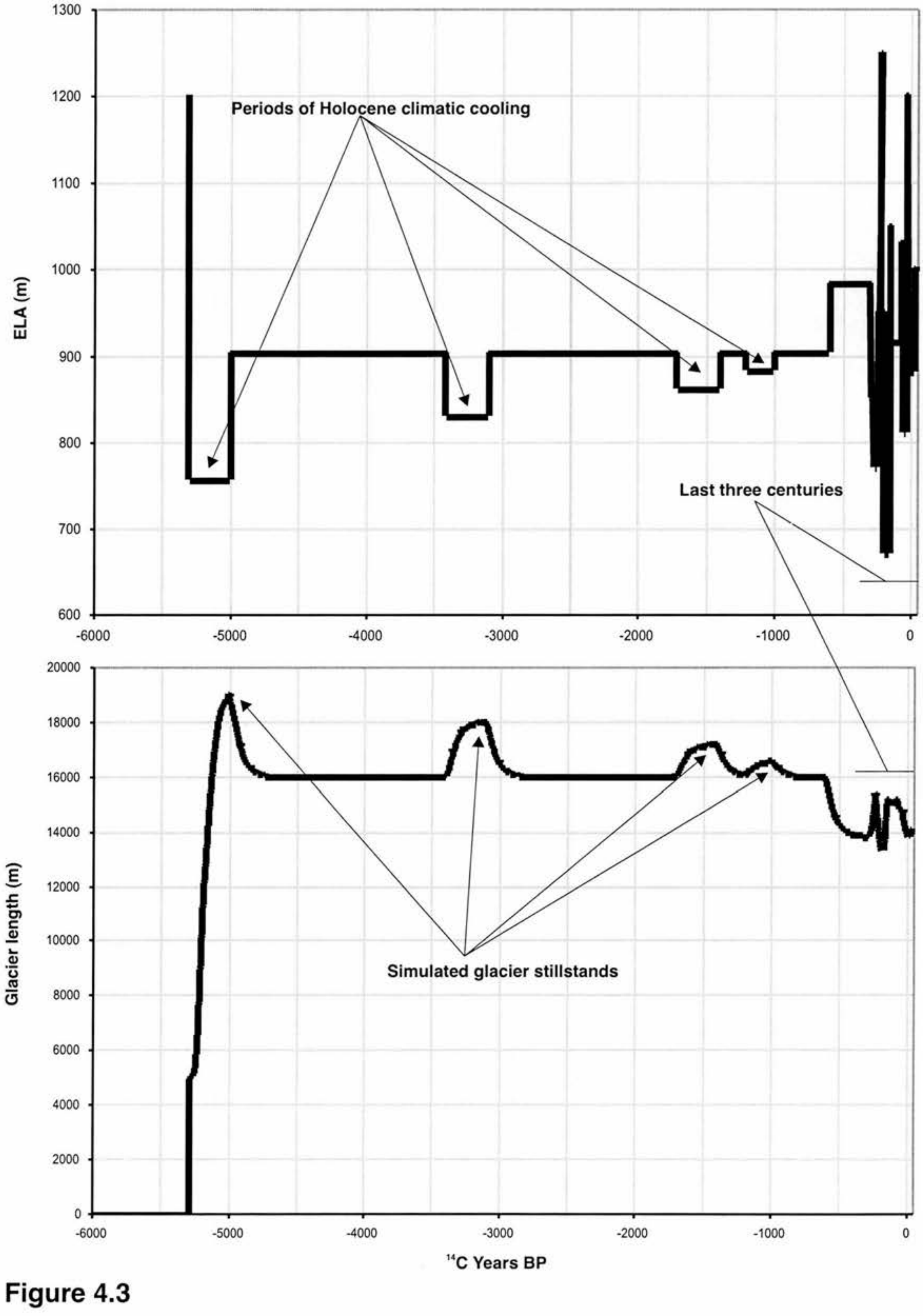
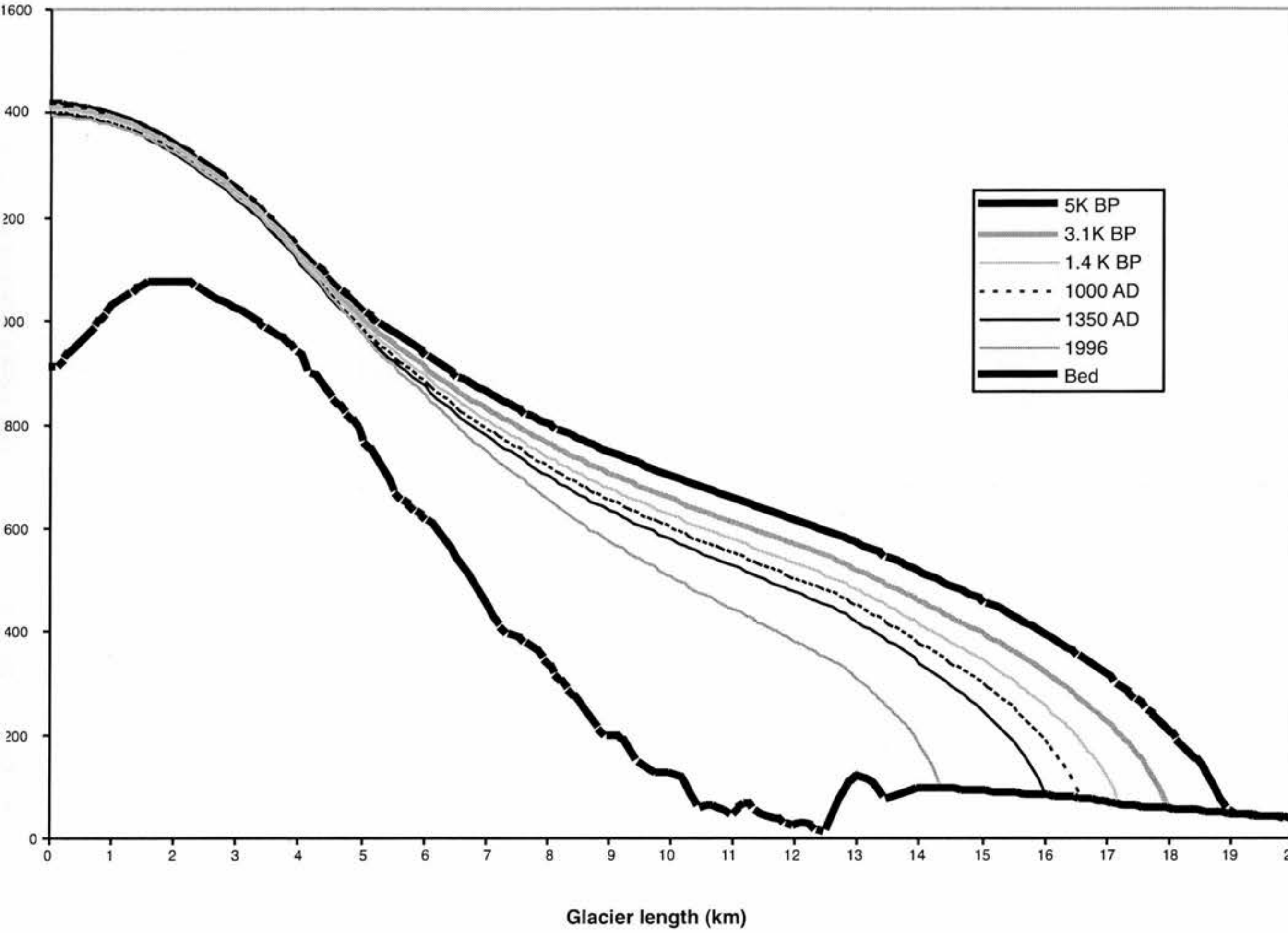


Figure 4.3

Reconstructed ELA record and simulated glacier length for Sólheimajökull from 5000 years BP to the present. The lower chart shows the glacier length simulation, where the glacier was assumed to have reached equilibrium with the climate during several glacier still-stands between 5000 years BP and recent centuries (at c. 300 years BP).

The corresponding ELA reconstruction is shown in the upper chart. The ELA reconstruction shows several periods of Holocene climatic deterioration that are of similar magnitude to the cooling events of the last three centuries.



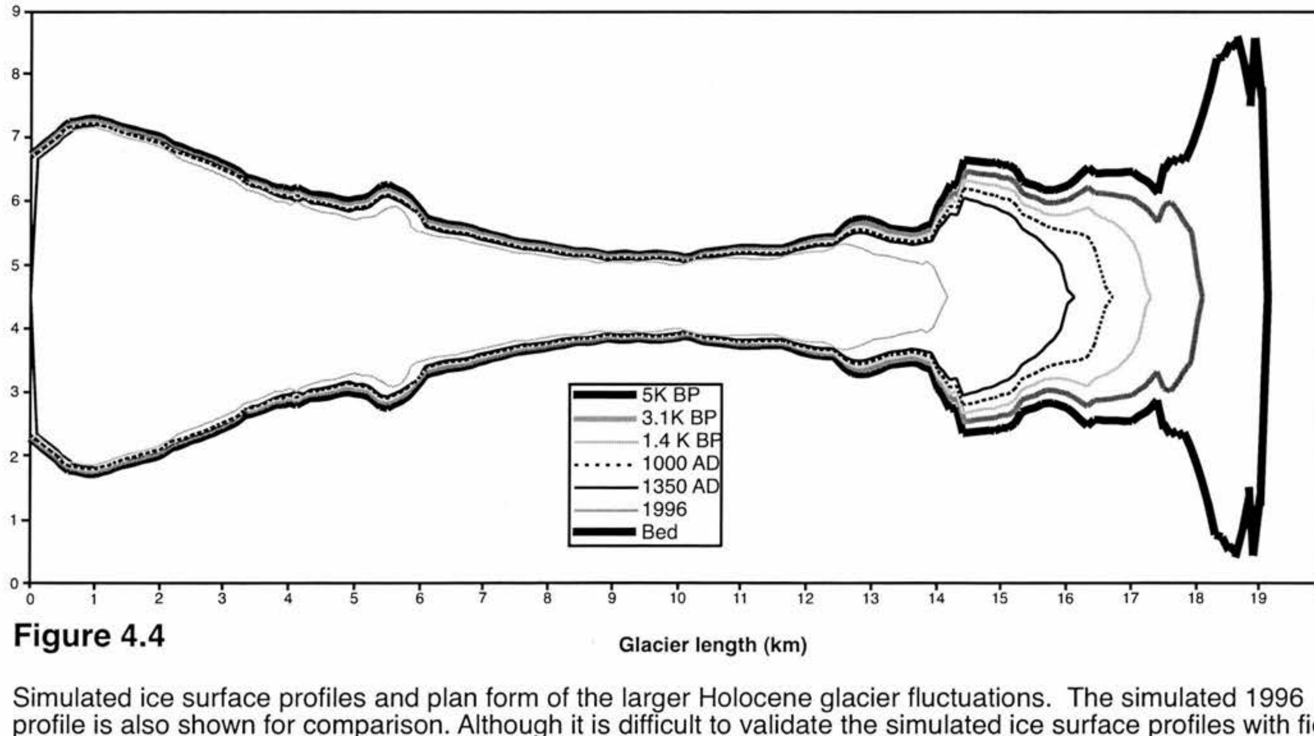


Figure 4.4

Simulated ice surface profiles and plan form of the larger Holocene glacier fluctuations. The simulated 1996 profile is also shown for comparison. Although it is difficult to validate the simulated ice surface profiles with field evidence, the simulated areal extent of the glacier fluctuations matches the map of Dugmore (1987) (Figure 1.6) well. The map and modelled glaciers both show that the largest Holocene fluctuation involved an increase in glaciated area of 20 km^2 relative to the recent ice limits. It is also evident that the glacier becomes wider at two points (15 km and 18 km) corresponding to the topographic thresholds identified in Chapter 3.

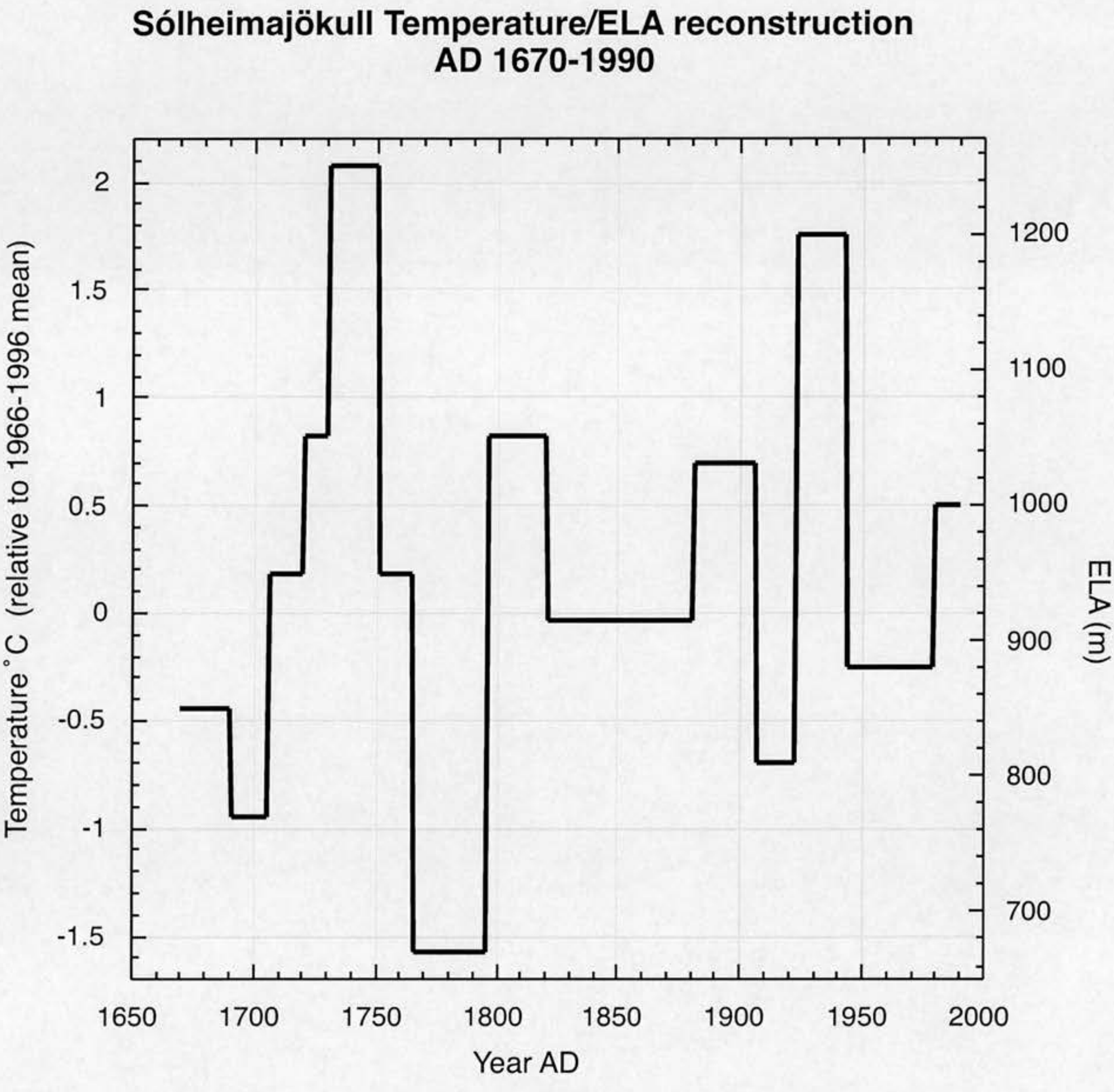


Figure 4.5

This chart illustrates the Sólheimajökull ELA reconstruction converted to temperature changes for the period from 1670 to 1990. The temperature is reconstructed relative to the 1966-1996 mean temperature. It is assumed that precipitation remained constant at 1966-1996 mean level.

The temperature fluctuations show large variability at a sub-century scale. The climate was especially cold from 1680-1700, 1780-1800 and from 1905-1920. Warm periods occurred from 1700-1730, and from 1920-1940.

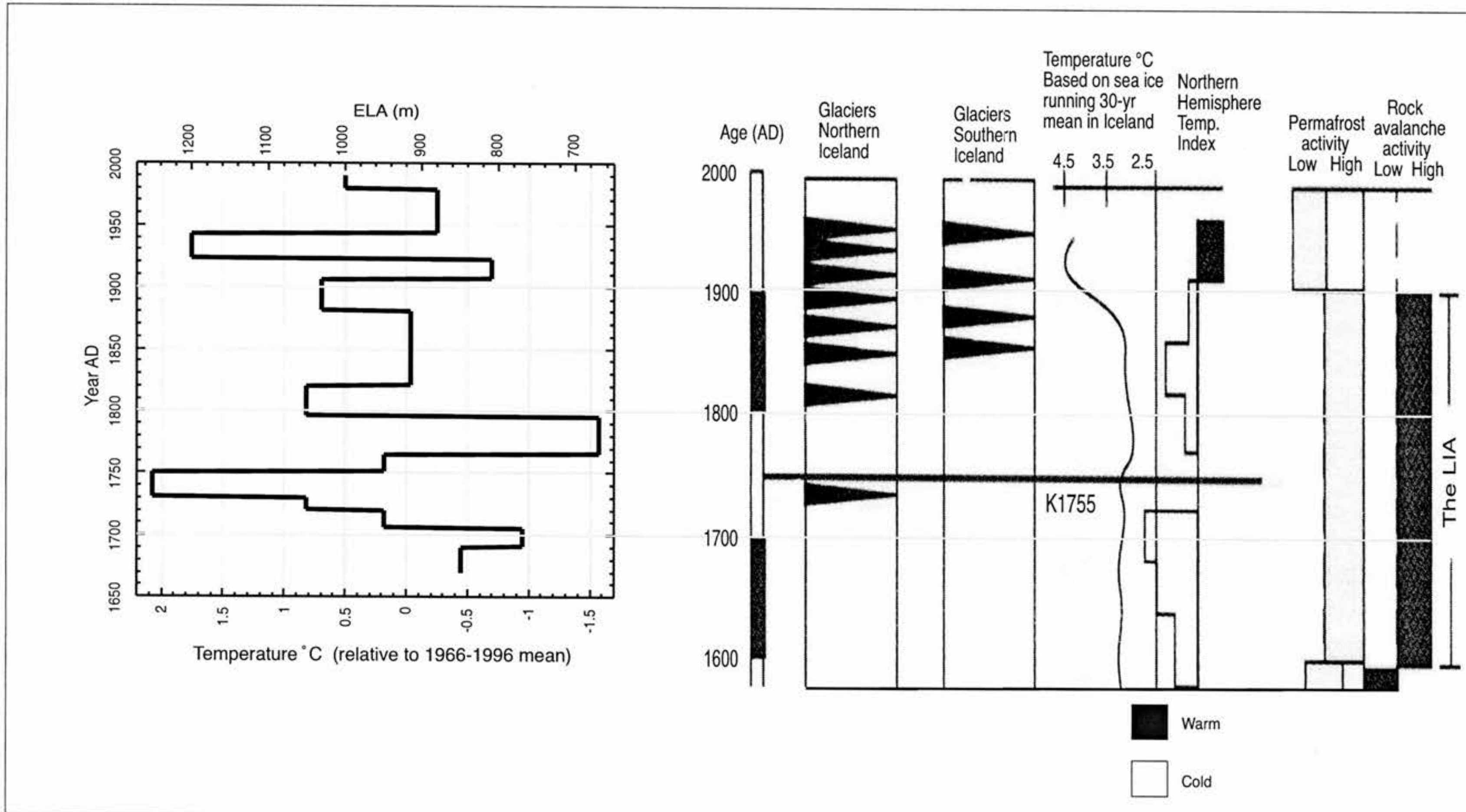


Figure 4.6

A comparison between the ELA/climate reconstruction at Sólheimajökull and wider climate proxy evidence in Iceland (adapted from Gudmundsson, 1997). The Sólheimajökull reconstruction shows larger and more frequent temperature oscillations than existing climate proxy evidence, including a warming event during the 'Little Ice Age' as defined by Gudmundsson (1997).

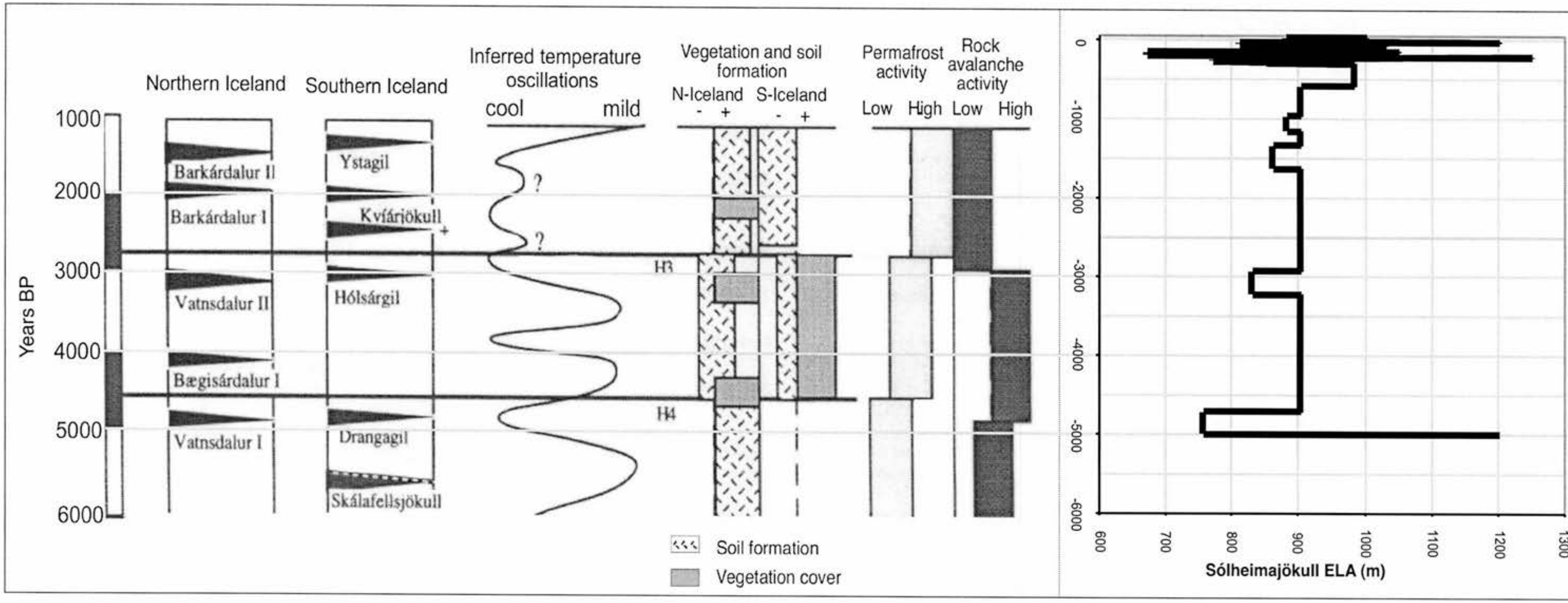


Figure 4.7

A comparison between the ELA reconstruction at Sólheimajökull and wider climate proxy evidence in Iceland (adapted from Gudmundsson, 1997). Periods of low ELA correlate with periods of cold climate, as inferred qualitatively from evidence of glacier fluctuations, soil and vegetation formation, permafrost and rock avalanche activity. Glacier fluctuations in northern and southern Iceland are relatively synchronous (the names of individual local advances are shown). Unfortunately the ELA reconstruction reveals little about intervening warm periods in Iceland.

Chapter 5: Predicting the Response of Sólheimajökull to Climatic Warming

Introduction

Over the last decade, concern over anthropogenic climatic change has provided a strong focus for research in the field of glaciology and climatic modelling. This has included major international interdisciplinary research programs such as the Intergovernmental Panel on Climatic Change (IPCC). Several reigonal reasearch programs have also made contributions; examples include the 'Scientific Committee on Antarctic Research, Global Changes and the Antarctic' program (SCAR GLOCHANT), the European Union funded MAGICS program, and in the UK, the NERC funded 'Arctic Ice and Environmental Variability' (ARCICE) program. One of the outcomes of these scientific programs has been the development of projections of future climate based on output from General Circulation Models (GCMs). This has naturally led several researches to investigate the response of glaciers to the climatic warming scenarios. The main objective of these studies has been to predict changes in global sea level resulting from the retreat of valley glaciers and small ice caps (Oerlemans and Fortuin, 1992, Oerlemans *et al.*, 1998).

The aim of this chapter is to quantify the response of Sólheimajökull to enhanced greenhouse warming. There are two main objectives. The first is to predict the changes in glacier length of Sólheimajökull to six warming scenarios that were recently used in EISMINT valley glacier modelling experiments on 12 different glaciers worldwide (Oerlemans *et al.*, 1998). It is hoped that this will help to broaden the results of Oerlemans *et al.*, (1998). It is useful to include Sólheimajökull because it is located on a highly maritime coastal margin in the central North Atlantic and has an area of 44 km^2. At present, the EISMINT study has included few glaciers of intermediate size (50 km^2) and their sample is biased toward the European Alps where many glaciers have recently been retreating. Maritime glaciers such as Sólheimajökull in Iceland and Nigardsbreen in southern Norway have recently been advancing (Dowedswell *et al.,* 1997). It is important to understand how an initial state of advance will influence the response of glaciers to climatic warming scenarios.

The second objective of this study is to help predict the response of Icelandic glaciers to climatic warming in more detail. A warming scenario for Nordic countries based on downscaling of output from a GCM (Johannesson *et al.*, 1995) is used so the results are specific to Iceland. The response of Icelandic glaciers to future warming has been predicted for Hofsjökull (Johannesson, 1991, 1997). However it is difficult to know to what extent these results are representative to the island as a whole. Iceland is characterised by a large climatic gradient and glaciers of very different geometry (Bjornsson, 1979). There are theoretical grounds for believing that outlet glaciers in south Iceland will have a larger and faster response to climatic change than the interior ice caps. The present study allows this hypothesis to be tested because Sólheimajökull and the outlets of Hofsjökull can be considered extreme cases. Sólheimajökull has a valley glacier geometry and high local precipitation, which results in the glacier snout descending to 100 m above sea level. In contrast, the outlets of Hofsjökull are broad flat lobes, precipitation is less and the glaciers terminate on a plateau or mountain slope at altitudes of c. 800 m. It is likely that the dynamic response of most Icelandic glaciers might lie between these two extremes. A study of Sólheimajökull will therefore allow the response of all Icelandic glaciers to climatic warming to be generalized with more certainty.

Another reason for examining the response of Icelandic glaciers to climatic change is to examine the local impact on the landscape. Although south-flowing maritime glaciers do not influence hydroelectric power generation like the outlets of Hofsjökull, changes in the length of Sólheimajökull might cause significant geological hazards that could result in loss of life. Sólheimajökull is known to periodically cause glacier floods by damming up local canyons, and it has also been the route of outburst floods following volcanic eruptions (Thorarinsson, 1940, Dugmore 1987). Of greater concern is the possibility that glacier retreat might promote a volcanic eruption in the Katla caldera underlying Myrdalsjökull. Recently, a significant correlation between glacial retreat and volcanic eruptions has been shown for the Oerafajökull volcano and overlying ice cap in south eastern Iceland (Gudmundsson, 1998). Thinning glaciers promote eruptions because they cause local isostatic uplift, which results in pressure release in the magma chamber of the underlying volcano.

Methodology

Experiment 1: Calculating the response of Sólheimajökull to EISMINT warming scenarios

A realistic way to simulate the response of glaciers to climatic change is to use a fully coupled mass balance glacier flow model such as that described in Chapter 2. This was the methodology used recently in the EISMINT experiments on a selection of 12 glaciers worldwide (Oerlemans *et al.*, 1998). The EISMINT methodology is followed in this study so that the results for Sólheimajökull can be directly compared to their findings. As part of the procedure has already been described in Chapter 2, it will be treated briefly here. Three steps are followed:

- A mass-balance history is reconstructed for the glacier from the record of historic length fluctuations. An attempt is made to simulate the fluctuations as closely as possible, by forcing the flow model with a reference mass-balance profile perturbed by a series of step functions. This is the 'dynamic calibration' method described in Chapter 2 and applied to reconstructing a mass-balance history in Chapter 4. The objective is to take the initial state of balance into account at the beginning of the future projection, and to guarantee that the model is well tuned.
- Seven future climate scenarios are considered. The simulation is started in 1990 and the model is integrated forward in time until AD 2100. During the first model run, it is assumed that the future climate will remain the same (in terms of mass balance) as that of the last 30 years, i.e. no climatic warming. In subsequent experiments, the warming scenarios considered are temperature increases of 0.1, 0.2 and 0.4 K per decade. The experiment is also repeated for each warming scenario to include the effects of an increase in precipitation of 10% per degree of warming

- The time-dependent response of the glacier in each warming experiment is normalised against its 1990 volume. This allows the response of Sólheimajökull to be compared to other glaciers in the EISMINT study.

The initial conditions are the glacier surface profile and state of mass balance of Sólheimajökull in the year 1990 (advancing). These values are taken from the dynamic calibration experiment in Chapter 2. The mass-balance gradient is assumed to be the reference mass-balance profile (Equation 2.11). The initial ELA of 920 m is used. This is the 1961-1990 mean ELA as determined in the dynamic calibration. For each warming scenario, the model is run forward in decadal time slices starting at the year 1990. At the beginning of each new decade, the warming rate and increase in precipitation is imposed on the glacier as a step change. This is calculated from Equation 2.14 as a vertical shift in ELA. The procedure is repeated until the simulation reaches the year AD 2100.

Experiment 2: Calculating the response of Sólheimajökull to a Nordic warming scenario
A warming scenario specific to Nordic countries (Johannesson *et al.*, 1995) is also used to make projections of future extent for Sólheimajökull. This is the most realistic warming scenario currently available for Iceland. The objective of applying this scenario to Sólheimajökull is to provide a realistic assessment of how Icelandic glaciers might respond to climatic changes. The scenario specifies a warming of 0.25 K per decade in mid summer and 0.35 K in mid winter with a sinusoidal variation through the year (Johannesson *et al.*, 1995). Precipitation is assumed to increase by 5% per degree of warming. The methodology for calculating the retreat of Sólheimajökull for this warming scenario is identical to that described above in the EISMINT experiments; integration is started in 1990 with the ice-surface profile and mass-balance conditions derived from the dynamic calibration, and the model is forced in decadal time slices. However in this experiment, the integration is continued over a longer period from AD 1990 to the year AD 2200.

Results

Experiment 1

Figure 5.1 shows the response of Sólheimajökull to the seven EISMINT climatic change scenarios. A summary of the findings in terms of length changes, volume changes and ELA changes is shown in Table 5.1. The variability in response to the warming scenarios is very large, ranging from an increase in glacier length by 1 km for the 'no change' scenario to a reduction in length of 12 km by the year AD 2100 for a warming of 0.4 K/decade. With regard to volume changes, the glacier gains 20% in volume by AD 2100 if the climate remains similar to the 1961-1990 mean. Otherwise the glacier loses c. 20% of its volume by AD 2100 for a warming of 0.1 K/decade, c. 50% for a warming rate of 0.2K/decade and c. 85% for a 0.4 K/decade warming. The effect of an increase in local precipitation is to reduce the amount of glacier retreat, but for the scenarios considered here, the precipitation increase is not large enough to compensate for the increase in melting. A very large precipitation increase would be required to compensate for even the smallest amount of climatic warming. For example if the temperature increased by only one degree over the 21st Century, precipitation would have to increase by 70% in order to keep the ELA at the 1961-1990 level.

Figure 5.2 shows how the retreat of Sólheimajökull compares to the other glaciers in the EISMINT warming scenario. All volumes are normalized relative to each individual glacier's volume in 1990. Sólheimajökull has a similar response to the larger valley glaciers in the Swiss and French Alps which are currently advancing or in balance, such as the Unterer Grindelwaldgletscher, Rhonegletscher, and Glacier d' Argentière. Sólheimajökull exhibits a smaller change in normalised volume than Haut Glacier d' Arolla, Nigardsbreen, the Hintereisferner and Storglaciären. When the response of Sólheimajökull is plotted on a chart along with the mean scaled response of 12 glaciers in the EISMINT study, it is apparent that the retreat of Sólheimajökull is typical of glaciers in that study (Figure 5.3). The glacier looses 60% of its initial ice volume by the year AD 2100. In detail, Sólheimajökull has a slightly faster retreat rate than the mean retreat rate of the 12 glaciers in the study of Oerlemans *et al.* (1998). Yet the final change in volume is similar because Sólheimajökull is initially advancing, which reduces the overall reduction in ice volume.

Experiment 2

The response of Sólheimajökull to the warming scenario specified for Iceland (Johannesson *et al.*, 1995) is dramatic (Figure 5.4). Despite an initial state of advance, the glacier starts to retreat by the year 2020. By the year 2070 the rate of retreat becomes extremely rapid at c. 250 myr^{-1}. At this stage most of the glacier has become an ablation zone. By the year AD 2100, the glacier has retreated by 10 km to a length of 5 km. Changes in ice volume are shown in Figure 5.4. By the AD 2100, only 35% of the initial ice volume remains and the glacier has disappeared entirely by around AD 2180. A summary of the results is presented in Table 5.1.

The findings of this study allow the response of Sólheimajökull and Hofsjökull to climatic change to be compared. This is interesting because Sólheimajökull and Hofsjökull are influenced by a similar macro-scale climate, but have very different local climates and ice geometries. Figure 5.4 shows the predicted changes in length and volume for two outlet glaciers of Hofsjökull and for Sólheimajökull for the warming scenario of Johannesson *et al.* (1995) (T. Johannesson, written comm. 1999). The response of the two glaciers is quite different. The margin of Hofsjökull is predicted to retreat by c. 2.5 km by the year AD 2100 and in comparison, the terminus of Sólheimajökull is predicted to retreat by c. 10 km by AD 2100. Sólheimajökull also undergoes a larger relative (scaled) change in volume than Hofsjökull. By AD 2100, Sólheimajökull is predicted to decrease in volume by 65%, while the simulation shows that Hofsjökull decreases in volume by 40% over the same time period. By the end of the simulation, the volume of Hofsjökull and Sólheimajökull both approach zero.

Table 5.1: Predicted retreat of Sólheimajökull to climatic warming scenarios.

The glacier has an initial (1990) length of 14.3 km, volume of 3.06 km^3 and ELA of 920 m. The variability in response is very large, illustrating that climatic change scenarios need to be tightly constrained in order to accurately predict the response of Sólheimajökull.

Warming Scenario	Length at AD 2100 (km)	% of 1990 volume left at AD 2100	ELA at AD 2100 (m)
No change.	15.3	117	920
0.01	12.9	75	1096
0.01+	13.4	81	1071
0.02	7.7	43	1272
0.02+	10.0	50	1221
0.04	2.5	11	1624
0.04+	3.6	19	1522
Johannesson *et al.* 1995 (c. 0.03).	5.1	31	1382

Discussion

Comparing the response of Sólheimajökull to 12 glaciers worldwide

Comparing the response of Sólheimajökull to climatic change with the results of the EISMINT study (Figure 5.2) draws several conclusions. The glaciers exhibiting the largest response to the warming scenario were either:

- Small glaciers with short response times covering small elevational ranges. These glaciers have fast response times, and small increases in temperature can cause the ELA to quickly rise above the highest level of the glaciers.
- Glaciers already in a strongly retreating mode in 1990. These tend to be small glaciers or glaciers located in a continental setting. The continental setting is important as the largest temperature changes over the last century have occurred in the centre of continents (Briffa and Jones, 1993), and also because continental glaciers are less likely to have been influenced by an increase in precipitation. This is because they are located further from the ocean and from the influence of migrating storm tracks.
- Glaciers with a geometry that promote a large response. Nigarsbreen is the best example. It has an extremely flat accumulation area located on a plateau at an elevation close to the present ELA. A small rise in the ELA results in most of the glacier becoming an ablation zone.

One of the aims of this chapter was to see if Sólheimajökull underwent a proportionally large glacier response in comparison with other glaciers in the EISMINT study. This seemed likely because Sólheimajökull is located in a high precipitation area, and has a glacier geometry that promotes a large response to climatic change (Chapter 2). It can be concluded that is not the case and that the response of Sólheimajökull is similar to that of large valley glaciers in the European Alps. It is surprising that

Sólheimajökull undergoes a smaller response than Nigardsbreen considering that both glaciers have a similar climatic setting. This confirms that the response of Nigardsbreen to climatic warming is an exception, relating to its geometry. We should not necessarily expect other maritime outlet glaciers terminating around sea level to respond as dramatically to climatic change as Nigardsbreen.

An implication of this finding is that the local topography underlying a glacier greatly influences its response to climatic change, and this needs to be included in any attempt to predict future changes in ice volume. Therefore simple methods such as the 'fixed geometry approach' need to consider the individual geometry of glaciers explicitly. At present, schemes such as that used by Oerlemans and Fortuin (1992) do take geometry into account implicitly, because they predict a larger response of maritime glaciers to climatic change than continental glaciers. Part of the rationale for specifying this different response is the fact that maritime glaciers often descend to lower altitudes than continental glaciers. However this rationale relies on assumption that topography is relatively uniform from place to place. Large errors might result in making this assumption if the response of glaciers to climatic change is predicted for glaciers in contrasting topographic settings. For example, if the response of Icelandic glaciers is compared to that of the Alps, the fixed geometry approach will predict a larger change for Icelandic glaciers because local precipitation is higher in Iceland than in the Alps. However, the majority of Icelandic glaciers (with the exception of Sólheimajökull and a few other confined outlet glaciers) are ice caps resting on relatively undissected topography. In contrast, the topography of the Alps is highly dissected and valley glaciers are the norm. This means that most Icelandic glaciers would undergo a smaller change in volume than glaciers in the Alps in response to a climatic warming, despite the large amount of precipitation in Iceland. However, it must be kept in mind that the topography of Iceland is unique, and the possible source of error in the 'fixed geometry approach' described above would not be very important when considering global changes in ice volume.

Comparing the response of Sólheimajökull and Hofsjökull to climatic warming

For some time we have known that large gradients in mass-balance sensitivity exist across Icelandic glaciers (Ahlmann and Thorarinsson, 1940, Bjornsson, 1979). One result of this study is that it is now possible to quantify how these differences influence the response of Icelandic glaciers to climatic change. The results of Experiment 2 show that Sólheimajökull has a larger response than Hofsjökull to the climatic warming scenario of Johannesson *et al.* (1995), both in terms of changes in length and volume. The larger changes in length at Sólheimajökull are related to the geometry of the glacier. Large changes in the tongue-like terminus of Sólheimajökull result from relatively small changes in the volume of the ice cap. This is in contrast to Hofsjökull, which has a circular plan form. Here, large changes in glacier volume result in relatively small changes in the extent of the ice cap margin.

The findings of Experiment 2 indicate that changes in volume of Sólheimajökull are also larger than changes in the volume of Hofsjökull in response to climatic warming. This indicates that the different response of the two ice masses is not entirely geometrical. One reason for this difference is that Sólheimajökull has a larger mass-balance sensitivity than Hofsjökull. The mass-balance sensitivity of

Sólheimajökull was calculated with an energy-balance model in Chapter 2, while the mass-balance sensitivity of Hofsjökull was calculated with a degree-day model in Johanesson (1997). In this chapter, the different mass-balance sensitivity was accounted for by imposing an rise in ELA at Sólheimajökull of 160 m for each 1 K rise in temperature. This is a larger figure than that used by Johannesson (1997) at Hofsjökull, where a 110 m rise in ELA was imposed for every 1 K increase in temperature. While the calculated mass-balance sensitivities of Sólheimajökull and Hofsjökull no doubt contain some uncertainty, the finding appears to be robust. Sólheimajökull has a large mass-balance sensitivity because it descends to a low altitude. This means that the ablation season is long at Sólheimajökull, and that mass-balance is sensitive to climatic warming all year round. In contrast, the ablation season is shorter at Hofsjökull and melting occurs mainly during the summer months (Johannesson, 1997).

Another factor that contributes to the larger response of Sólheimajökull to climatic change in comparison to Hofsjökull is that Sólheimajökull has a shorter response time. In Chapter 3 we saw that Sólheimajökull has a volume response time of 40-70 years (at glacier lengths considered in this chapter). The response time of Hofsjökull is longer at c. 100 years (Johannesson, 1991, Johannesson, 1997). This means that changes in glacier volume are faster at Sólheimajökull than at Hofsjökull. In Figure 5.4 it is difficult to disentangle the relative importance of the mass-balance sensitivity and response times in explaining the different scale of response between Sólheimajökull and Hofsjökull.

Climatic warming scenarios and glacier response

In Experiment 1, we saw that the magnitude of glacial retreat at Sólheimajökull varies enormously, depending on the climatic scenario considered. This implies that it will be very difficult to estimate changes in the length and volume of Icelandic glaciers unless we are able to accurately predict future climatic changes. In Chapter 4 we saw that Sólheimajökull has undergone large fluctuations in glacial extent during recent history. This occurred before anthropognic influences on climatic changes may have become important, implying the existence of a significant natural variability in the climate system. The climatic warming scenario of Johannesson *et al.* (1995) was produced with statistical downscaling of output from a GCM. While the authors are clearly aware of the tentative nature of the scenario there are several issues that need to be considered further.

It is now becoming clear that the global climate system displays a wide spectrum of modes of variability as far as can be determined from a variety of observational records (Simmonds, 1997). In Chapter 4 we saw that the climate of Iceland in the last 300 years has experienced sub-century scale temperature changes of up to 2°C. An important consideration so far neglected in simulations of future climate is how these natural climatic cycles will influence the climate of Iceland over the coming centuries. It is probable that any climatic warming will be superimposed on a natural climatic cycle. This behavior may already be evident in the recent history of glacier fluctuations and climatic changes in Iceland. While 20th Century warming has been clearly evident in the interior of continents (Briffa and Jones, 1993), 20th Century climatic trends in the North Atlantic have been more complicated (Lamb 1979, Mysak and Power, 1991). A 20th Century cooling trend is evident in Iceland, which is related to an increased incidence of sea ice off the Icelandic coastline in the late 1960s, 1970s and early 1980s (Einarsson, 1991). In Chapter 3 it was shown that this cooling trend was responsible for the glacier advance of Sólheimajökull from 1970 to 1995. The problem is that GCMs do not currently take natural climatic variability associated with sea ice cycles into account. Indeed, this has been one of main criticisms of attempts to make future projections of climate in the North Atlantic (Battisti *et al.,* 1996). One implication might be that warming rates for the North Atlantic have been overestimated.

Another problem with GCM simulations is that that they average precipitation over wide areas where it is possible for precipitation changes to vary locally. Although the scenario of Johannesson *et al.* (1995) includes a predicted increase in precipitation, it is fairly modest change. This can be illustrated by comparing the predicted changes (5%) with the standard deviation of interannual variability in precipitation at the Icelandic climate station Vestmannaeyjar over the last 100 years (15%). Although an increase in precipitation of the order of 15% over a century would not have as dramatic an effect on glaciers as the predicted temperature rises, it would result in a smaller glacial retreat than predicted by Johannesson *et al.* (1995).

Until climatic warming scenarios become more realistic, it will be very difficult to predict how Icelandic glaciers will influence the landscape. The climatic warming scenario of Johannesson *et al.* (1995) is the best currently available for Iceland. The 0.1^{+} and 0.4 K/decade EISMINT warming scenarios can be viewed as extreme cases, and at this stage it is not possible to determine which projection of future glacier length is the most realistic. This means that it will be difficult to predict the timing and nature of glacier related hazards resulting from glacier retreat during the 21st Century.

Summary

The future response of Sólheimajökull to prescribed climate warming scenarios has been predicted with a numerical model. Sólheimajökull is projected to retreat by 10 km over the 21st Century, and will lose 65% of its ice volume if the climate warms by c. 3°C per decade as suggested for Nordic countries by Johannesson *et al.* (1995). Under this warming scenario, the glacier will melt entirely by around AD 2180. Sólheimajökull exhibits a wide range of responses to other climate warming scenarios. For example, if the climate warms by 0.1 degree per decade, and if precipitation increases by 10% for each degree of warming, Sólheimajökull will retreat by only 800 m and loose only 20% of its 1990 ice volume by the year AD 2100. In contrast, if the climate warms by 0.4 degrees per decade, then the glacier will retreat by 12 km and only 11% of the initial volume will remain by AD 2100.

Sólheimajökull exhibits a larger response to a climate warming scenario than the outlets of the Hofsjökull in central Iceland. One implication of this result is that the predicted retreat of Hofsjökull cannot be considered representative of all Icelandic glaciers. Any attempt to predict changes in ice volume for the whole of Iceland needs to consider the different response of coastal outlet glaciers and the broad lobes of glaciers in the interior. Interestingly, the response of Sólheimajökull to future warming scenarios is very similar to that predicted for large glaciers in the European Alps. This appears to reinforce the finding that topography strongly influences glacier response to climatic change.

It will be very difficult to predict changes in the volume and extent of Icelandic glaciers until climatic change scenarios are improved. Present GCMs do not adequately incorporate the effects of natural climatic variability associated with decadal-scale changes in sea ice extent in the North Atlantic. This natural climatic variability may result in large regional differences in glacial retreat rates, and may even promote periods of glacial advance during an overall pattern of retreat, as is evident in the 20^{th} Century pattern of glacier fluctuations at Sólheimajökull.

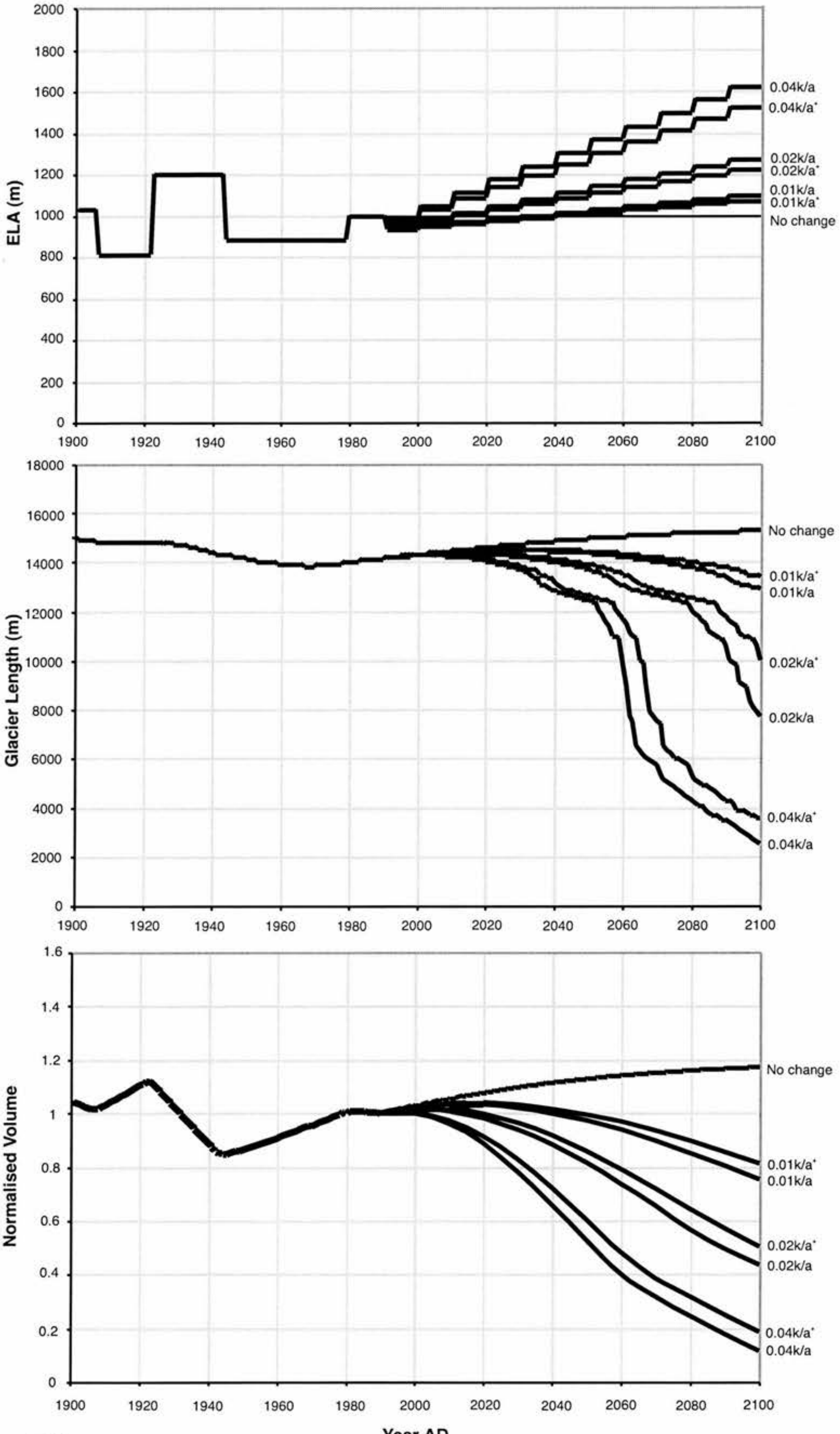


Figure 5.1

Response of Sólheimajökull to the climatic warming scenarios specified in the EISMINT experiment. The climatic warming scenarios are 0.01, 0.02 and 0.04 K of warming per year (0.01Ka^{-1}, 0.02 Ka^{-1}, and 0.004 K^{-1}). The simulations, including a 10% increase in precipitation for each 1 K increase in temperature, are also shown (0.01Ka^{-1}+, 0.02K^{-1}+ and 0.04K^{-1}+) along with a scenario where conditions remain similar to the mean climate from 1960-1990 (no change). Projected changes in ELA, glacier length and normalized glacier volume are shown from AD 1900 to 2100. The predicted change in glacier volume and length is large when compared to changes that have occurred during the 20th Century.

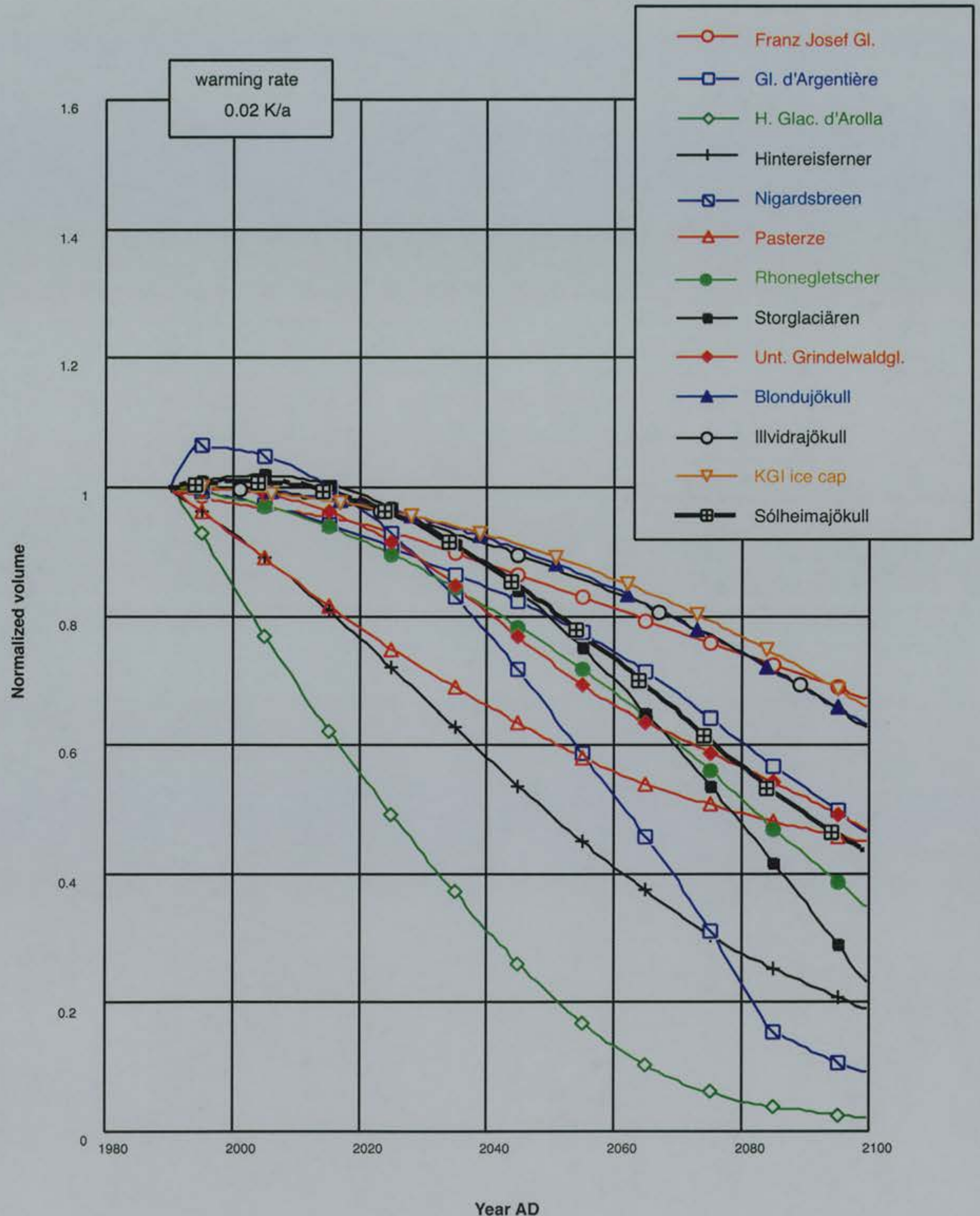


Figure 5.2

A comparison between the response of Sólheimajökull and 12 other glaciers worldwide to the 0.02 K/year EISMINT warming scenario. The predicted response of Sólheimajökull is similar to that of large glaciers in the European Alps. Small glaciers, especially those already retreating, undergo the largest changes. The ice caps undergo the smallest changes. The response of Nigardsbreen is very large. This is because the accumulation zone of Nigardsbreen is flat and lies at an altitude close to the present ELA.

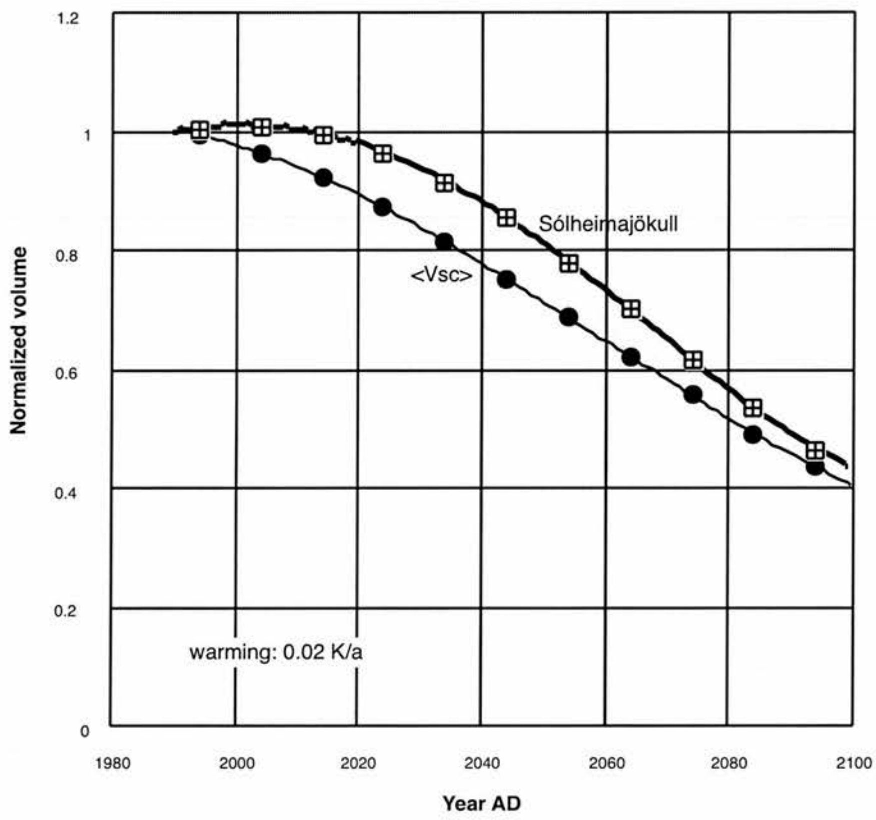


Figure 5.3

A comparison between the response of Sólheimajökull and the scaled response of 12 glaciers worldwide. <Vsc> refers to a scaling where each glacier in the EISMINT study has equal weight. Sólheimajökull has a faster rate of volume reduction compared to the mean volume reduction rate for the 12 glaciers in the study, but it takes Sólheimajökull longer to begin its volume reduction because it is initially advancing in 1990. This means that Sólheimajökull undergoes a very similar volume reduction to the mean of the 12 glaciers in the EISMINT study.

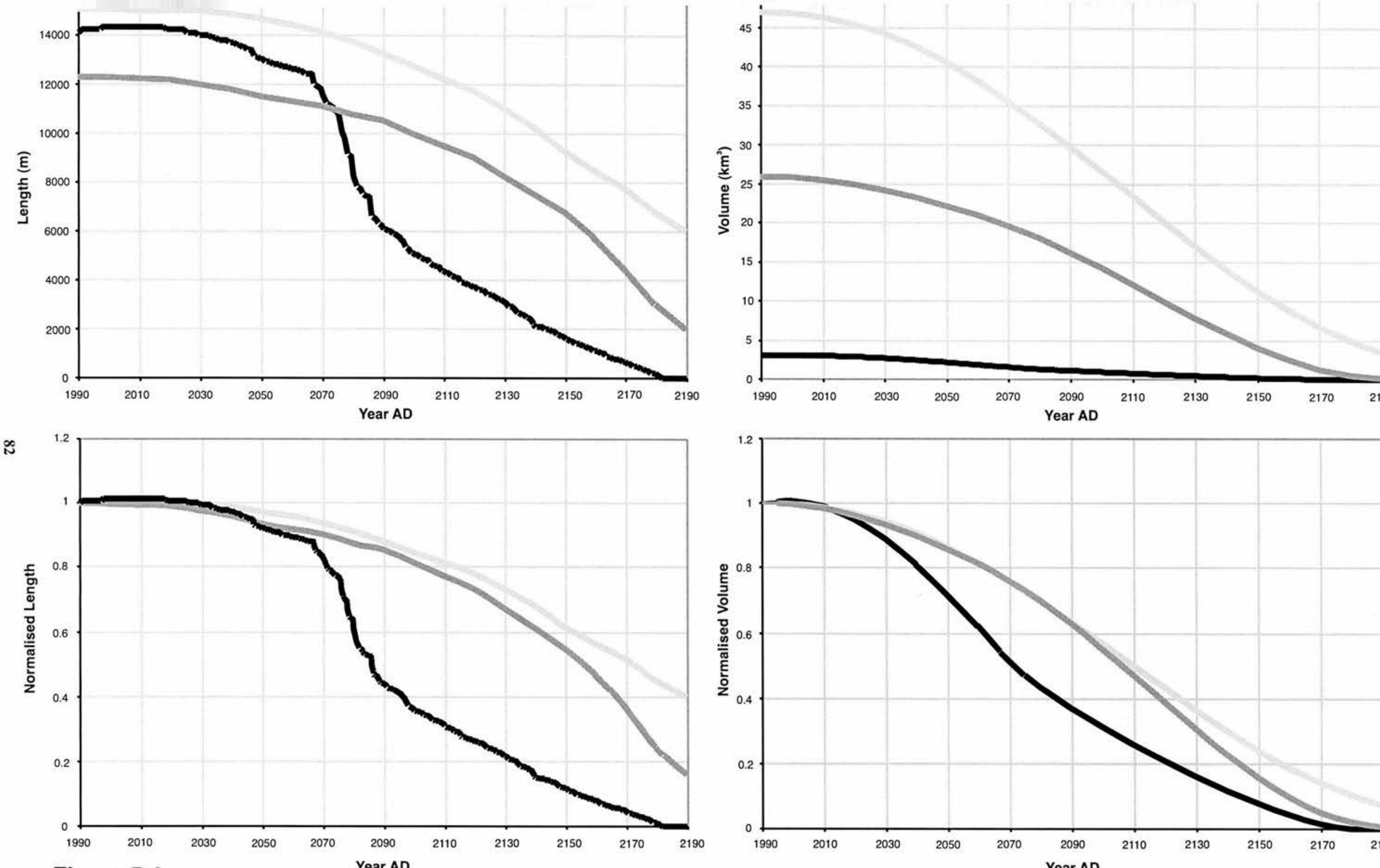


Figure 5.4

A comparison between the projected changes in length and volume of Sólheimajökull and two outlet glaciers of Hofsjökull (Blöndujökull and Illvidajökull) to the Iceland-specific climatic warming scenario. Changes in actual and normalised length and volume are shown. Sólheimajökull displays more dramatic changes in length, normalised length and normalised volume, but Hofsjokull undergoes much larger changes in actual volume. This is because the outlets of Hofsjökull cover a greater surface area than Sólheimajökull.

Chapter 6: Climate, Topography and the Holocene fluctuations of Icelandic Glaciers

Introduction

The aim of this chapter is to develop a classification of Icelandic glaciers in terms of their response to climatic change. By drawing on the findings of previous chapters, we can examine the important factors governing the response of Icelandic glaciers to climatic change. Climatic and topographic conditions can be linked to variability in mass-balance sensitivity, changes in glacial extent, response time, and moraine formation. This is achieved in the first part of the chapter. If we understand these relationships then predictions can be made for all Icelandic glaciers. In the second part of this chapter, a classification scheme is developed which incorporates these factors. The classification scheme is then applied to Icelandic glaciers so that predictions can be made regarding the response of common glacier types to climatic change.

In the final part of this chapter, the classification scheme is used to gain insight into the spatial and temporal variability of Holocene glacier fluctuations in Iceland. The classification scheme is also used to test the two hypotheses constructed in Chapter 1 regarding the Holocene fluctuation of Icelandic glaciers. The results are used to answer three research questions:

- Can spatial variability in Holocene glacier fluctuations can be explained by local differences in climatic and topographic setting in Iceland?
- Is Sólheimajökull a good choice for climatic studies?
- Can other glaciers be identified where complete Holocene moraine records are present?

Key Relationships: Glacier Fluctuations and Climatic Changes in Iceland

Climatic gradients and mass-balance sensitivity

The aim of this section is to quantify the influence of climatic gradients on mass-balance sensitivity in Iceland by drawing on the results of Chapter 3 and Chapter 5. In Chapter 5 we saw that Sólheimajökull responded to a Nordic climatic change scenario by undergoing larger changes in volume and length than Hofsjökull in central Iceland. Part of the reason for this larger response to climatic change is that Sólheimajökull has a larger mass-balance sensitivity than Hofsjökull. This means that Sólheimajökull experiences a larger change in mean specific balance than Hofsjökull for the same change in mean annual temperature. The difference in mass-balance sensitivity between Sólheimajökull and Hofsjökull is a reflection the of precipitation gradient that exists across the island (Chapter 1). Hofsjökull has a smaller mass-balance sensitivity than Sólheimajökull because it lies inland where mean annual precipitation is lower than on the south coast. The lower mean annual precipitation at Hofsjökull is largely responsible for the glacier terminating at a higher altitude (c.700 m) than Sólheimajökull (c. 100 m). The higher altitude of the ablation zone on Hofsjökull means that the mass-balance is only sensitive to changes in summer temperature, while Sólheimajökull will respond to temperature changes all year around (Chapter 3).

Because the difference in mass-balance sensitivity between Sólheimajökull and Hofsjökull can be related to the precipitation gradient in Iceland, it is possible to make predictions about the mass-balance sensitivity of other Icelandic glaciers, based on their local precipitation regime. One way to test the applicability of this finding is to use an empirical relation derived by Oerlemans and Fortuin (1992) to calculate the mass-balance sensitivity of a range of Icelandic glaciers located in areas with different amounts of mean annual precipitation. The equation derived by Oerlemans and Fortuin relates mass-balance sensitivity (S_T) to mean annual precipitation (P_{ann}) with a logarithmic function.

$$S_T = 0.512+0.662 \log (P_{ann}) \tag{6.1}$$

This equation is derived from a line of best fit in a plot of mass-balance sensitivity verses mean annual precipitation. The data points are derived from the results of a mass-balance modelling study of 12 glaciers where good mass-balance observations exist and glacier hypsometry is known (Oerlemans and Fortuin, 1992). The glaciers are located in areas of different mean annual precipitation worldwide. Equation 6.1 shows that the mass-balance sensitivity depends on the continentality of the glacier. This is in agreement with the explanation of the difference in mass-balance sensitivity demonstrated between Sólheimajökull and Hofsjökull in this study.

An advantage of Equation 6.1 is that it can be applied to any glacier as long as the local precipitation is known. Nonetheless, estimating precipitation is a non-trivial matter, especially in mountain areas where enormous local variability in precipitation occurs (Barry, 1981). By their nature, glaciers tend to be located in high precipitation, areas whereas weather stations are generally located in inhabited areas where precipitation is lower. An example is the mountain ice cap Oerafajökull, which is predicted to receive > 4 myr^{-1} of precipitation according to a precipitation map derived from weather station data (Figure 1.3), but recent mass-balance measurements indicate 6-8 myr^{-1} (Björnsson *et al.*, 1998). To reduce this potential source of error in estimating mass-balance sensitivity, winter mass-balance measurements can be used to adjust the precipitation figures where available.

With this method and Equation 6.1, the mass-balance sensitivity of several Icelandic glaciers is calculated (Table 6.1). The glaciers in Table 6.1 are chosen because they represent the climatic extremes in Iceland and capture the influence of the precipitation gradient on the mass-balance sensitivity of a range of glaciers. Precipitation for each glacier is estimated from climate station data (Figure 1.3) (Eythorsson and Sigtryggsson, 1971). The precipitation estimates are also checked against winter mass-balance measurements from high altitude accumulation areas where possible, with data supplied by O. Sigurdsson (written comm., 1998).

Table 6.1: Mass-balance sensitivity in Iceland.

Precipitation is estimated for different regions in Iceland and the mass-balance sensitivity is calculated with Equation 6.1. Extreme values of precipitation are chosen so that the climatic gradient is well captured. Drangajökull in northwest Iceland is excluded because very little is known about precipitation and mass balance in this area.

Glacier	Precipitation (estimated from map) (myr^{-1})	Winter balance (averaged over several years) (myr^{-1})	Precipitation value used (myr^{-1})	Mass-balance sensitivity ($myr^{-1}K^{-1}$)
S. Vatnajökull (Oerafajökull).	>4	6-8 Björnsson *et al.*, (1998).	8	1.1
N. Vatnajökull (Bruarjökull).	2-3	2-2.5 Björnsson *et al.*, (1998).	2.5	0.8
South Myrdalsjökull (Sólheimajökull).	>4	Not measured.	5	1.0
N. Myrdalsjökull.	3 c. 2.5 at closest climate station.	Not measured.	3	0.8
N. Hofsjökull (Satujökull).	2-3	c. 2 O. Sigurdsson pers. comm. (1998).	2	0.7
Trollaskagi (B).	1.6-2	1.5 Björnsson, (1972).	1.5	0.6

Table 6.1 confirms that the mass balance of glaciers located on the south coast of Iceland is more sensitive to temperature change than the mass-balance of interior ice caps such as Hofsjökull. Perhaps unexpectedly, Equation 6.1 calculates that most Icelandic glaciers (excluding those located on the south coast of Iceland) have a broadly similar sensitivity. The implication of Table 6.1 is that we should expect Icelandic glaciers on the south coast to undergo larger changes mass balance following a uniform climatic change than glaciers located in central and northern Iceland. In the absence of glacio-dynamic effects associated with local differences in topography, we would expect larger volume changes to occur for glaciers located in southern Iceland in response to a temperature change. Furthermore, Table 6.1 illustrates that outlet glaciers on the lee-side of large ice caps are expected to undergo smaller changes in mass-balance than the more maritime margins in response to a temperature change.

Estimating the mass-balance sensitivity of glaciers using Equation 6.1 also involves some uncertainty relating to differences in geometry between individual glaciers. This is because Equation 6.1 implicitly includes the assumption that the topography underlying glaciers is similar from location to location. For example in Table 6.1, the mass-balance sensitivity of southern Myrdalsjökull is calculated as 1.0 $myr^{-1}K^{-1}$.

In reality, southern Myrdalsjökull consists of an ice cap margin terminating at 700-800 m above sea level, and the outlet glacier Sólheimajökull which terminates at 100 m above sea level. This means that the mass-balance sensitivity of Sólheimajökull is expected to be larger than the mass-balance sensitivity of the margin of Myrdalsjökull terminating at higher altitude. This is confirmed in Chapter 2, where the mass-balance sensitivity of Sólheimajökull is calculated as 1.4 $myr^{-1}K^{-1}$ with the mass-balance model. Similar local variability in the mass-balance sensitivity related to topography is also expected in other parts of Iceland, and this variability is not evident in Table 6.1. This implies that the relationship between glacier fluctuations and climatic changes in Iceland cannot be fully understood without considering topography in more detail.

Topography and the scale of glacier response

So far it has been shown that there is a difference in mass-balance sensitivity between the south coast and central Iceland, but before conclusions can be made regarding past fluctuations in glacier length, the influence of topography on glacier dynamics needs to be considered further. In Chapter 3 it was shown that the topography of the glacier trough influenced the fluctuations of Sólheimajökull by locally increasing or decreasing the sensitivity of the glacier front to climatic change. In Chapter 4 the importance of equilibrium was illustrated; it is possible for the glacier to reach a different extent depending on how long a particular climatic period is maintained. In Chapter 5 it was shown that changes in length at Sólheimajökull were much greater than changes in the length of Hofsjökull for a uniform climatic change. The purpose of this section is to consider the influence of topography on glacier sensitivity by using these findings to generalise about how topography influences the sensitivity of all Icelandic glaciers.

The interactions between topography and glacier extent examined for Sólheimajökull have wider validity in Iceland because many other Icelandic glaciers rest on mountains with a similar form. This is because the topography of Iceland is relatively uniform due to its volcanic formation. It is common for glaciers to form on dissected plateaux and descend from mountain ice caps to sandur plains. In northern Iceland, cirque glaciers form in the valley heads of dissected plateaux. We would expect many of the features of topographic control on glacier sensitivity shown in Chapter 3 to play a role with these glaciers. For example:

- The terminus of an outlet valley glacier should undergo large changes in extent in response to climatic change if the valley is of uniform width. This is a geometrical effect, where a small change in glacier volume on the ice cap leads to a large change in glacier length. In cases where the valley is not of uniform width, the glacier will undergo smaller changes in glacier length at points of valley widening.
- For glaciers presently terminating at the mouth of a valley opening onto a sandur plain, we should expect the glacier terminus to be insensitive to further advance.
- Broad ice cap margins will undergo small changes in extent in response to climatic change, even if considerable changes in ice cap volume occur. This behavior was illustrated in the case of Hofsjökull in Chapter 5.

- Cirque glaciers will generally fluctuate by small amounts. However some cirque glaciers may be more prone to advance if they flow into a valley with a small topographic slope. In the case of Sólheimajökull, the growth of the glacier is controlled by a positive feedback loop where the glacier is stable at a length of 2 km, but advances from a length of 5 km to 11 km without a further change in climate. The large increase in glacier length is due to the elevation mass-balance feedback where the glacier grows much larger once it begins to increase in thickness. The elevation-mass-balance feedback is more important when bed slope is small (Oerlemans, 1989).

The relationship between response time, climate and topography

If the response of glaciers to a climatic change is to be predicted for a region, the possible variations in response time between different glaciers needs to be understood. This is because the response time of a glacier may largely determine whether or not a glacier responds to a climatic change. In cases where the response time is much longer than the length of a climatic change, the high frequency climatic changes may not be evident in the frontal variations of a glacier (Hubbard, 1997). In contrast, a nearby glacier with a shorter response time may experience high frequency frontal fluctuations in response to the same climatic change.

In Chapter 3 and Chapter 5, numerical modelling experiments show that the response time of Sólheimajökull is shorter than the response time of Hofsjökull. The aim of this section is to identify the topographic and climatic characteristics responsible for this difference in response time between the two glaciers, so that the response time of the remaining Icelandic glaciers can be estimated.

Johannesson *et al.* (1989) shows that the response time of a glacier can be estimated by the equation:

$$V(t) = h_{0max}/(-b_T) \qquad (6.2)$$

Where h_{0max} is a scale for the (maximum) thickness of the glacier and $(-b_T)$ is a scale for the mass balance near the terminus. Equation 6.2 implies that the response time is directly proportional to the ice thickness and indirectly proportional to the mass-balance gradient at the snout of a glacier. The advantage of this equation is that the response time can be calculated without detailed numerical experiments such as those performed in Chapter 3.

With Equation 6.2 and a few simplifying glaciological assumptions, the response time of Icelandic glaciers can be linked to their climatic and topographic setting. Mass-balance at the snout of a glacier is mostly related to elevation. Glaciers extending to lower elevations experience large net ablation at the snout, because in the absence of an inversion, the free air temperature is higher at lower elevations. This means that melting rates are higher (through an increase in turbulent heat flux and incoming long-wave radiation) and the melt season is longer. Both factors result in a more negative mass-balance at the snout.

The altitude to which a glacier extends depends on the amount of local precipitation, and the degree of topographic dissection. Glaciers in maritime climate zones that are well channelled by topography extend to the lowest elevations. On the other hand, broad ice cap lobes located in drier climatic zones tend to terminate at higher elevation. As a result, we would expect the mass-balance gradients of outlet glaciers descending toward the south coast of Iceland to be much larger than the mass-balance gradients of ice cap margins in central Iceland.

The thickness of a glacier depends on the slope of the sub-glacial topography, where glaciers on small slopes reach greater ice thickness. This is a simple relationship evident in the basal shear stress equation (Equation 2.7) where for a specified shear stress, ice thickness is inversely proportional to the bed slope.

To summarise, topographically confined outlet glaciers located on steep mountain slopes Iceland will have shorter response times than thick ice caps located on gentle slopes in the dry interior of southern Iceland. This explains why the response time of Sólheimajökull to climatic change is shorter than the response time of Hofsjökull. This reasoning can be used to qualitatively estimate response times of Icelandic glaciers where basic information is available on ice thickness and the terminus altitude.

Topography and moraine formation

Before we can classify glaciers according to their response to climatic change, some generalisations need to be made regarding the role of topography on moraine formation. This is because the classification scheme is ultimately used to enhance our understanding of the spatial pattern of Holocene glacier fluctuations in Iceland as is evident in the moraine record.

The process of moraine formation is complicated and it is not possible to identify all the factors ultimately responsible for whether or not a moraine forms in front of a glacier, and the form that a moraine will take if deposited. The natural variability in glacier size, shape, thermal regime, underlying topography and sediment sources combine to produce huge spatial variation in the amount, type and trajectories of sediment in transport and form of resulting moraines (Kirkbride, 1995). Nonetheless, some generalisations can be made as many of the processes mentioned above are influenced by the sub-glacial topography and topography of the glacial terminal environment. Furthermore, in Iceland some variables are similar for all glaciers. For example, Icelandic glaciers are temperate in nature and generally have high ice fluxes (Björnsson, 1979), the underlying substrate is highly erodable and the majority of glaciers carry most of their sediment load in the basal transport system. Cirque glacier are slightly different in that they have a significant supraglacial as well as subglacial debris source. In Iceland generally, the system is relatively simple and it can be assumed that moraines will form in front of glaciers if the glacier front remains stable and if the debris is not flushed out by meltwater. This concept is now treated in more detail.

Sub-glacial meltwater streams and proglacial streams can limit the formation of moraines in two ways. Aggressive sub-glacial meltwater streams can preferentially incorporate debris from basal ice layers into meltwater streams which then evacuate debris through meltwater outlets. Such debris will not contribute toward moraine formation. For example, Boudhusbreen in Norway and Breidamerkurjökull in Iceland evacuate up to 95% of their debris via meltwater streams (Hooke, 1989, Björnsson, 1996). This is more likely to occur under glaciers located on flat lowlands, where ice surface slopes are small and ice flow is not strongly confined by topography. In such cases, meltwater conduits are prone to lateral migration as conduits respond to minor changes in the ice surfaces slope (Shreve, 1972).

In contrast, confined valley glaciers tend to have stable sub-glacial meltwater streams and thus more basal debris. It is typical for two main conduits to develop near the lateral margins, and at the snout they combine into a single outlet. This allows debris to be freely deposited as moraines around the ice margin, except for a narrow zone at the centre of the terminus where the debris is incorporated into the meltwater stream.

In the proglacial environment, meltwater can limit the formation of moraines in two ways. On sandur plains where bedrock slopes are small, meltwater streams are free to migrate laterally. Over time, a braided stream network will erase evidence of earlier-formed moraines through erosion or burial. In such cases, former moraines are only likely to be preserved on valley sides. Moraines are most likely to survive when meltwater streams are located in bedrock channels. This will occur where stream discharge is high, as in areas of high relief and high precipitation. Another way in which former moraines can be eroded is through the impact of outburst floods (jökulhlaups). These occur when glacier ice is melted through volcanic activity (eruptions or enhanced geothermal output) or where ice-dammed lakes are drained catastrophically. Outburst floods contain great energy and have the potential to erode elaborate moraine sequences in a single event, or bury them during the deposition of vast outwash terraces. In such cases moraines are only likely to survive on valley sides. This process is described with regard to survival of Holocene moraines around the jökulhlaup-prone margins of Oerafajökull by Gudmundsson (1998) and around the Myrdalsjökull and Eyjafjallajökull by Spedding (1997).

To conclude, moraines are unlikely to survive in front of glaciers terminating on sandur plains, especially if the glacier is unconfined. On the other hand, valleys provide ideal environments for moraine deposition and preservation. These findings are integrated in the classification scheme in the following section in order to predict whether or not moraines are likely to form and be preserved in front of Icelandic glaciers. The classification scheme is intended as a tool to aid in the understanding of the spatial distribution of Holocene moraines in Iceland.

Icelandic Glaciers and their Response to Climatic Change

A classification scheme for Icelandic glaciers

Within the context of understanding the spatial and temporal variability in the Holocene moraine record in Iceland, one way forward is to develop a classification of Icelandic glaciers which illustrates the role of the climatic and topographic factors so far considered in this chapter. The aim of the classification scheme is to isolate glaciological characteristics particular to common glacier-types which in turn can be used to understand the occurrence or absence of Holocene moraine sequences in front of Icelandic glaciers. The simplest approach is to classify according to glacier size. Following the classification of Haeberli (1995), Johannesson and Sigurdsson (1998) suggested that Icelandic glaciers fall into three categories:

- The smallest, somewhat static, low shear stress glaciers (cirque glaciers) reflect yearly changes in climate and mass-balance almost without delay.
- Larger, dynamic high shear stress glaciers (outlet glaciers) react dynamically to decadal variations in climatic and mass-balance forcing with an enhanced amplitude after a delay of several years.
- The largest outlet valley glaciers provide strong and efficient smoothed signals of secular trends with a delay of several decades.

An advantage of this scheme is that it establishes a sense of the amplitude and timing of the response which increases with increasing glacier size. The main limitation is that the classification scheme does not take into account factors such as mass-balance sensitivity, influence of topography on glacial extent or the role of topography in moraine formation.

One improvement on this scheme is to consider the climatic and topographic setting of individual glaciers. In Figure 6.1, examples of the new glacier categories are illustrated. Glacier size is linked to the three glacier types; the largest glaciers terminate on sandur plains, at the base of the mountain. On a slightly smaller scale are glaciers terminating on mountain slopes or within valleys. The smallest glaciers are the cirque glaciers. These three categories are equivalent to the subdivision of Haeberli (1995). In addition, the plan form and bed slope are considered. An 'unconfined glacier' descends from an ice cap and glacier area increases with decreasing elevation. On the other hand, confined glaciers are channeled by a valley in their lower reaches, and their accumulation zone is generally well defined and much wider than their terminal reaches. Sólheimajökull and the outlet glaciers of Oerafajökull are considered good examples of the latter.

The above scheme is used to classify Icelandic glaciers in Table 6.2. The aim of the table is to illustrate the three important factors for each glacier type; the response time, the magnitude of length response and the potential to form moraine sequences. Table 6.2 anticipates that the largest and fastest glacier response to a climatic change is to be produced from outlet valley glaciers in southern Iceland. They are also the glaciers which have the greatest potential to create moraine records. In contrast, unconfined glaciers terminating on sandur plains will undergo slow, small changes in length in response to climatic changes. They are also less likely to create moraine sequences.

In Figure 6.2, a composite satellite image of Iceland is shown. It is evident from the satellite image that unconfined glaciers terminating on sandur plains are the most numerous glacier type. On the other hand, outlet valley glaciers are extremely rare. This is a reflection of the topography of Iceland, which is largely undissected due to its young geological age. The overall impression is that there are relatively few glaciers which are likely to form moraine sequences. The outlet valley glaciers and cirque glaciers offer the best possibility. Detailed predictions for each glacier type in the classification scheme are presented next.

Table 6.2: Topography and Icelandic glaciers.

The table presents the predicted response of Icelandic glaciers to climatic change. The different glacier types are ranked from most sensitive to least sensitive. The most sensitive glaciers are least numerous and vice versa.

Glacier type	Location	Examples	Abundance	Mass-balance sensitivity	Hypsometric sensitivity	Pinning point near snout?	Expected behavior	Response time	Moraine retention ability.
Valley glaciers.	South Iceland.	Sólheimajökull Lambatungnjökull.	Rare.	High.	High.	No.	Large changes in length and volume.	<50 years	Excellent.
Valley glaciers terminating on sandur.	South and central Iceland.	Oerafajökull, Mulajökull.	Moderately common.	High.	High.	Yes.	Large changes in volume. Modest fluctuations in length.	<50 years	Good on lateral ice margins. Poor on sandur plain.
Cirque glaciers.	Central and northern Iceland.	Glaciers of Trollaskagi, Flateyriskagi and Kerlingafjoll.	Common.	Low.	Medium.	No.	Modest changes in length and volume.	<50 years	Excellent.
Unconfined glaciers terminating on slopes.	All regions.	Much of Langjökull, Eyjafjallajökull.	Common.	Medium.	Low.	No.	Modest changes in length and volume.	>50 years	Fair.
Unconfined glaciers terminating on sandur.	Mostly in central Iceland.	Northern margins of Vatnajökull, Myrdalsjökull. Much of Hofsjökull.	Common.	Medium.	Low.	Yes.	Small changes in length. Modest changes in volume.	c. 100 years	Poor.

Predictions

Unconfined glaciers terminating on sandur plains

Unconfined glaciers terminating on sandur plains are common in Iceland (Figure 6.2). They form low-gradient glacier lobes that descend in widening channels to terminate on sandur plains (Figure 6.1). Braided streams and thick sequences of glacier-derived fluvial sediments dominate their terminal environments. Examples include the northern margins of Myrdalsjökull, the western and northern margins of Vatnajökull, the western margin of Langjökull and most of Hofsjökull. The glaciers are unlikely to fluctuate greatly in extent as they have widening channels and terminate at the break of slope between the mountain and the sandur plain. Both these features enhance the stability of the snout and inhibit further advance. Response times are likely to be of medium length due to the large ice thickness. The combination of high sediment budgets (Björnsson, 1996) and multiple meltwater outlets resulting from the low profile outlet glaciers (Shreve, 1972) favours maximum rates of sediment production and the formation of sandur plains. They are poor environments for moraine preservation.

Unconfined glaciers terminating in upland areas

Unconfined glaciers terminating in upland areas consist of broad lobes that terminate on the mountain slope (Figure 6.1). Their terminal environments are mostly characterized by bedrock and moraine ridges. Outlet streams tend to be incised into bedrock. They characterise the margins of Langjökull, Drangajökull, Eyjaflallajökull and the smaller ice caps. Unconfined glaciers terminating in upland areas typically have widening channels which makes them relatively insensitive to changes in mass balance. They also tend to respond slowly to climatic change as they are located at higher altitudes where ablation rates are lower. Moraine deposition is less influenced by meltwater, and due to their insensitive nature, there is potential for deposition of moraine ramparts. Unfortunately, moraines are likely to be reduced by periglacial processes over time and they are difficult to date with tephrochronology because deep soil profiles are less common than in valleys.

Outlet valley glaciers terminating on sandur

Valley glaciers terminating on sandur plains are the most conspicuous glaciers in Iceland, descending over large altitudinal gradients from high plateau icefields and from ice-filled calderas, through valleys to sandur plains (Figure 6.1). Their terminal environments are dominated by sandur, but moraines and ice marginal lakes may also be present. They occur at two scales in Iceland. Smaller examples include Gigjökull and Svinafellsjökull. Larger examples include Breidamerkurjökull, Mulajökull, Skeidararjökull and Hofdabrekkujökull. It is common for these glaciers to have an extremely favourable hypsometry. However, the scale of response may be reduced because the glacier can become pinned at the sandur plain/mountain slope boundary during advance. Response times are likely to be short as the glaciers descend to low altitudes. Moraine building potential is high on the lateral margins and a superimposed moraine may form at the mountain/sandur plain boundary, perhaps if an overdeepening is present.

Outlet valley glaciers

Outlet valley glaciers are confined to the coastal ice caps Myrdalsjökull and Vatnajökull (Figure 6.2). Sólheimajökull and Lambatungnajökull are the best examples. In the case of Sólheimajökull, the glacier flows through the dissected flank of a large volcano. At Lambatungnajökull, the glacier flows into an older geological area. The terminal environments are similar to those of alpine glaciers, and one or two rivers flowing along the valley centre drain meltwater. Outlet valley glaciers have ideal configurations for exhibiting a large response to a change in mass balance. Furthermore, because they terminate at low altitudes, they are likely to have short response times. Moraines are likely to spread over several kilometres in front of the snout and the valleys are excellent environments for the preservation and dating of moraine sequences.

Cirque glaciers

Cirque glaciers mostly occur in the Trollaskagi and Flateyri Peninsulas in northern Iceland (Figure 6.2). They typically cover less than 3 km^2 and tend to be located at the heads of large valleys that dissect the plateau surfaces (Figure 6.1). The glaciers usually terminate within valleys, and meltwater is often evacuated via a single meltwater stream. The fluctuations of cirque glaciers are unlikely to be inhibited by topography. They will advance and retreat but changes in extent will be small and moraine loops tend to be closely packed over a short distance. In this context where the cirque glaciers are believed to be temperate, response times are likely to be short because their ice thickness is so small. The terminal environment is excellent for the preservation and dating of moraines as soil is well preserved in the valleys and depositional sequences are relatively undisturbed by meltwater.

Understanding Spatial and Temporal Variability in the Holocene Moraine Record of Iceland

The next step is to use the classification scheme to provide insight into spatial and temporal variability of the Holocene moraine record in Iceland. The interpretation of Holocene moraine records in Iceland is at an early stage and has so far lacked this integrated approach. In some cases moraine sequences have been directly related to climatic changes (Stötter *et al.*, 1999), and in one case, the pattern of moraines has been interpreted as a response to a distinctive history of ice dynamics (Dugmore and Sugden, 1991). Moraine sequences seem to indicate two extreme types of Holocene behaviour. Until recently the widely held view was that many Icelandic glaciers reached their Holocene maximum during the Little Ice Age advances of the mid 19^{th} Century (Björnsson, 1979). In some cases there is good stratigraphic evidence to support such an inference (Thorarinsson, 1966, Sharp and Dugmore, 1985). In other cases, the application of reliable dating methods has shown that some moraine sequences indicate a maximum glacier expansion during the mid-Holocene, and several periods of glacier expansion since with the latest being the Little Ice Age advance. This latter behavior has been identified by Dugmore (1989) Gudmundsson (1998) Stötter *et al.* (1999) and Kirkbride and Dugmore (2000). Table 6.3 summarises the characteristics of glaciers and their moraines where long Holocene records have been found.

Table 6.3: Characteristics of glaciers exhibiting Holocene moraine records.

Location and characteristics of glaciers and their moraine sequences in Iceland for glaciers where Holocene moraines are known.

Glacier	Characteristics	Holocene moraines
Sólheimajökull (Dugmore, 1987, 1989).	Outlet valley glacier	Moraine loops and till sheets spread over 6 km in front of terminus.
Gigjökull (Kirkbride and Dugmore, 2000).	Outlet valley glacier terminating on sandur	Large moraine rampart near Little Ice Age limit.
Oerafajökull (Gudmundsson, 1998).	Outlet valley glacier terminating on sandur	Lateral moraines and large moraine rampart (Kviarjökull) within 1-3 km of Little Ice Age ice limits.
Kerlingarfjöll (Kirkbride and Dugmore, 2000).	Cirque glacier	Small moraines located close to the present glacier.
Trollaskagi cirque glaciers (Stötter *et al.*, 1999).	Cirque glaciers	Small moraines just outside Little Ice Age ice limits.

The classification scheme developed in this chapter supports a new hypothesis to explain the spatial pattern of Holocene glacier fluctuations in Iceland. Complete Holocene moraine records are only known from the glaciers listed in Table 6.2 which undergo large changes in glacier length in response to climatic changes. They are also found exclusively around valley and cirque glaciers, where there is good potential to preserve moraine sequences.

In light of these new findings, the moraine record at Sólheimajökull is re-examined. From the findings of this chapter, it is clear that Sólheimajökull has a climatic and topographic setting that promotes large amplitude fluctuations in glacier length in response to climatic change. In addition, because Sólheimajökull terminates within a valley it has good potential to deposit and preserve moraine sequences. These factors are largely responsible for the glacier exhibiting the longest and most detailed moraine sequence in Iceland. Because large amplitude changes in glacial extent can be easily explained at Sólheimajökull, and because the glacier fluctuations dated by tephrochronolgy at Sólheimajökull are synchronous with other glacier fluctuations in Iceland, it seems unlikely that ice-divide migration played a role in the Holocene evolution of the glacier as argued by Dugmore and Sugden (1991). The advantage of this new theory is that it can explain the pattern of Holocene moraines at Sólheimajökull, and also give insight into wider geomorphic evidence in Iceland. These are the hallmarks of a successful scientific theory (Popper, 1984).

This new classification scheme also provides insight into the wider pattern of Holocene moraines in Iceland. The different scale of Holocene glacier fluctuations is explainable at sites where complete

Holocene moraine records are known: the largest Holocene fluctuations occurred at Sólheimajökull, where large changes in glacier length occur in response to small climatic changes. Slightly smaller glacier fluctuations occurred during the Holocene at Oerafajökull. Here, mass-balance sensitivity is high but advance is restricted by a topographic threshold at the sandur plain. Cirque glaciers exhibited the smallest Holocene fluctuations, reflecting their smaller mass-balance sensitivity. Finally, the absence of Holocene moraine sequences in the forelands of the main ice caps can be related to the topography of Iceland.

The absence of Holocene moraines predating the Little Ice Age on many Icelandic glaciers might reflect two related processes. The first is related to glacier sensitivity. Most Icelandic glaciers terminate on sandur plains, where the scale of an advance in response to a climatic change will be small. It is suggested that these glaciers probably experienced small changes in glacier length during mid-Holocene cold stages. This means that if moraines were deposited during periods of advance, they would be located close to the present glaciers. Furthermore, a second important process limits the ability of many Icelandic glaciers to form moraine sequences. The immediate glacier foreland on sandur plains is a high energy environment where it is likely that moraines (if deposited) will be subsequently eroded by aggressive meltwater streams or buried beneath fluvial outwash deposits. As these glaciers are most numerous, it is readily understandable as to why older Holocene moraines are absent from the forelands of most Icelandic glaciers.

An alternative explanation related to glacier response time may explain the lack of mid-Holocene moraines in the forelands of many Icelandic glaciers. Earlier in this chapter it is explained that steep coastal outlet glaciers in Iceland have shorter response times than the thick lobate ice cap margins in central Iceland. If climatic cooling events during the mid-Holocene only persisted for decades, as is evident in the climate of the last 300 years, it is possible that the larger ice cap lobes in Iceland did not respond. If this was the case, then moraine sequences would only be found near the margins of glaciers with short response times, such as the outlet valley glaciers and the cirque glaciers. Table 6.3 shows that mid-Holocene moraines are only found near these glacier types.

Because there are several possible explanations for the absence of complete Holocene moraine records in the forelands of Icelandic glaciers, it is impossible to determine which process is most important. This is a problem common to many geomorphological and geological studies, where it is sometimes impossible to reconstruct the precise nature of processes that have occurred in the past. In other words, to return to Figure 1.1, if a moraine record is absent (far right of diagram), it is impossible to invert the arrows. We can learn little about regional climatic changes in this case. On the other hand, in this chapter it is evident that several glaciers produce excellent moraine records by exhibiting large responses to climatic change and by advancing into valleys where moraine records can be preserved. In this case, the moraine record can be inverted with a numerical model to reveal important information about climatic change. The moraine record of Sólheimajökull is one such example, and in the following chapter the significance of climatic changes are discussed.

A final potential use for the classification scheme is that it can be used to help guide future searches for Holocene moraine records in Iceland. Complete moraine sequences are most likely to be found around the margins of outlet valley glaciers in southern Iceland, such Lambatungnajökull in southeastern Iceland, or in front of cirque glaciers in central and northern Iceland.

Summary

Theory suggests that we should expect regional and local differences in the response of Icelandic glaciers to climatic change. Differences in mass-balance sensitivity mean that glaciers located along the south coast of Iceland are likely to have the largest response to a temperature change, while glaciers in central and northern Iceland will have a smaller response. Topography acts as a filter, either enhancing or reducing this signal locally. In the case of valley glaciers the sensitivity to climatic change is enhanced. This is due to a mass-balance feedback, where topographic confinement allows the glaciers to descend to lower elevations where the ablation season is longer. If the bed slope is small, the sensitivity is further enhanced as increasing ice thickness leads to higher accumulation and vice versa. The dynamic response of valley glaciers also depends on the nature of their terminal environment. Glaciers terminating within valleys will be more responsive because advance and retreat is restricted by topographic barriers. On the other hand, the sensitivity of valley glaciers terminating on sandur plains will be reduced during their advance cycles.

The potential for Icelandic glaciers to deposit and preserve moraine sequences is also influenced by topography. The main factor is the degree to which meltwater and sediments are partitioned during deposition. Moraines are most likely to be preserved where meltwater is evacuated by stable outlet streams, as in valleys where single outlet streams often form. Moraines will also be present on mountain slopes where stable outlet streams quickly incise into bedrock. Sandur plains are the least favorable environment for moraine deposition and preservation. This is because debris is preferentially entrained by sub-glacial meltwater conduits within the ice, and moraines are quickly eroded or buried on sandur plains by the lateral migration of meltwater streams.

Glaciers in Iceland are generally located where the topography is young and evolving dynamically, and where valleys are absent. It is typical for Icelandic glaciers to terminate in broad lobes at the boundary between a mountain slope and sandur plain. Such glaciers undergo relatively small advances in response to climatic change and they are unlikely to preserve moraine sequences. Valley glaciers are less numerous, but they undergo large changes in glacier length in response to climatic change and have potential to deposit and preserve moraine records. Moraines are most likely to be found in the older geological zones in the forelands of valley glaciers and cirque glaciers or in younger geological areas where valleys exist.

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This study offers an alternative hypothesis to that of Dugmore and Sugden (1991) to explain the large extent of Sólheimajökull during the mid-Holocene. It also provides an explanation for the wider spatial pattern of moraines in Iceland and the absence of moraines in some areas. The overriding conclusion is that where present, moraine records reflect climatic changes directly. If a numerical model is used, a moraine record can be inverted into a climate signal.

a

b

c

d

Figure 6.1

Examples from the glacier classification scheme. **a** (Skeidarajökull) A confined glacier terminating on a sandur plain. **b** (eastern Hofsjökull) An unconfined glacier terminating on a sandur plain. **c** (Solheimajokull) The main tongue is a valley glacier while the ice cap margin at high altitude is an unconfined glacier terminating on a slope. **d** A cirque glacier on the Flateyjarskagi peninsula in northern Iceland. (Aerial photographs courtesy of O. Sigurdsson, Natural Energy Authority, Iceland).

Figure 6.2

A composite satellite image of Iceland. The ice caps and different topographic zones are clearly visible. Valley glaciers are restricted to the southern part of Myrdalsjökull and Vatnajökull, while unconfined glaciers are the most common glacier type, especially on Hofsjökull and the northern margins of Vatnajökull and Myrdalsjökull. The cirque glaciers of northern Iceland are also visible.

Chapter 7: Holocene Climatic Change

Introduction

There is a growing need to understand the nature and causes of past climatic changes. This is because natural climatic variability needs to be taken into account if we are to understand present climatic changes and accurately predict future climatic changes (Simmonds, 1997). The aim of this chapter is to use the ELA reconstruction from Chapter 4 to help identify the pattern and forcing mechanisms of Holocene climatic changes in Iceland. An attempt will be made to link the Sólheimajökull ELA record to changes in atmospheric and ocean circulation.

Iceland is an ideal location for the study of palaeo-environments, because it is situated close to the convergence of temperate and polar air and water masses (Chapter 1). However, palaeoclimatic records from Iceland have not made a large contribution to past climatic studies. The main reason is that until recently, a persistent view was that the terrestrial glacial record showed little evidence for climatic cooling events during the Holocene between c. 8000 years BP and c. 2500 years BP; indeed glaciers were believed absent (Björnsson, 1979). A more general view is that the Holocene climate experienced millennial-scale variability, with alternating periods of warming and cooling, and cool intervals recurring at 1500 to 2500 year intervals (Denton and Karlen, 1973, O'Brian *et al.*, 1995, Bond *et al.*, 1997, Bianchi and McCave, 1999). Recent geomorphological evidence from Iceland indicates that the former view is incorrect: the Holocene climate of Iceland also experienced variability (Dugmore, 1989, Gudmundsson, 1997, Stötter *et al.*, 1999). However, the dynamics of these climatic changes are not well understood. One problem is that the pattern of Holocene glacier fluctuations in Iceland shows some spatial and temporal variability. This means that the climatic interpretation of glacier fluctuations in Iceland has sometimes been difficult (Dugmore and Sugden, 1991). Nevertheless, in the previous chapter it is argued that glacier fluctuations in Iceland can be related to climatic changes directly.

A further barrier to successfully relating glacier fluctuations to climatic changes in Iceland is that the dynamics of climatic variability in Iceland are generally not well understood. The main uncertainty lies in how climatic changes in Iceland relate to wider changes in atmospheric circulation and ocean currents. There are several examples of climatic conditions in Iceland that appear to differ from global climatic conditions. The Little Ice Age is known from documentary sources to have been climatically variable in Iceland and its existence has even been questioned (Ogilvie, 1992). In another example, the global climate is known to have warmed considerably in the 20th Century (Briffa and Jones, 1993), although in Iceland there has been a cooling during the same period (Einarsson, 1991).

This chapter aims to improve our understanding of the dynamics of climatic change around Iceland by critically interpreting the ELA reconstruction from Chapter 4. The relationship between other climate proxy records and the Sólheimajökull ELA reconstruction is considered. Furthermore, knowledge from observational and modelling studies of climatic variability are drawn on to explain climatic change in

Iceland. Three time scales are considered:

- Decadal-scale changes. This includes events such as the cooling in the 1960s, and warming in the 1920s.
- Century-scale climatic changes. Examples include the medieval warming and Little Ice Age.
- Millennial-scale changes. This includes climatic cycles of 1500 year duration, first identified by Denton and Karlen (1973).

Decadal-Scale Climatic Oscillations

In Chapter 4 it is shown that the climate of the last three centuries (18^{th}, 19^{th} and 20^{th}) has been characterised by large magnitude decadal-scale shifts in temperature. The purpose of this section is to seek an explanation for these climatic changes in terms of changes in oceanic and atmospheric circulation around Iceland. In Chapter 1, it is shown that the climate of Iceland is influenced by a periodic influx of sea ice around the north and west coast. During these periods the climate of Iceland is cooler, and there are suggestions that atmospheric circulation is also different.

It is also shown in Chapter 4 that the climate of Iceland during the 20^{th} Century was characterised by large temperature changes, and that the ELA record is capable of resolving these decadal-scale temperature changes. In the Introduction it is explained that the 1960s and 1970s were a period of climatic cooling associated with a high sea ice incidence in Icelandic waters. In Figure 7.1, the Sólheimajökull ELA record for the 20^{th} Century is compared to the Koch sea ice index, which is the number of weeks per year when sea ice has a direct affect on the coast of Iceland. It is clear that the cool climates reconstructed from 1900-1920 and 1940-1985 coincide with periods when sea ice was present for large parts of the year in Iceland, while the warmer period from 1920-1940 occurred when sea ice was relatively absent from Icelandic waters. It seems likely from this analysis that the Sólheimajökull ELA record mirrors the provenance of sea ice around Iceland. Both records exhibit decadal-scale variability.

Changes in atmospheric circulation may also be responsible for the variations in ELA reconstructed at Sólheimajökull. In Figure 7.1, the Sólheimajökull ELA record is plotted against an index of the North Atlantic Oscillation (NAO), an indicator of mean atmospheric circulation. It can be seen that the NAO index registers changes at a much higher resolution than the Sólheimajökull ELA record. There is no obvious correlation between the NAO index and ELA at Sólheimajökull, which means that a clear relationship between the NAO and ELA record at Sólheimajökull is not evident. However, this does not preclude the presence of an indirect relationship. Statistical analysis of the NAO index has revealed spectral powers of 20 years (Cook *et al.*, 1998), a similar length to the decadal-scale changes registered in the Sólheimajökull ELA record and in the Koch sea ice index. Furthermore, analysis of sea-level pressure data and winter sea-level pressure data have led Mysak and Venegas (1998) to suggest that the NAO index and sea-ice conditions in the Barents and Greenland Sea are related.

Negative phases of the North Atlantic Oscillation are associated with a strong East Greenland Current, an extensive ice cover in the Greenland Sea and weak Icelandic Low. During these phases, there is a southward displacement of the zonally averaged surface westerlies, and sea ice anomalies appear in the ocean around Iceland. This results in cold conditions in the Iceland, Greenland and Barents Seas. Temperatures are colder in Iceland and east Greenland. Northern Europe is colder, while the Mediterranean is wet under the influence of the westerlies. During these phases we would expect the Sólheimajökull ELA record to indicate low temperatures.

Conversely, positive phases of the North Atlantic Oscillation are associated with a weak East Greenland Current, a low sea ice cover in the Greenland Sea and a vigorous Icelandic Low (Mysak and Venegas, 1998). Temperatures are warmer in Iceland and east Greenland. During these phases, there is a poleward displacement of the zonally averaged surface westerlies. Iceland experiences wet and warm conditions, and lies directly in the path of the storm tracks. Northern Europe experiences mild and wet conditions. During these phases we would expect the Sólheimajökull ELA record to indicate higher temperatures.

Sea-ice conditions do not change directly in response to changes in atmospheric circulation because the ocean exhibits a lagged response. Between the two extremes of atmospheric circulation evident in the North Atlantic Oscillation, sea ice propagates in a clockwise direction around the Arctic, reappearing in the Greenland Sea every 10 years (Mysak and Venegas, 1998). There is evidence to suggest that this cyclical self sustaining decadal-scale climatic oscillation of the atmosphere-sea-ice-ocean system has been in existence for some time (Mysak and Power, 1991)

One question worth considering in more detail is exactly how the abovementioned changes cause a lowering in ELA on Sólheimajökull. The Icelandic climate experiences larger changes in temperature than the mean hemispheric conditions. This is related to a temperature feedback from the oceans due to the presence or absence of sea ice in the Iceland and Greenland Seas. The temperature difference between the Atlantic and polar water masses is 2-4°C (Lamb, 1979). Sea ice usually causes the greatest lowering of temperatures during winter and spring (Eythorsson and Sigtryggsson, 1971). The mass balance of Sólheimajökull is sensitive to changes in temperature all year around, so it is expected that temperature changes in winter and spring would cause fluctuations of the ELA. During severe sea ice years when ice approaches the south coast of Iceland, the effect on ELA would be profound. This is because the mass balance of Sólheimajökull is most sensitive to changes in summer temperature.

We now look at the history of the relationship between sea-ice incidence and the Sólheimajökull ELA reconstruction. In Figure 7.2, the documentary record of sea ice incidence is compared to the Sólheimajökull ELA record for the period from 1700 to 1850. The coldest period of last 300 years according to the documentary record of Icelandic climate (Ogilvie, 1992) clearly appears in both the Sólheimajökull ELA record and the sea ice index. The warm interval known from the earlier decades of the 18th Century also registers as a period of low sea ice incidence. There is a slight mismatch between the sea ice record in the early 19th Century, but during this time the reconstructed ELA record is of lower quality because the glacier snout was fixed at a topographic threshold.

The Sólheimajökull ELA record suggests that a short-term climatic oscillation has existed in Iceland from c. 1700 to the present. It is argued here that this relates to periodic sea-ice anomalies that may have a wider relationship to the North Atlantic Oscillation. The periodicity can be constrained to some extent with the model. It must be smaller than the glacier response time (c. 40 years) because glacier equilibrium was not reached during the period from AD 1700 to the present. If the number of steps in the ELA forcing (14) used in the inverse modelling is divided by the length of the reconstruction (320 years) a mean period of 22 years is found.

The Sólheimajökull ELA record appears to indicate that the decadal-scale climatic changes around Iceland are dominated by two extreme modes (Figure 7.3). During periods of low ELA (AD 1760-1800, 1820-1860, 1900-1920, and 1940-1980), temperature conditions in Iceland were dominated by cold ice-bearing waters of the East Iceland Current. The NAO experienced mainly negative modes. This means that there was a mean southward displacement of the zonally averaged surface westerlies. As a result, polar easterly winds influenced Iceland's climate. The climate was significantly different to that of the last decade, and living conditions were difficult. Polar bears were even able to migrate to Iceland across the pack ice (Ogilvie, 1992). During periods of high ELA (AD 1700-1750, 1800-1820, 1860-1900, and 1920-1940) temperature conditions were mild in Iceland. North Atlantic Drift waters extended further north and the sea-ice margin was located hundreds of kilometres from the Icelandic coast. The NAO was dominated by the positive mode. This means that the mean position of the zonally averaged surface westerlies migrated to the north (Figure 7.3). The Icelandic Low played an important role in the climate, resulting in wet and stormy conditions. These climatic conditions have dominated Iceland since the mid-1980s. More widely, the associated northerly migration of the mean position of the zonally averaged surface westerlies that has been evident since the mid-1980s has caused a warming in northern Europe, carrying as far north as Siberia (Hartmann *et al.*, 2000).

Recently a link has been made between the North Atlantic Oscillation and the hemisphere-wide Northern Hemisphere Annular Mode (NAM), which is also called the Arctic Oscillation (Wallace, 2000). This climatic phenomena represents mean atmospheric conditions over the entire Arctic, and the North Atlantic Oscillation is a part of the cycle. The NAM is defined from a sea-level pressure field over a weighted average of grid points throughout the Northern Hemisphere. It includes a phase difference, where the North Atlantic sector of the Arctic is out of phase with the Bering Sea/Beaufort Sea sector. This means that extensive sea ice conditions and high pressure in the North Atlantic sector occurs synchronously with low pressure anomalies and a minimum sea-ice cover in the Bering Sea/Beaufort Sea sector, and vice versa (Figure 7.3).

The NAM and NAO are indistinguishable in terms of their spatial signatures and time-dependent behavior (Wallace, 2000). This means that if temperature changes in Iceland can be linked to the NAM, then hemispheric or global signal is involved. This is supported by evidence which shows that last three extreme periods of this oscillation (late 1960s, early 1980s and early 1990s) have occurred during synchronous North Atlantic Oscillation and El Nino-Southern Oscillation events. This suggests global teleconnections and a single forcing mechanism (Mysak *et al.*, 1996).

If temperature changes in Iceland can be related to hemispheric signals, then it should be possible to compare the Sólheimajökull ELA record with wider climate proxy indicators. Indicators of Arctic sea ice extent in the North Atlantic sector are especially relevant, because they represent the mean conditions of the Arctic Oscillation over decadal time-scales (Mysak and Power, 1991). As a final test of the explanation of climatic variability presented in this section, the Sólheimajökull ELA reconstruction is plotted against a record of the August sea ice extent in the Barents Sea in the region surrounding Svalbard over the period from AD 1650 to the present (Vinje, 1999) (Figure 7.4). The Barents Sea sea-ice record was reconstructed from a variety of sources mainly associated with shipping movements over the past centuries. The early part of the record was reconstructed from documentary sources in ship log books kept by whalers, sealers and fisherman. More recently, wider shipping records, aircraft observations and satellite images have been used.

Figure 7.4 shows that record of the sea-ice extent in the Barents Sea and the ELA reconstruction from Sólheimajökull are remarkably well correlated. The two records are in phase: periods of low ELA associated with cold temperatures in Iceland are associated with an increase in sea ice extent in the Barents Sea. In contrast, periods of high ELA corresponding to high temperatures in Iceland are associated with a decrease in sea ice extent in the Barents Sea. This relationship holds for the main climatic periods identified in this section. Indeed, the early part of the 18th Century is shown in both records to be similar in scale to the present warming. The late 18th Century is clearly the coldest period in recent history. Both records indicate the strong warming from AD 1920-1940. Even subtleties such as the cooling between 1905 and 1920 evident in the Sólheimajökull ELA record are evident, as are increases in extent of the Barents Sea ice edge. The correlation between these two climate proxy records may be particularly good as both are influenced by mean summer temperature conditions. This correlation strongly supports the inference that the Sólheimajökull ELA reconstruction is picking up a trend in temperature associated with changes in sea-ice extent over the entire North Atlantic sector of the Arctic.

Century-Scale Climatic Oscillations

One interesting question worth exploring is whether or not a Little Ice Age climatic event is present in the Sólheimajökull ELA reconstruction. It is widely believed that the climate of the 17th, 18th and 19th Centuries was considerably colder than the climate of the 20th Century (Grove, 1988). However, recently the assertion that the Little Ice Age was uniformly cold in the North Atlantic has been dismissed (Ogilvie, 1992, Barlow, 1997). One way to track a possible Little Ice Age signal in Iceland is to compare the ELA reconstruction at Sólheimajökull with some well known climate proxy indicators. In this case, the GISP2 accumulation record (central Greenland) and the Nansen Fjord marine sediment record (coastal east Greenland) are chosen because they have a high resolution, and cover the time period in question.

In the previous section, it became evident that changes in the climate of Iceland should be synchronous with climatic changes in Greenland. This is because during cold phases, an expansion of sea ice into the Greenland Sea would result in colder temperature in both areas. In the case of the Nansen Fjord marine sediment record, this should be clearly indicated by an influx of cold water. The relationship between the GISP2 record and the climate of Iceland is slightly more complicated. This is because the GISP2 ice core was drilled on the summit of the Greenland ice sheet, further from oceanic influences. Nonetheless, the GISP2 accumulation record may correlate with changes in ELA at Sólheimajökull because during cold phases in Iceland, an increase in sea ice in the Greenland Sea would result in a reduction in air temperatures. Lower air temperatures then reduce the moisture-holding capacity of the atmosphere. In addition, a high sea-ice cover in the precipitation source region would also decrease accumulation rates in central Greenland.

All three proxy records show evidence of climatic variability during the last 300 years (Figure 7.5). The Nansen Fjord record probably shows the clearest indication of a Little Ice Age signal between AD 1700 and 1900 when the fjord was dominated by an influx of polar water (Jennings and Weiner, 1996). The GISP2 record also shows this to be a period of cool climate, as indicated by an decrease in accumulation (Meese *et al.*, 1994). At Sólheimajökull, the period of climatic cooling is tightly constrained between AD 1760 and 1920, when there is an absence of warm decadal-scale climatic events. In agreement with Ogilvie (1992), the Sólheimajökull ELA record shows that this period was not uniformly cold. The peak of the cooling (as discussed earlier) seems to have been in the 1780s and 1790s. It can be concluded that a Little Ice Age signal is present, but it is shorter and more climatically variable than the view of the Little Ice Age as proposed by Grove (1988).

Less information is available during the years prior to 1700 AD, both in terms of the quality of ELA reconstructions and other climate proxy records. In Figure 7.5 it can be seen that a Medieval Warm Period is not resolved by the Sólheimajökull ELA record, although there is clear evidence for a Medieval warming in the GISP2 Ice Core and the Nansen Fjord record. One interesting point can be observed; the glacier advance dated to the mid 10th Century at Sólheimajökull occurred during a period believed to be consistently warm. One possible reason for this glacier advance is that a decadal-scale climatic cooling occurred during a period of overall favorable climate. An analogy can be drawn with the climate of the last few decades in Iceland which have seen the advance of Sólheimajökull. This has occurred during a century that is widely held to be the warmest in the last 1000 years (Briffa and Jones, 1993). This indicates that despite experiencing an overall warming, the climate of the 10th Century in Iceland may have been influenced by changes in sea-ice extent related to climatic processes that are still continuing at present. It would only be possible to test this hypothesis at Sólheimajökull if the glacier length record was more tightly constrained before AD 1700.

Millennial-Scale Climatic Oscillations

The broad periods of glacier expansion at Sólheimajökull at c. 5000, 3430, and from AD 570 to the present indicate global changes, as glacier expansion at these times is known to be a global phenomena (Denton and Karlen, 1973) (Figure 7.6). Using the dust and sea salt record in the GISP2 ice core, O'Brian *et al.* (1995) suggest that longer-term climatic changes such as the Little Ice Age are associated with changes in atmospheric circulation on a 1500 year time-scale. Colder periods result from an expanded polar vortex. This means that the mean position of the zonally averaged surface westerlies migrated to the south. Warmer intervals are associated with the northward movement of the cyclone tracks and calmer zonal atmospheric circulation (Figure 7.6).

Ocean currents also fluctuated over these millennial time intervals. Grain size data in marine cores south of Iceland also indicate fluctuations in ocean circulation at 1500 year intervals (Bond *et al.*, 1997, Bianchi and McCave, 1999). This data implies changes in ocean surface currents and probably sea ice extent, where cold phases are associated with a southward expansion of polar waters. Furthermore, the Nansen Fjord record shows a clear indication of a Mediaeval Warm Period and a Little Ice Age (Figure 7.5). Coarser resolution marine core data indicate that these 1500 year climatic cycles are superimposed on a longer cycle. During the course of the Holocene, the marine polar front migrated southward. This southward movement reflects cooling since the Holocene climatic optimum that occurred at around 8000 years BP in the North Atlantic (Koç *et al.*, 1993).

It appears that the Holocene atmospheric and oceanic circulation patterns in the North Atlantic over centuries to millennia have experienced two dominant modes, resembling the extremes of the North Atlantic Oscillation seen in the 20th Century (Tremblay *et al.*, 1997). If this is correct, then the climatic changes on all time scales discussed here result from similar climate dynamics. The primary forcing mechanism for the long-term changes is in dispute, but is thought to represent either changes in the solar constant (Denton and Karlen, 1973, Karlen and Kuylenstierna, 1996) or long-term changes in the abundance of volcanic aerosols in the atmosphere (Porter, 1986, Nesje and Johannesson, 1992).

The Sólheimajökull ELA record may provide insight into the interaction between short term and long term climatic changes. For example, in contrast to the variable Little Ice Age record, evidence from Sólheimajökull indicates that mid-Holocene climatic cooling may have been less variable. This is because the numerical modelling experiments in Chapter 4 show that if the climatic cooling was of similar magnitude to the climatic changes of the last few centuries (in terms of temperature or precipitation), then a cold climatic period of c. 300 years duration is required to simulate the glacier extent at this time. If the climatic cooling of the mid-Holocene was sustained for a long period as suggested here, it may indicate a weaker or absent decadal-scale climatic oscillation at that time. This is clearly speculative, but it is possible to find explanation for this apparent difference in climate proxy records.

One possible explanation is that gross oceanographic conditions were different around 5000 years BP compared to those of today. The East Greenland current was weaker and did not extent as far south, and the Icelandic current may not have been present (Koç *et al.*, 1993). Under these conditions, the mid-Holocene climate of Iceland would have been less influenced by sea ice, and decadal-scale climatic oscillations (if present) would have had a smaller effect on temperatures in Iceland. During longer-term cool phases, expansions of the polar vortex might have resulted in more intense westerlies in Iceland rather than significantly lower temperatures. A contemporary analogy could be made with southern Norway, where coastal glaciers have recently expanded in response to an increase in precipitation resulting from enhanced westerly air flow (Tvede and Laumann, 1997), while Icelandic glaciers have advanced in response to lower temperatures (Sigurdsson and Jonsson, 1995). The present distribution of ocean currents around Iceland was established at c. 3000 years BP (Koç *et al.*, 1993). This might explain why cool events in the Sólheimajökull ELA record correlate well with the rest of world up until c. 2000 years ago, but after this time the climate of Iceland has became increasingly variable and dominated by short-term climatic fluctuations.

Summary

The Sólheimajökull ELA record indicates that the climate of Iceland in the last three centuries was strongly influenced by a decadal-scale climatic oscillation that involved changes in atmospheric and oceanic circulation and critically, sea ice extent. Iceland is presently sensitive to this oscillation because of the proximity of the polar front to the north coast. This high-resolution cycle seems to be superimposed on a long-term oscillation with a wavelength of several thousand years. This millennial-scale climatic variation is responsible for the broad intervals of glacier expansion at c. 5000, 3000 and 1400 years BP. The timing of Holocene glacier fluctuations at Sólheimajökull was synchronous with global glacier trends. It is speculated that this longer-term climatic change has become increasingly less evident in Iceland as the polar front migrated southwards during the late Holocene. The most recent cold phase, the Little Ice Age, appears to have been of short duration, diminished by decadal-scale oscillations. In contrast, the mid-Holocene climate of Iceland is speculated to have been more stable because the polar front was located further to the north. In such a situation, temperature changes in Iceland would have been smaller but less variable on a decadal scale.

These findings have implications for interpreting contemporary climatic patterns. Warming in the Arctic since 1985 is associated with a decrease in sea ice extent in the Greenland and Barents Sea, and a northerly migration of the zonally averaged surface westerly winds. Although there is concern that these changes are related to increased Greenhouse gas emissions, it must be kept in mind that the Arctic experiences significant natural variability. Similar climatic changes in terms of scale and the rate of change are identified in this study during the middle half of the 18th Century. As this is prior to the Industrial Revolution, it is concluded that changes of this nature are inherent in the climate system. This means that it is difficult to distinguish between natural and human-induced climate changes.

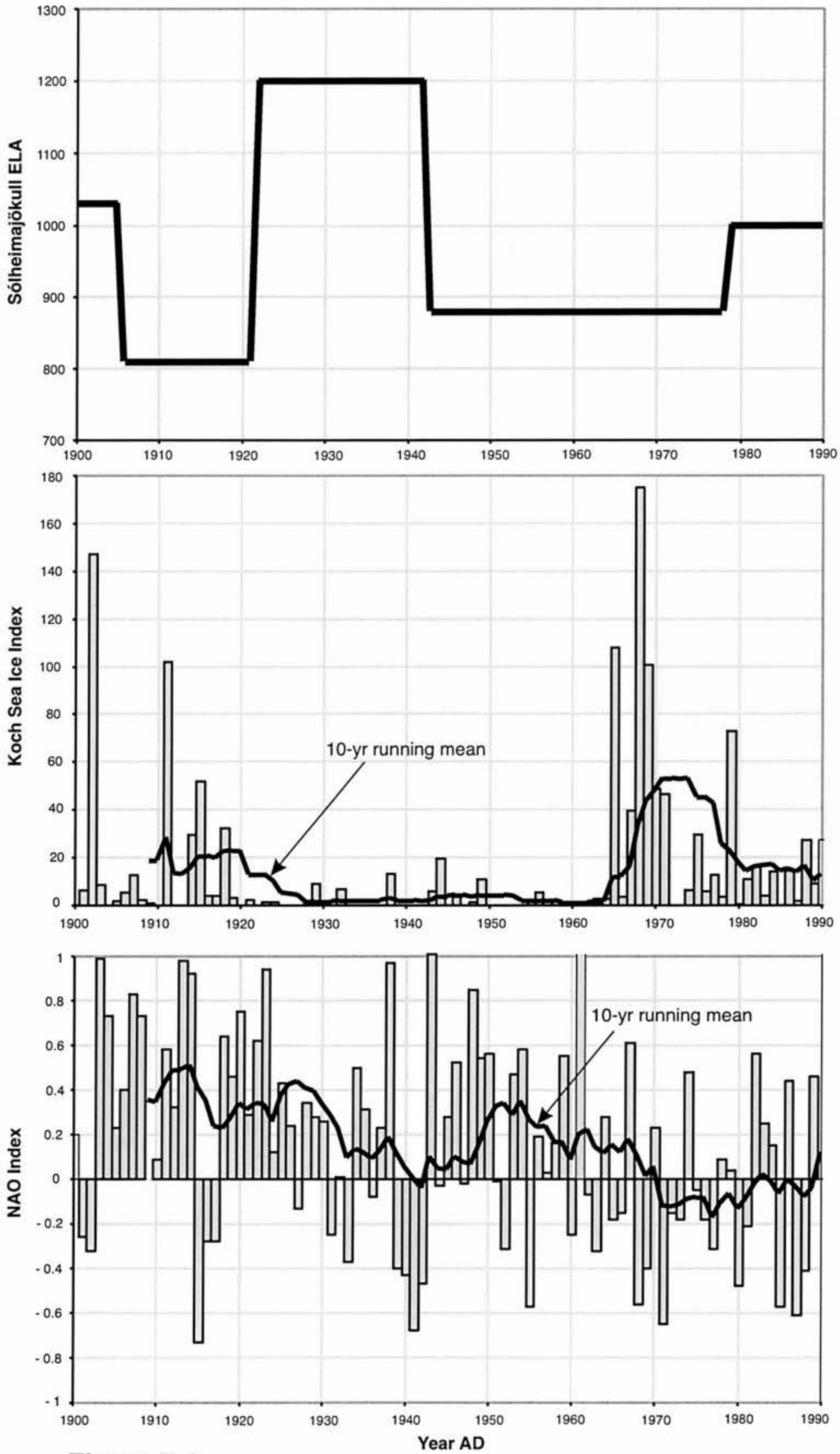

Figure 7.1

A comparison between the Sólheimajökull ELA record, the Koch sea ice index and the North Atlantic Oscillation (NAO) index (T. Jonsson written comm. 1998). There is a general agreement between the extreme states. For example the 1920s and1930s are characterised by low sea ice incidence, positive phases of the NAO and a high ELA. In contrast, the 1960s and 1970s are characterised by high sea ice incidence and negative phases of the NAO, and a low ELA.

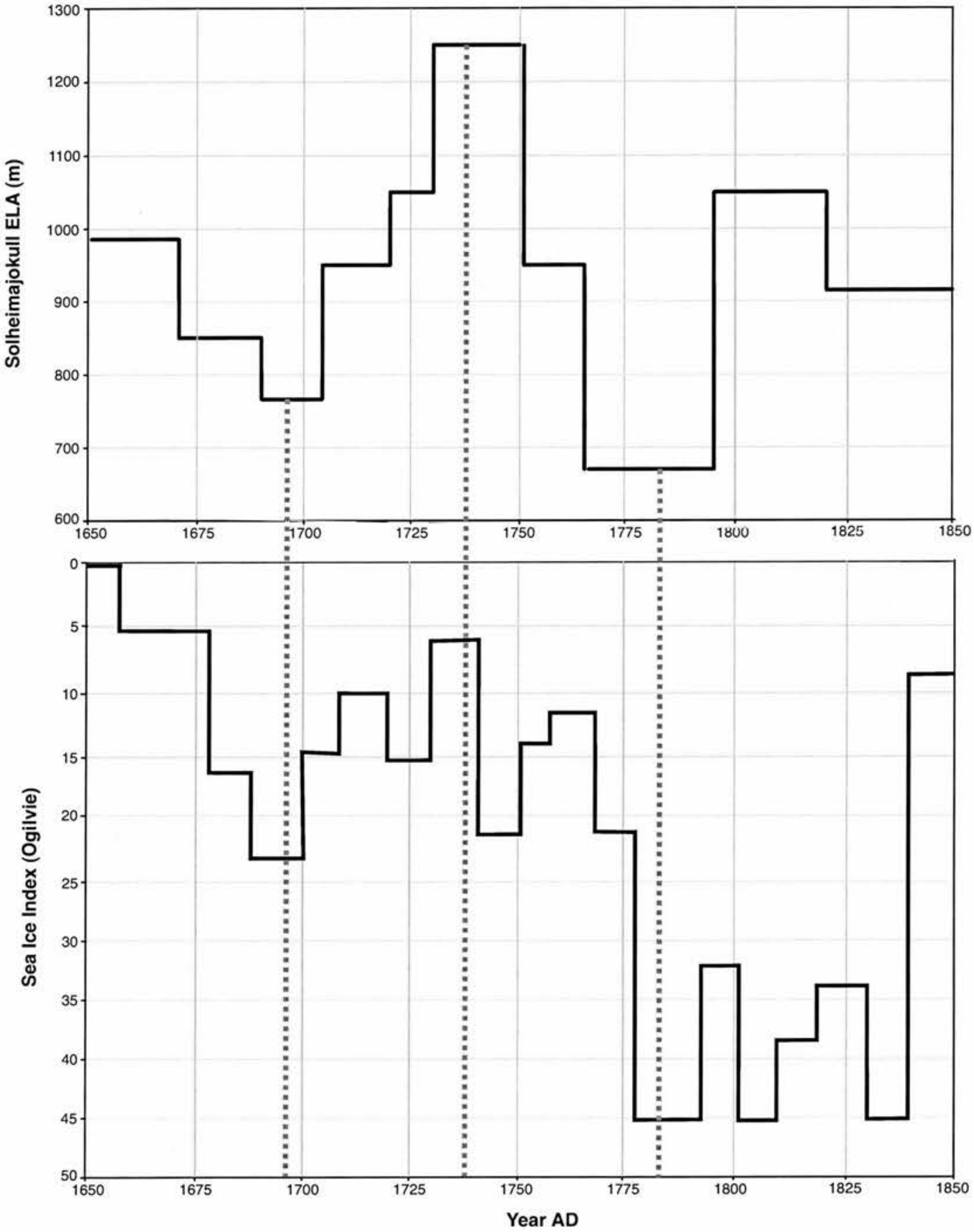

Figure 7.2

A comparison between the Sólheimajökull ELA record and a sea ice index for the period from AD 1650-1850 (Ogilvie, 1992). The years from 1700 to 1750 are warm according to the Sólheimajökull ELA record, and correspond to a period of low sea ice incidence. The last few decades of the 18th Century are the coldest of the ELA reconstruction and correspond to the most severe sea ice conditions in the last 300 years. A dotted line highlights the places where there is a good match between peaks of the two records.

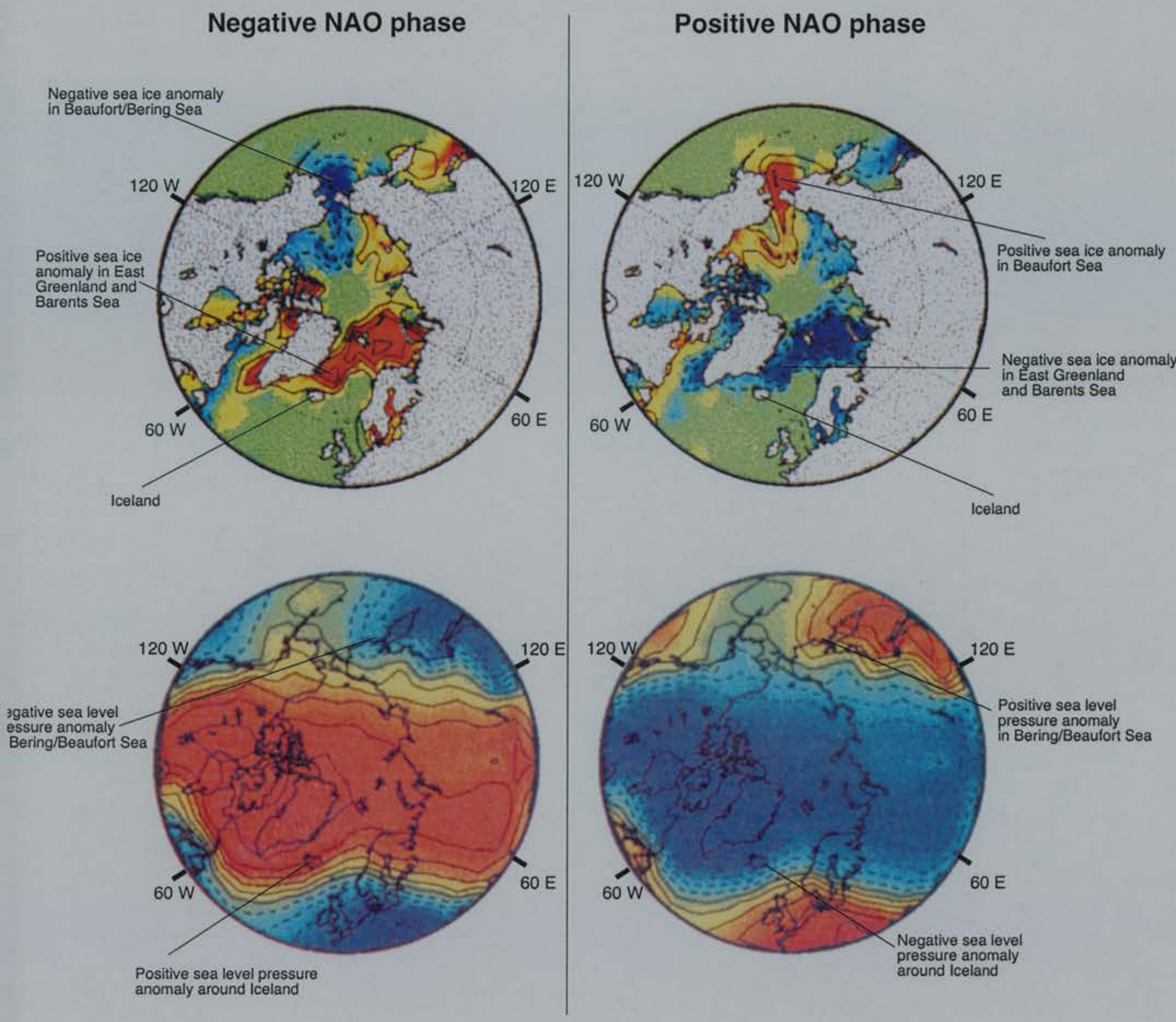

Figure 7.3

The figure shows two extremes of mean oceanic and atmospheric conditions in the Arctic over the period from 1950-1990 AD as reconstructed from pressure and sea ice concentration data (Mysak and Venegas, 1998). The upper charts are sea-ice anomaly maps. Red tones indicate a positive anomaly (more sea ice than normal) while blue tones indicate a negative anomaly (less sea ice than normal). Green tones reflect normal conditions, and the continents are white.

The lower charts are corresponding sea level pressure anomaly maps. Red tones indicate positive anomalies (higher pressure than normal) and blue tones indicate negative anomalies (lower pressure than normal).

The two extreme states reflect the positive and negative phases of the North Atlantic Oscillation. During positive phases (charts to the left), sea ice is extensive in the Greenland and Barents Sea. High pressure dominates over Iceland and the westerlies shift south toward the equator. Low pressure dominates over Britain and southern Scandinavia.
In the Bering Sea/Beauford Sea sector, an opposite phase signal is evident. Sea level pressure is low and sea ice is less extensive. During the positive phase of the North Atlantic Oscillation (charts to the right), the opposite conditions occur. Sea ice is less extensive in the Greenland/Barents Sea, and low pressure dominates over Iceland.

The climate of Iceland since AD 1700 has been dominated by switches between these two states.

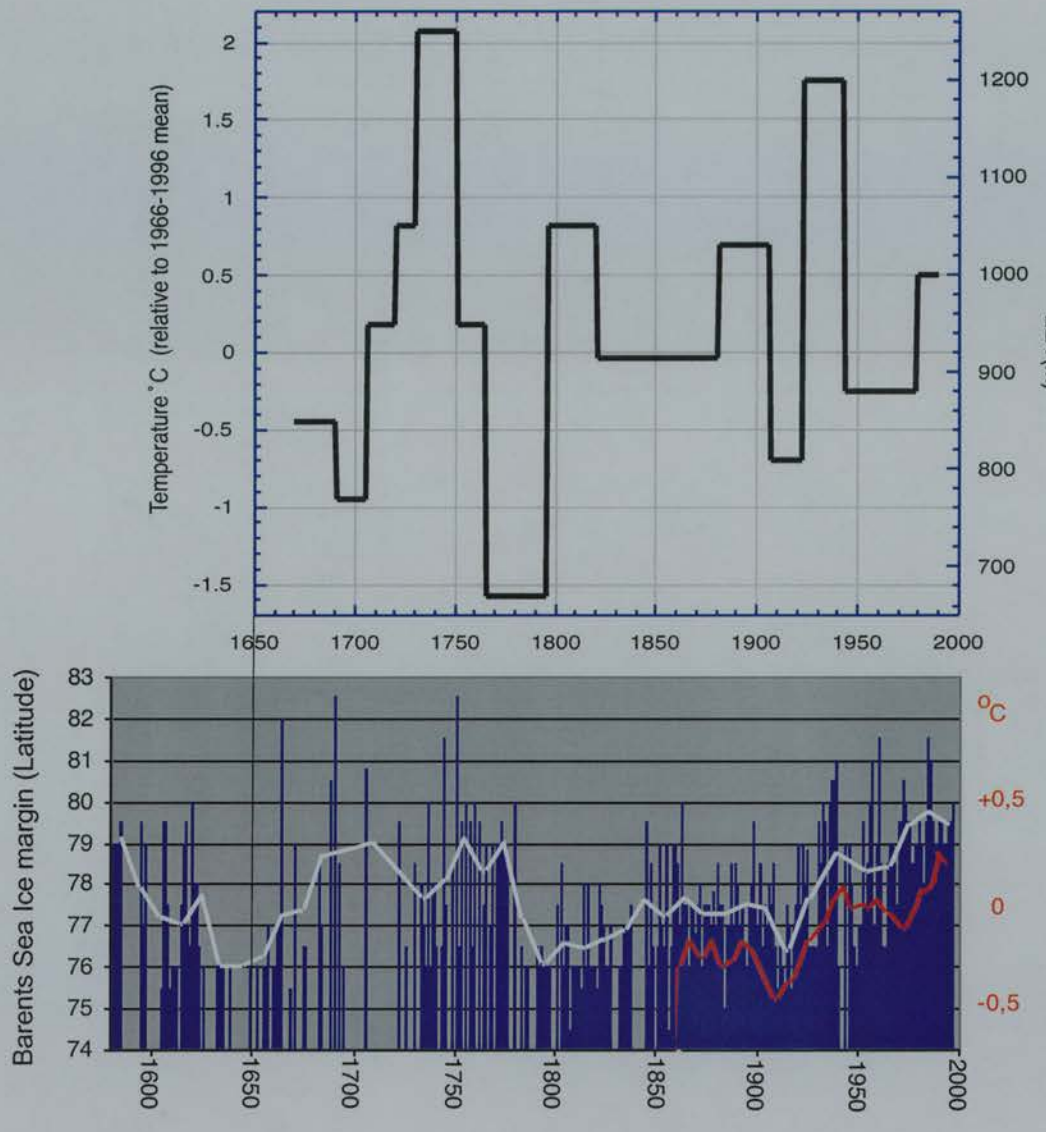

Figure 7.4

A comparison between reconstructed temperature and ELA at Sólheimajökull, and fluctuations in the sea-ice margin of the Barents Sea over the period from AD 1650 to 2000 (Vinje, 1999). The Barents Sea data represents the August ice edge position in the western Barents Sea between Svalbard and Franz Josef Land as reconstructed from records of whalers, sealers and fisherman. Each bar represents an annual ice edge position, and the white line is a smoothing with a 10-year mean. The red line is the atmospheric temperature deviation from the Northern Hemisphere mean temperature as recorded in climate stations from 1961-1990.

The Barents Sea ice edge fluctuations effectively show North-South migrations in the polar front and are correlated with the Sólheimajökull ELA reconstruction. The warming from 1700-1750 is visible in both charts and corresponds to a higher latitude sea ice edge position in the Barents Sea, comparable to that of the present day. The cooling in the 1780s and 1790s evident in the Sólheimajökull record is characterised by a large southward migration of the sea ice edge position in the Barents Sea. The cooling spike evident from 1905-1920 in Iceland is evident in another southward movement of the Barents Sea ice edge. The warming of the Arctic from 1920-1940 is clear in both records.

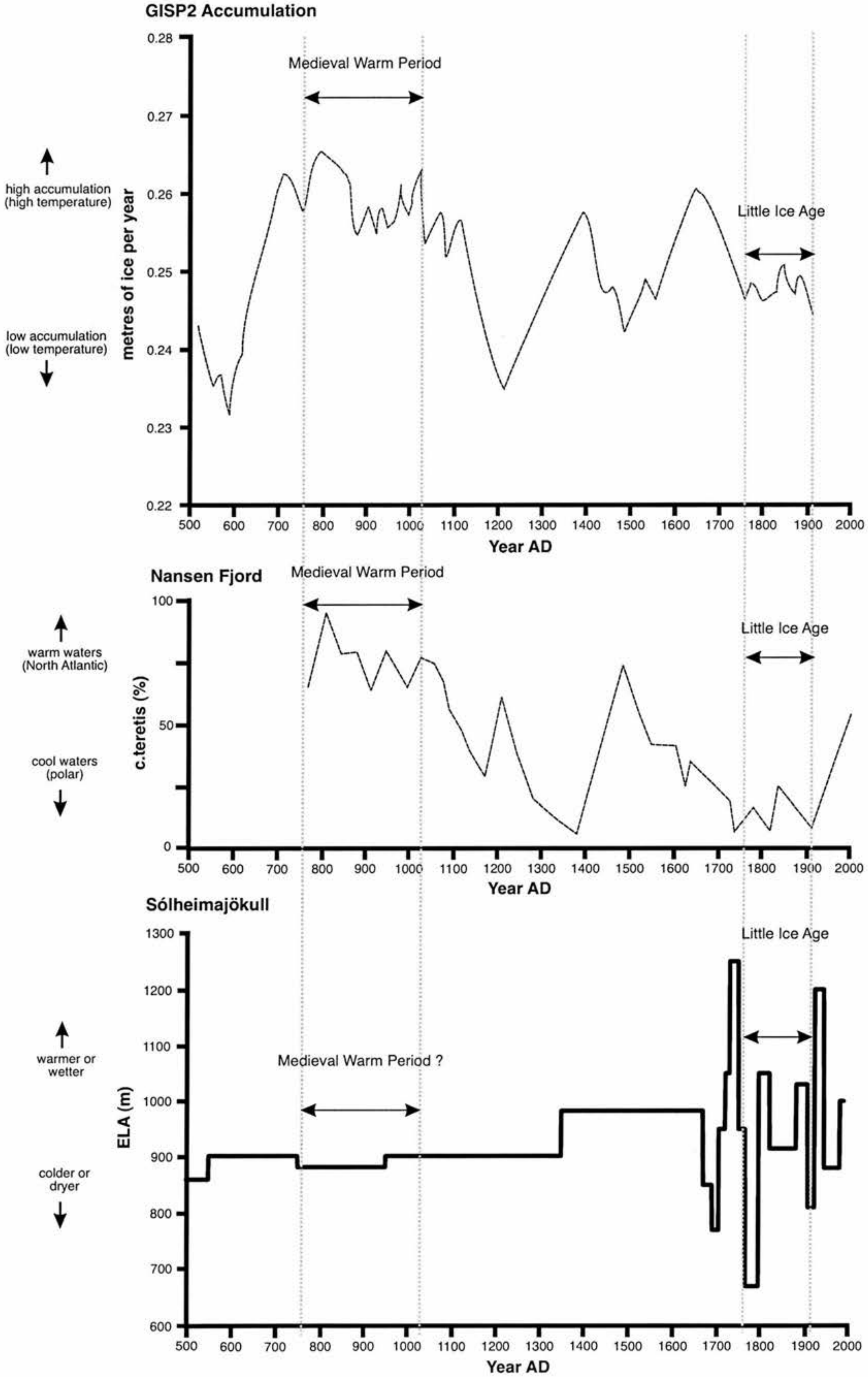

Figure 7.5

A comparison between the Sólheimajökull ELA record, the GISP2 accumulation record from central Greenland (Meese et al. 1994) and the Nansen Fjord Cassidulina terretis (formaminifera) record from a marine core in east Greenland (Jennings and Weiner, 1996). All three records show evidence of a Little Ice Age event, but a Medieval Warm period is not evident at Sólheimajökull. This may be because the ELA reconstruction is not well resolved.

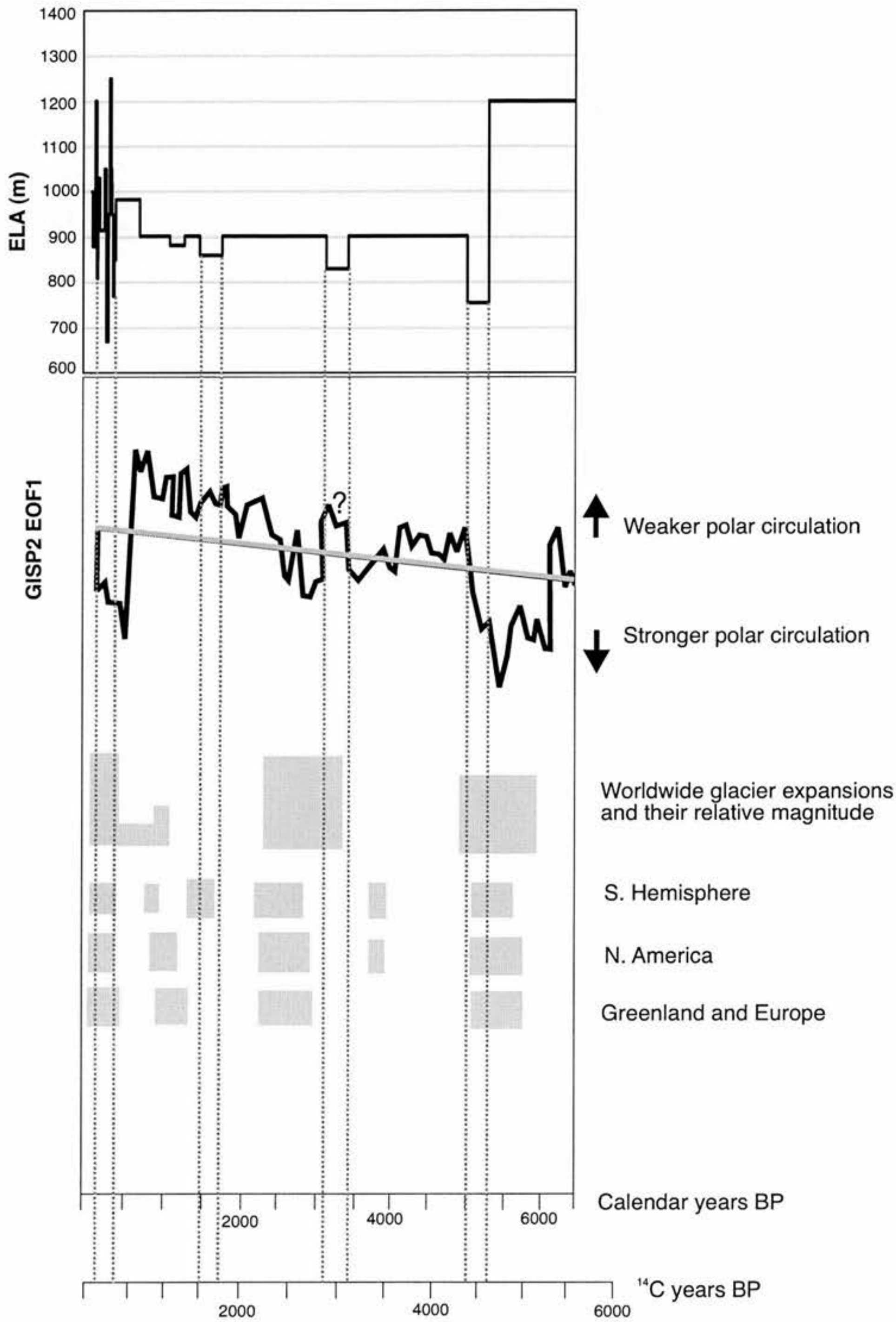

Figure 7.6

A comparison between the Sólheimajökull ELA record, the GISP2 EOF1 record and a synthesis of worldwide glacier fluctuations (O'Brian et al. 1995). There is general agreement between all three records in terms of the timing and magnitude of Holocene cool phases. This indicates that fluctuations in the ELA of Sólheimajökull reflect global climatic changes. Specifically, periods of low ELA at Sólheimajökull occur during periods of stronger polar circulation when the westerlies migrate southwards.

Chapter 8: Implications and Conclusion

Introduction

The aim of this chapter is to summarise the main findings of this thesis, and to discuss the implications for future studies. The chapter is divided in three sections. First, the hypothesis presented in Chapter 1 are revised in light of the thesis findings. Second, the contribution of the thesis to several fields of research is covered including:

- Mass-balance sensitivity and glacier response.
- Topography and glacier response.
- Inverse modelling of climatic changes.
- Predicting the response of Icelandic glaciers to climatic warming.
- The Holocene moraine record in Iceland.
- Holocene climatic changes.

Avenues for future research arising from the work are also discussed.

Revision of hypotheses

In Chapter 1, several hypotheses were presented as an aid to understanding the relationship between glacier fluctuations and climatic change in Iceland. In this section, the hypotheses are revisited: First it is explained if the hypothesis has been successfully tested within the thesis. In cases where more work is required, the original hypotheses are revised in light of the thesis findings.

Hypothesis 1

- Glaciers located on the south coast of Iceland have a larger mass-balance sensitivity than glaciers in central and northern Iceland.

In Chapter 3, the mass-balance sensitivity of Sólheimajökull was calculated at 1.4 $ma^{-1}K^{-1}$, a higher value than the mass-balance sensitivity of Hofsjökull in central Iceland (0.7-0.8 $ma^{-1}K^{-1}$) as calculated by a degree-day model by Johannesson (1997). This finding provides evidence in support of the hypothesis. In Chapter 6, an equation based on Oerlemans and Fortuin (1992) also showed that the mass-balance sensitivity of glaciers in southern Iceland is expected to be higher than the mass-balance sensitivity in central and northern Iceland, although unexpectedly, the mass-balance sensitivity in central and northern Iceland was found to be similar. This is because precipitation is significantly higher on the maritime southern margin of Iceland where orographic precipitation is intense.

These findings can only be truly tested if mass-balance measurements are undertaken on more glaciers in Iceland, especially in southern Iceland where only one glacier has recently been studied (Breidamerkurjökull) (Oerlemans *et al.,* 1999).

Hypothesis 2

- Glaciers confined in valleys will undergo a larger change in length in response to climatic change than broad ice cap lobes.

In Chapter 5, a comparison between projected length changes of Sólheimajökull and Hofsjökull in response to an atmospheric warming scenario specified for Iceland is in support of this hypothesis (Figure 5.4). It is important to remember that changes in valley glacier length are influenced by small changes in valley geometry. In Chapter 3 it was shown that glaciers stabalise at points of valley widening. It follows that this hypothesis might not hold if a glacier is located at such a topographic threshold. Generalisations are difficult as the response of individual glaciers depends on the bedrock topography.

Hypothesis 3

- Glaciers terminating on sandur plains occupy stable topographic positions and will be insensitive to further advance.

This hypothesis was successfully tested in Chapter 3 (Figure 3.3) where modelling shows that Sólheimajökull stabalises on a sandur plain beyond a length of 18 km, despite further climatic cooling. Beyond this critical length, changes in glacier length resulting from a uniform lowering of ELA are reduced by a factor of 5. It was also shown in Chapter 6 that many large outlet glaciers in Iceland terminating on sandur plains lack long Holocene moraine records. One reason may be that these glaciers have undergone small changes in glacier extent during the Holocene as a result of their topographic configuration.

Hypothesis 4

- Glacier expansion in Iceland occurs during cold periods when sea ice surrounds the Icelandic coastline.

This relationship is evident over the last 300 years, as is argued in Chapter 7 (Figures 7.1-7.5). Figure 7.4 provides strong supporting evidence for the hypothesis: Changes in the sea ice margin of the Barents Sea between AD 1650 and the present are compared to the Sólheimajökull ELA reconstruction. This shows that north-south changes in the late summer sea ice extent in the North Atlantic sector of the Arctic appears to be the most important factor in forcing glacier expansions in Iceland. There are some periods when the relationship is more complex as in the case of the ELA reconstruction at Sólheimajökull which shows a cooling in southern Iceland from 1940-1980, while sea ice became more extensive from the late 1960s until the early 1980s.

Hypothesis 5

- Climatic variability in Iceland has resulted from changes in ocean waters and atmospheric circulation on timescales of decades to centuries.

In Chapter 7 it is shown that changes in ocean waters, specifically in sea ice extent appear to be most significant in forcing glacier fluctuations on a decadal timescale. Although changes in atmospheric circulation may ultimately be responsible for the changes in ocean circulation, atmospheric circulation as evident in the NAO index has a high frequency oscillation which is difficult to interpret. In contrast, changes in sea ice extent appear to have a clear decadal signal, and have been responsible for the large temperature changes that have forced glacier fluctuations over the last three centuries. This hypothesis can be revised to read 'Climatic variability in Iceland has resulted from decadal changes in sea ice extent that may have been driven by switches between modes of the North Atlantic Oscillation.'

Hypothesis 6

- Valley glaciers located on the south coast of Iceland will undergo larger changes in length and volume in response to climatic warming than ice cap lobes in central Iceland.

This hypothesis was successfully tested with the modelling experiments in Chapter 5 (Figure 5.4). The modelling experiment shows that Sólheimajökull undergoes a larger change in glacier length and volume than Hofsjökull in response to the same climate warming experiment. However after 200 years, the length and volume of both glaciers approaches zero and as the glaciers become smaller, differences in their response are less apparent. While the models appear to confirm this relationship, it must be emphasised that the models have inherent limitations and that it will be difficult to accurately predict the future response of Icelandic glaciers to climatic change.

Hypothesis 7

- The Holocene fluctuations of Sólheimajökull reflect climatic changes rather than ice-divide migration. The large extent can be explained by an especially favourable climatic and topographic setting.

In Chapter 4 it was shown that the Holocene fluctuations of Sólheimajökull can be simulated with changes in temperature which are similar to that known from other climate proxy indicators in Iceland. Critically, it is also because climatic changes in the last few centuries have been of short duration which has meant that Sólheimajökull has not been in equilibrium with the climate. The large Holocene fluctuations of Sólheimajökull appear to reflect the favourable climatic and topographic setting of the glacier. While the ELA reconstructions presented in Chapter 4 show that the founding assumptions of the ice-divide migration theory of Dugmore and Sugden (1991) are in error, their hypothesis cannot be rejected outright. This is because fluctuations of the ice-divide were not explicitly modelled in this thesis. Until this is done, the hypothesis presented above must be viewed as a competing hypothesis.

Hypothesis 8

- The spatial variability of other Icelandic glacier fluctuations can be explained by local differences in their climatic and topographic setting. This may also explain why some glaciers appear to have reached their maximum extent during the 19^{th} Century while others display a more complete Holocene moraine record.

In Chapter 6 it was shown that a combination of factors resulting from the climatic and topographic setting of Icelandic glaciers can explain spatial variability in the Holocene moraine record in Iceland. Important processes appear to be differences in the scale of glacial advance resulting from a climatic change, and the moraine retention ability of glacier forelands. In each case, the relative importance of these factors is different. The only means of discriminating between the processes is to study the glacier with a numerical model in conjunction with studies of the proglacial sediments. The predictions in Chapter 7 are seen as a first step, but in reality an encompassing hypothesis such as this cannot provide all the answers.

The Contribution of the Thesis

Mass-balance sensitivity and glacier response

The mass-balance sensitivity of Sólheimajökull as calculated with the energy-balance model is large (1.4 $myr^{-1}\ K^{-1}$), meaning that a large change in mass balance results from a small rise in air temperature. The large mass-balance sensitivity reflects the climatic location and hypsometry of the glacier. Due to high local precipitation and topographic dissection, Sólheimajökull descends to near sea level. The climate of Iceland is especially mild on the south coast due to the proximity of warm North Atlantic Drift waters. This means that ablation occurs year round on the lower part of the glacier. Changes in summer temperature have the largest impact on mass-balance, although changes in spring, autumn and winter temperature are also significant. In contrast, changes in summer precipitation have little impact on mass-balance.

It is accepted in the glaciological literature that in temperate areas, glacier mass-balance is most sensitive to changes in summer temperature. A key finding of this thesis is that significant ablation can occur year round on maritime glaciers terminating near sea level. Studies relating glacier fluctuations to climatic changes need to keep this in mind when searching for a correlation between changes in mass balance and climatic variables. The mass balance of Sólheimajökull is more sensitive to temperature changes than Nigardsbreen (0.9 $myr^{-1}K^{-1}$) (Oerlemans, 1992) a glacier in Norway with similar characteristics. This is because Nigardsbreen is located further inland than Sólheimajökull. The mass balance of Sólheimajökull is less sensitive to climatic change than the mass balance of the Franz Josef Glacier in the Southern Alps of New Zealand (1.6 $myr^{-1}K^{-1}$) (Oerlemans 1997b). This finding is expected because the Franz Josef Glacier also terminates at low altitude near the sea and the climate of the west coast of New Zealand is considerably warmer than the south coast of Iceland.

Topography and glacier response

The effect of valley geometry on glacier response is important. An increase in valley width leads to a local decrease in the sensitivity of the glacier snout to climatic change. Two types of topographic threshold are identified at Sólheimajökull. One topographic threshold occurs at a glacier length of 15 km where the valley widens by 1 km. This threshold influences changes in glacier length, but it does not have a significant impact on glacier volume. This means that for a series of step lowering in ELA, the glacier snout fluctuates by a smaller amount in terms of glacier length as the threshold is approached. On the other hand, glacier volume increases at a uniform rate as the glacier continues to thicken despite the smaller scale of advance. The second topographic threshold occurs where the glacier valley widens into a sandur plain. At this point the glacier essentially decouples from climatic changes, and further changes in the climate will not lead to significant alterations in the length or volume of the glacier.

The identification of topographic thresholds in a landscape is important as they influence how a glacier responds to climatic change (Mercer, 1961). Past interpretations of glacier fluctuations can potentially be in error if topographic thresholds are not identified. Previous modelling studies of valley glaciers have indicated the importance of the bed slope in influencing glacier sensitivity (Oerlemans, 1989), but few modelling studies have highlighted the importance of changes in valley width. Hubbard (1997) illustrated how changes in valley width influenced the fluctuations of Quaternary glaciers in the Chilean Lake District, but to the authors knowledge, this relationship has not been identified in modelling studies of contemporary non-calving glaciers.

There are few cases in the glaciological literature where the influence of topography on glacier response time has been examined. Glacier response time is longer at topographic positions where the glacier snout is undergoes larger changes in glacier length in response to climatic change, and shorter when the opposite condition occurs. This relationship between glacier response and topographic thresholds is not considered in the simple scheme currently used to estimate the response time of glaciers (Johannesson *et al.*, 1989). For example, Johannesson's formula will overestimate the response time if a glacier advances into a widening trough (response time will be shorter). On the other hand, the response time will be underestimated by Johannesson's formula if a glacier advances into a trough where the bed slope is small (response time will be longer). This behaviour has also been confirmed by numerical experiments with an ice flow model for glaciers with different bed slopes and plan forms (Oerlemans, 1989, J.Oerlemans pers. comm. 1999).

Inverse modelling of climatic changes

The record of past changes in glacier length at Sólheimajökull can be used to reconstruct a climate proxy record. This is evident because the reconstructed ELA trend for the 20th Century correlates with the instrumental record of temperature from an Icelandic weather station. Furthermore, a scale factor can be used to relate changes in ELA to changes in air temperature. These reconstructed temperature changes are similar to changes in air temperature recorded at Icelandic weather stations. The ELA reconstruction for the 18th and 19th Centuries also captures climatic changes known from the documentary record of Icelandic climate (Ogilvie, 1992). Limitations in the ELA reconstruction method are identified: Assuming that the glacier reached equilibrium at any time can lead to large errors in the climatic reconstruction. Another potential source of uncertainty in inverse modelling is the influence of topography on changes in glacier extent. The 19th Century ELA reconstruction at Sólheimajökull is of lower quality because the glacier is located at a topographic threshold during this time. This source of uncertainty can be isolated if topographic thresholds are first identified using an experiment such as that described in Chapter 3.

Very few inverse modelling studies of glacier fluctuations have been undertaken previously, despite the fact that a method has been available since the founding studies of Nye (1965). The findings of this study suggest that inverse modelling can be used to accurately reconstruct climatic change if a good record of past changes in glacier length exists. However, the time gap between former glacier length positions should be shorter than the glacier response time, otherwise significant errors might occur in the ELA reconstruction. These errors can be very large if a glacier is incorrectly assumed to reach equilibrium at a position for which the only field evidence is a 'snapshot' such as a painting or written description. If the time gap between known former glacier positions is too long, then there is no choice but to assume that the glacier reached equilibrium with the climate. In such cases, inverse modelling of glacier length may not provide results that are any more reliable than ELA reconstructions derived from simpler methods such as accumulation-area ratios (Porter, 1975).

Predicting the response of Icelandic glaciers to climatic warming

In Chapter 5, the response of Icelandic glaciers to climatic warming scenarios is investigated. The response of Sólheimajökull is calculated for a set of climate warming scenarios that are used in a worldwide modelling study of valley glaciers and small ice caps (Oerlemans *et al.*, 1998). The predicted response of Sólheimajökull is similar to that of large valley glaciers in the European Alps. The response of Sólheimajökull to different warming scenarios is variable. For a warming rate of 0.1 K/decade, and with an increase in precipitation of 10% per degree of warming, Sólheimajökull retreats by only 2 km by the year AD 2100. Conversely, if the climate warms by 0.4 K/decade and precipitation does not increase, then Sólheimajökull retreats by 12 km by the year 2100. This large range of possible responses means that future climatic warming scenarios need to be tightly constrained in order to accurately project future changes in glacier length. At present, there is some uncertainty in the development of future climatic warming scenarios because natural climatic variability in the Arctic is not well simulated in climate models (Battisti *et al.,* 1997).

The predicted retreat of Sólheimajökull is also compared to that of Hofsjökull in central Iceland for a climatic warming scenario specified for Icelandic glaciers (Johannesson *et al.*, 1995). The two glaciers respond differently to climatic change; Sólheimajökull undergoes larger changes in glacier length and volume than Hofsjökull. This suggests that the projection of future glacier length for Hofsjökull (Johannesson, 1997) cannot be considered representative of Icelandic glaciers. We should not expect the response of glaciers to climatic warming to be uniform across the island; the most dramatic retreat of glaciers will occur along the south coast of Iceland.

Glacier sensitivity and the Holocene moraine record in Iceland

In Chapter 6, Icelandic glaciers are categorised according to their sensitivity to climatic change. The sensitivity of a glacier to climatic change in Iceland is shown to depend on its climatic and topographic setting. The precipitation gradient across the island and the relatively predictable topography allow patterns of sensitivity to emerge. Glaciers which undergo the largest changes in extent in response to climatic change are the outlet valley glaciers that flow toward the south coast of Iceland. These glaciers undergo the largest and fastest changes in glacier length in response to a climatic change. They are the least numerous of all glacier types. Most Icelandic glaciers terminate on sandur plains, which means that the scale of glacier advances will be small. The Icelandic glaciers that undergo the smallest changes in glacier extent are the unconfined lobes of large ice caps terminating on sandur plains. They form large parts of all the main ice caps. The classification scheme is used to predict which glaciers are more and less likely to form moraine sequences. Unconfined glaciers terminating on sandur plains are unlikely to form moraine sequences because they advance and retreat over small distances and tend to flush most of their debris into meltwater streams. On the other hand, confined glaciers terminating in valleys and cirque glaciers are more likely to deposit moraine records, because moraines are spread over larger distances and are deposited on valley sides out of reach of meltwater.

This knowledge is used to interpret the pattern of Holocene moraines. Long Holocene moraine records are found around the terminal margins of outlet valley glaciers and cirque glaciers. Holocene moraine records are also found on the lateral margins of outlet valley glaciers terminating on sandur plains. The spacing of the moraine records supports findings from the classification scheme developed in Chapter 6. At Sólheimajökull, an outlet valley glacier (most sensitive), Holocene moraines are spread over 6 km from the present glacier. At Oerafajökull, where several valley glaciers terminate on sandur plains (sensitive), the moraines are spread over 2-3 km. At Oerafajökull, the largest Holocene changes occurred at the glaciers which currently terminate within their valleys, such as Virkisjökull (Gudmundsson, 1998). In Tollaskagi and central Iceland, cirque glaciers expanded only c. 1 km beyond their Little Ice Age extent during the mid-Holocene. This is because they undergo small changes in glacier length in response to climatic change.

This new theory, combined with the ELA reconstruction in Chapter 4 is at odds with the hypothesis that the Holocene fluctuations of Sólheimajökull were influenced by ice-divide migration (Dugmore and Sugden, 1991). It appears that the ELA reconstructions which were the underlying assumptions of the Dugmore and Sugden (1991) hypothesis were in error. However it is not possible to reject the Dugmore and Sugden (1991) hypothesis outright because the models used in this thesis do not test their hypothesis directly. The implication of this conclusion is that fluctuations of Sólheimajökull primarily reflect climatic changes. This is good news for palaeoclimatic studies. It is hoped that the best moraine chronologies in Iceland (such as that present at Sólheimajökull) can be used to help understand local and regional climatic changes.

Holocene climatic change

The reconstructed ELA record at Sólheimajökull is used to understand the nature and variability of Holocene climatic change. Vertical shifts in ELA can be related to changes in incidence of sea ice around Iceland, which in turn represent mean oceanic and atmospheric conditions. Two regimes of Arctic climate prevail. During periods dominated by negative phases of the North Atlantic Oscillation, high pressure is prelevant over Iceland and the East Greenland current carries more ice from the Arctic Ocean to the north coast of Iceland. Temperature is lower by approximately 2°C. During positive phases of the North Atlantic Oscillation, the Icelandic Low dominates the climate of Iceland and sea ice is restricted to the coastal fringe of Greenland, several hundred kilometres from Iceland. Temperature is higher by approximately 2°C.

During the last 300 years, the climate of Iceland appears to have been dominated by decadal-scale shifts in climate between extremes of the North Atlantic Oscillation. This has resulted in a large amplitude signal in the Sólheimajökull ELA record. It is possible to identify a Little Ice Age signal between approximately AD 1750 and 1920, although it was not a uniformly cold period; even this period experienced decadal-scale changes in temperature. The model of the Little Ice Age presented here is closer to the climate record in historic documents (Ogilvie, 1992), than that of a sustained period of cooling (AD 1550 to 1850) suggested from a synthesis of evidence worldwide (Grove, 1998). As these climatic variations took place prior to the Industrial Revolution, they exemplify the frequency and scale at which natural climatic variations occur. This means that the difficulty involved in identifying a climatic warming signal resulting from increased Greenhouse emissions is increased.

It is difficult to use the modelling results to derive an accurate picture of earlier Holocene climatic changes. Yet it is possible to show that the large glacier extent can be simulated with a climatic change of similar magnitude to the cool phases of the last few centuries. The timing of earlier Holocene glacier fluctuations at Sólheimajökull is similar to that of other glaciers worldwide (Denton and Karlen, 1973) and it has recently been shown that these periods are also associated with global atmospheric and oceanic changes (O'Brian *et al.*, 1995, Bond *et al.*, 1997). Yet during the mid Holocene, the polar front was located further to the north than it is today (Koç *et al.*, 1993). Therefore it is possible that the mid Holocene climate of Iceland was more stable than it is currently. Cooling events may have been less dramatic but persisted for longer periods of time. If these cool periods were long enough to allow glaciers to reach equilibrium, it may explain why some glaciers reached their maximum extent during this time.

Future Research Directions

Glaciological studies in Iceland

One of the uncertainties of this study is that no mass-balance measurements have been made on Sólheimajökull. This means that it is difficult to test the mass-balance model properly. This uncertainty also has an impact on the glacier-flow modelling. Ideally the flow model should be tested with the dynamic calibration where the mass-balance input, glacier bed, and former glacier profiles are all well known. It is possible that the choice of flow parameters made in this study may have been affected by an uncertainty in the reference mass-balance profile, although in Chapter 2 (Figure 2.5) it was shown that the ELA reconstruction is fairly insensitive to the choice of flow parameters because the pattern of ELA reconstruction is very similar for a doubling and halving of the flow parameters. This uncertainty can be reduced if a mass-balance programme is initiated.

It is also important to initiate an ice velocity measurement program on Sólheimajökull. This would enable a rigorous testing of the ice flow model, and would provide further insight into the evolution of glacier profiles and response times in southern Iceland. Ideally, short term velocity measurements would be made over a grid on the glacier, and annual velocity would be measured by tracing the movement of markers on the ice surface. This could be carried out in conjunction with a mass balance program.

An interesting test of the modelling results would be to compare the derived mass-balance sensitivity (Equations 3.1 and 3.2) for Sólheimajökull with the mass-balance sensitivity of another maritime outlet glacier in Iceland. The mass-balance sensitivity is important because it underpins the climatic reconstructions (Chapter 4) and the projections of future glacier length (Chapter 5). It may soon be possible to compare the mass-balance sensitivity of Sólheimajökull with the derived mass-balance sensitivity of Breidamerkurjökull, where a glacio-meteorological experiment and mass-balance measurements have been undertaken (Oerlemans *et al.*, 1999). The climate sensitivity of the Langjökull ice cap will also be assessed in the coming years under an ARCICE project carried out at the University of Southampton.

Geomorphic studies

A implication of this study is that some Icelandic glaciers respond to climatic changes more directly than others. In other words, the most responsive glaciers are realistic indicators of climatic changes. The fluctuations of these glaciers will include information on the magnitude and timing of climatic changes. More detailed studies should be undertaken on Sólheimajökull, which has the longest and most detailed moraine record of all Icelandic glaciers. In contrast, glaciers that are topographically pinned will exhibit a similar response to climatic changes of different magnitude. On such glaciers, any attempt to invert a climate signal from former ice front positions will result in a wide envelope of climatic possibilities.

The classification scheme presented in Chapter 6 may be used to select glaciers for future study that are likely to contain unexamined moraine records. Holocene moraine sequences are expected to be found near glaciers that undergo a large response to climatic change, and where moraine forming potential is not inhibited by topography. The most promising sites include the valleys of southeastern Iceland near the margins of topographically confined outlet glaciers of Vatnajökull, and in the forelands of cirque glaciers in northern Iceland. Both these areas contain many glacier forelands that have not been studied in detail. The glaciers in southeastern Iceland are interesting because they are exhibit characteristics which promote a large response to climatic change, and should contain good soil profiles for dating with tephrochronology.

It is difficult to reconstruct climate with an inverse modelling study unless the glacier length record is nearly complete. If glacier length positions are only known at intervals of hundreds or thousands of years, then climatic inferences will be not be very specific. It is possible that further studies of Holocene moraine records in Iceland may not necessarily lead to an improvement in knowledge as the general features of the Holocene climate of Iceland are already well known. The key to unravelling the signal in records of glacier fluctuations is to increase the dating resolution of the moraine record. The best hope is the application of tephrochronology, but unfortunately south eastern and northern Iceland have less complete tephra records than areas near the central volcanic zone. One possibility would be to study a moraine record in conjunction with a sediment record in an ice marginal lake, or to integrate moraine studies with a higher resolution climate record in a nearby ice core. This may yield especially important climatic information if a site can be chosen where many tephra layers can be traced between different climate proxy records.

The findings of this thesis suggest that certain glaciers will be more prone to forming long moraine sequences than others. This may have implications for the study of moraine records more widely. For example, the lack of similar moraines in nearby valleys has been used as evidence against a Younger Dryas age for the Wohaiho Loop Moraine near the Franz Josef glacier in New Zealand (Mabin, 1996), despite solid dating evidence for the age of the moraine (Denton and Hendy, 1994). Yet it is possible that the Franz Josef glacier deposited a Younger Dryas moraine because it exhibited the largest local response. This hypothesis seems valid because the late glacial climatic event seems to have been small in the southern hemisphere, and the Franz Josef glacier is known to be very responsive because it receives high amounts of precipitation and descends to a low elevation (Oerlemans 1997b).

Studies of climatic variability in Iceland

One of the findings of this study is that the climate of the last three centuries has been characterised by decadal-scale changes in temperature. Temperature changes in Iceland are larger than in other areas because they are enhanced by changes in sea ice extent. This finding implies that climatic changes in Iceland may have local significance, but their regional impact might be open to question. This has been addressed by Kelly *et al.* (1987) with regard to the regional significance of the Koch Index, the indicator of sea-ice conditions near Iceland. By statistically comparing the sea ice index with instrumental temperature and pressure records they showed that the sea ice index is a regional, rather than hemispherical, indicator of climatic change. Heavy ice years are correlated with lower temperatures in the northern North Atlantic (Barents Sea, Northern Scandinavia and European Russia). This is due to the Arctic Oscillation, which implies that half of the Arctic will be out of phase at any one time (Mysak and Venegas, 1998). During a cold phase in the North Atlantic (Negative phase of NAO), sea ice is extensive in the Greenland and Barents Sea. At this time, sea ice reaches a minimum extent in the Bering and Beaufort Sea and warming is evident in southern Alaska and along the Russian pacific coast. During warm phases in the North Atlantic, an opposite phase signal is present.

It is possible to explain local departures from overall climatic trends if we improve our understanding of decadal-scale climatic variability in the Arctic. For example, decadal-scale changes between AD 1900 and 1990 resulted in the climate of Iceland undergoing an overall cooling trend during the 20th Century (Einarsson, 1991). This was mainly related to a large decrease in ocean surface temperatures in the North Atlantic during the Great Salinity Anomaly in the late 1960s and 1970s (Mysak and Power, 1991). In contrast, a warming trend is clearly evident in the centre of continents outside the influence of the regional changes in sea surface temperature (Briffa and Jones, 1993). Future studies of the climate of the North Atlantic region should focus on improving our understanding of natural climatic variability. This can best be undertaken by studying the climate system with coupled ocean-atmosphere models and high-quality climate proxy records. Recent work in this field is making progress (Polyakov *et al.*, 1999). Inverse glacier modelling studies may play a role in the development of climate proxy indicators, especially in areas where detailed records of past glacier length exist. Until this natural variability is well understood, it will be very difficult to predict the course of future changes in the climate of the Arctic or predict future changes in the extent and volume of glaciers.

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Appendix 1

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Ice-thickness measurements on Sólheimajökull, southern Iceland
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Ice-thickness measurements on Sólheimajökull, southern Iceland and their relevance to its recent behaviour

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Abstract

We present a radio-echo sounding and ice elevation survey along a central flowline of Sólheimajökull, a 15 km long outlet glacier of the Myrdalsjökull ice cap. The glacier reaches a maximum thickness of 433 m and occupies an overdeepened parabolic trough. The trough form contrasts with nearby canyons incised by jökulhlaups, indicating that it has formed predominantly as a result of glacial erosion. The ice surface shows few large undulations and has a consistent decline in altitude, indicating regular ice flow conditions. The parabolic profile, a gentle, consistently declining bed and ice surface slope, wide accumulation area and long narrow snout are characteristics of a glacier that is sensitive to climatic change. The recent glacier advance (1970-1995) is likely to have resulted from a dynamic response to changing mass balance conditions over the last few decades, in line with other maritime glaciers in the North Atlantic.

Introduction

Sólheimajökull drains from between two domes of the Myrdalsjökull ice cap through a valley 1 to 2 km wide (Figure 1). The ice surface and sub-glacial topography of the upper catchment has been surveyed by Björnsson (in Lawler et al., 1994); however little is known of the outlet glacier. This paper presents an ice radar and GPS survey conducted in 1996 and 1997 as part of a wider project that aims to understand the climatic significance of the fluctuations of Sólheimajökull. Sólheimajökull has one of the most detailed geomorphic records of glacier fluctuations in Iceland (Dugmore, 1989). Although the Holocene fluctuations are similar in timing to other glaciers in Iceland (Stötter, 1994), suggesting a response to a regional climate signal, Sólheimajökull is known for its anomalously large extent during the mid Holocene. Dugmore and Sugden (1991) suggested that this was caused by changes in the position of the M_rdalsjökull ice divide over time. However, other hypotheses may be valid. For example, the large extent may reflect a favorable topographic configeration as the geometry of the glacier trough can influence a glacier's response to a climate signal (Firbish and Andrews, 1984). Also, Sólheimajökull may have exhibited irregular flow in the past through the influence of sub-glacial flood events from the Katla caldera. The ice radar and bedrock profiles from this study give an opportunity to constrain some of these alternative hypotheses.

Survey Methods

The survey was located along a central flowline of the glacier, from the ice-divide to the snout (Figure 1). In addition, two cross profiles were undertaken in order to determine the trough cross section. Seventy soundings were made at 250 m intervals (Table 1). The sounding points were located by differential GPS (Magellan Promark 10). A base station was established at Skogar, 5 km from the glacier and the remote unit was placed midway between the radar transmitter and receiver. At least 10 minutes of psuedorange data were collected at each point, and positions were computed using Magellan post processing software (Magellan, 1991). Accuracy for the positions is +/- 2 to 5 m RMS for X, Y and +/- 6 to 15 m for Z co-ordinates depending on satellite geometry and duration of data logging at each location. Computed positions often varied by over 100 m from uncorrected fixes, justifying the use of differential GPS. Altitude checks were made against the GPS results using a digital aneroid barometer.

Ice depth was determined using a portable monopulse ice radar constructed following the principles of Watts and England (1975). Antennae systems for both transmitter and receiver were resistively dampened dipoles, and were adjusted to change frequency depending on ice thickness. (2.5 MHz for most readings and 5 MHz for ice thinner than ~ 150 m). The antennae were oriented either parallel or perpendicular to the direction of flow, depending on surface geometry, degree of crevassing and signal strength. The transmitter and receiver were separated by 50 m for each sounding. The basal return signals were interpreted in the field on a Fluke Scopemeter 96, but were also downloaded onto a PC so that spurious traces could be re-examined (Figure 2). The accuracy depends largely on the time resolution of the digital storage oscilloscope, where the corresponding points on the transmitted pulse and the ice/bedrock echo can usually be determined to the nearest time-step. In practice these limitations resulted in an accuracy for the thickness determinations of ~5%, and usually less than +/- 10 m. Ice thickness was calculated from the two-way travel time allowing for the spacing between transmitter and receiver, and time for the air pulse to reach the receiver (Bogorodskii et al., 1985).

Results

The long profile of the bed is notable for its lack of large undulations (Figure 3). The most complex topography occurs near the snout, where there is evidence of a sub-glacial hill and an overdeepening. The glacier bed remains close to sea level for 3.5 km inland from the snout, and would become a fjord if ice free under higher sea levels (for example, during the early Holocene). The greatest ice thickness of 433 m was located at points 22 and 23, where the relief of the valley sides is highest (Table 1). The ice is thinner at the top of the survey line at the southern rim of the Katla caldera. The results of two cross sections are shown in Figure 3. Cross section 1 was taken at point 23, 4.5 km from the snout where the glacier reached its maximum thickness (Figure 1). The trough approximates a parabolic profile with steep sides and a gently undulating floor. Cross section 2 was taken at point 34, 1.25 km from the snout (Figure 1). Here the trough was found to be asymmetrical, associated with divergence of the snout around Jökulhaus, with the deepest point occurring below the main (NW) ice stream.

The surface of Sólheimajökull has a gentle long profile with few undulations. An exception occurs near the snout (around Jökulhaus) where the sub-glacial hill is expressed on the glacier surface as transverse and longitudinal crevasses. North of the 1300 m contour, the ice surface levels off dramatically at the transition from shallow ice on the caldera rim to deep ice within the basin (Table 1). A comparison between our survey results and the 1:50000 Iceland Geodetic Survey topographic map (DMA M_rdalsjökull Sheet, 1812 (ii)) reveals that the glacier is 20-80 m thicker in its middle and lower reaches and extended ~ 500 m further in 1996 than when the DMA map was compiled (Figure 4). The date of compilation for the DMA map is unknown, but it pre-dates 1990 and probably occurred before 1980 (O. Sigurdsson pers. com., 1998). Between 1970 and

1995 the glacier advanced a net distance of 482 m (Sigurdsson, 1998). Therefore the map and our survey may represent a before and after snapshot, showing the changes in extent and ice thickness that have occurred on Sólheimajökull during recent times.

Discussion

There are a number of implications following from this work. First, a comparison between the subglacial topography map published in Lawler et al. (1996), and our data shows very good agreement near the ice cap, but the difference increases markedly on Sólheimajökull. Here, Lawler et al. (1996) estimated ice thickness based on the surface slope, and a basal yield stress of 1 bar, assuming perfect plasticity (Paterson, 1994). Our ice thickness measurements differed by up to 200 m (>50%) from the theoretical estimates. The near parabolic profile of the sub-glacial trough contrasts strongly with nearby sub-aerial canyons that have formed predominantly as a result of jökulhlaups from the Katla volcano. This suggests that glacial erosion has been more important than fluvial erosion in trough development. Further, the predictable crevasse patterns, lack of surface bulges or folded moraines and the gently declining ice surface profile suggests that Sólheimajökull flows regularly. We found no evidence in support of Sólheimajökull exhibiting surge-type behaviour, although a velocity monitoring programme is needed to be certain.

Finally, the relatively simple geometry of the subglacial topography and surface profile are characteristics of a glacier that is sensitive to climatic change. Specifically, this includes a wide flat accumulation area, long narrow confined snout, a lack of sub-glacial undulations and a gently inclined bed (Oerlemans, 1989). These characteristics mean that changes in Equilibrium-line Altitude will be effective over a large area of the glacier, and topographic barriers will not inhibit the dynamic response. In addition, Sólheimajökull has a high mass balance gradient (Björnsson, 1979) that results in the glacier snout reaching to low altitude where ablation rates are high. The net result is that Sólheimajökull is active and is likely to have a short response time (Jóhannesson et al., 1989), and experience large changes in extent for small changes in mass balance. The present glacial advance is likely to represent a simple response to changing mass balance conditions over the last few decades, in line with other maritime glaciers in the North Atlantic (Dowdeswell et al., 1997). Similarly, other Holocene fluctuations might reflect a simple climate response.

Conclusions

1. Sólheimajökull was found to occupy an overdeepened parabolic trough, ranging from one to two kilometres in width and reaching a maximum depth of 430 m. The glacier was thicker and had a greater extent in 1996 than as shown on the DMA map. This reflects the glacial advance of 482 m that occurred between 1970 and 1995.

2. Sólheimajökull has the characteristics of an active glacier that flows regularly, is very sensitive to climatic change, and responds quickly. The current advance is likely to reflect recent changes in mass balance. No evidence of surge-type behaviour could be found.

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Figures

Figure 1.

Map showing Sólheimajökull in southern Iceland, the glacier surface and the location of our survey lines.

Figure 2.

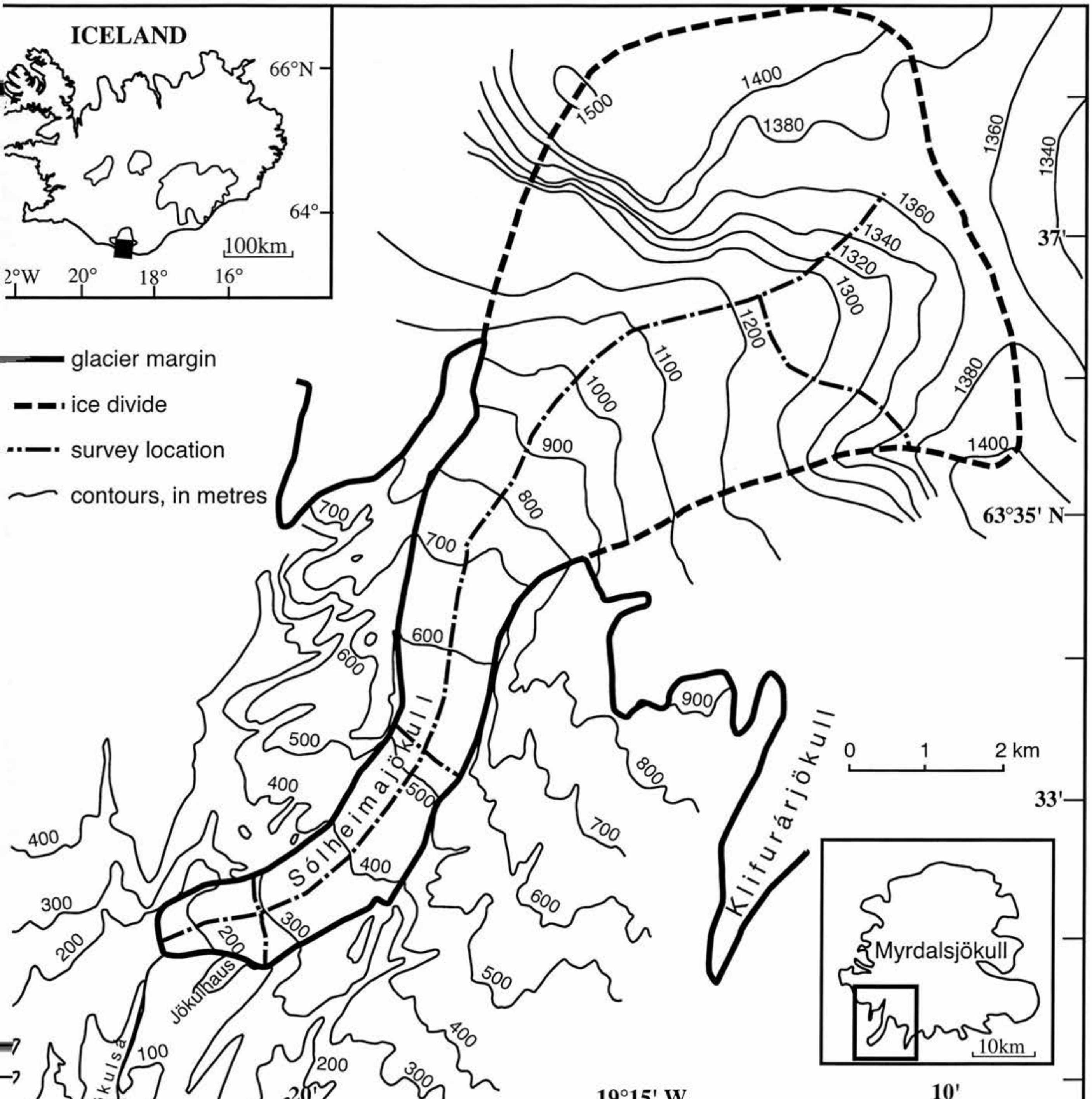
Downloaded oscilloscope screen for a typical basal return signal (Point 2). The cursors are located at the air and return pulses, and the two way travel time is 3.16 microseconds.

Figure 3.

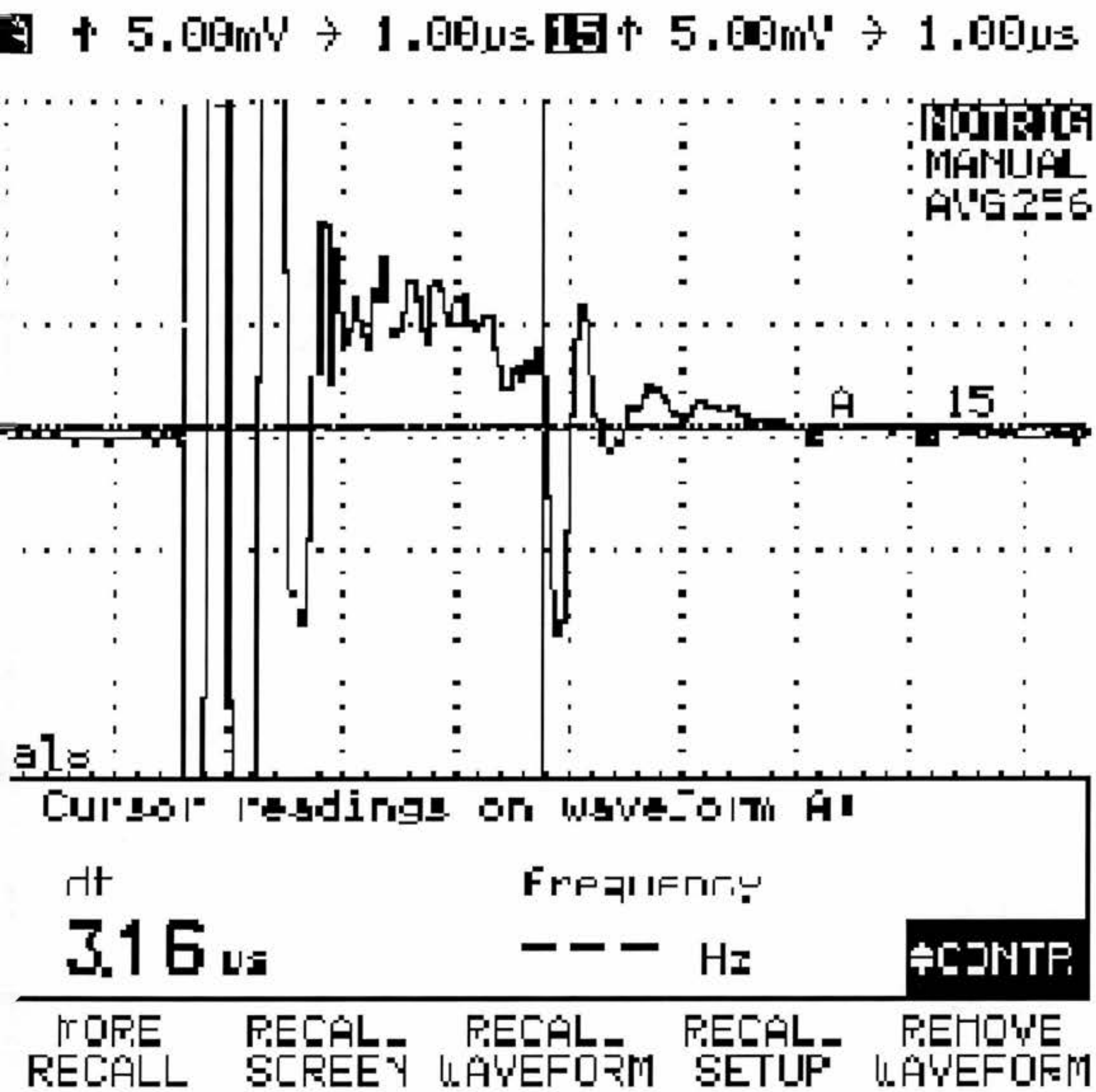
The results of the glacier survey; glacier long profile and cross sections (crosses delineate individual sounding points +).

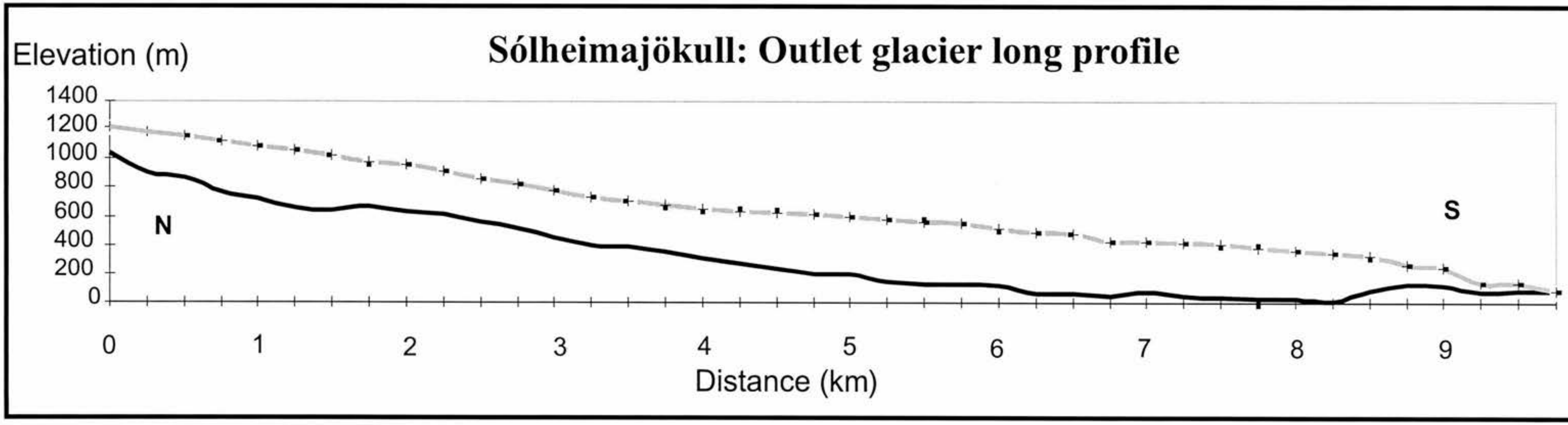
Figure 4.

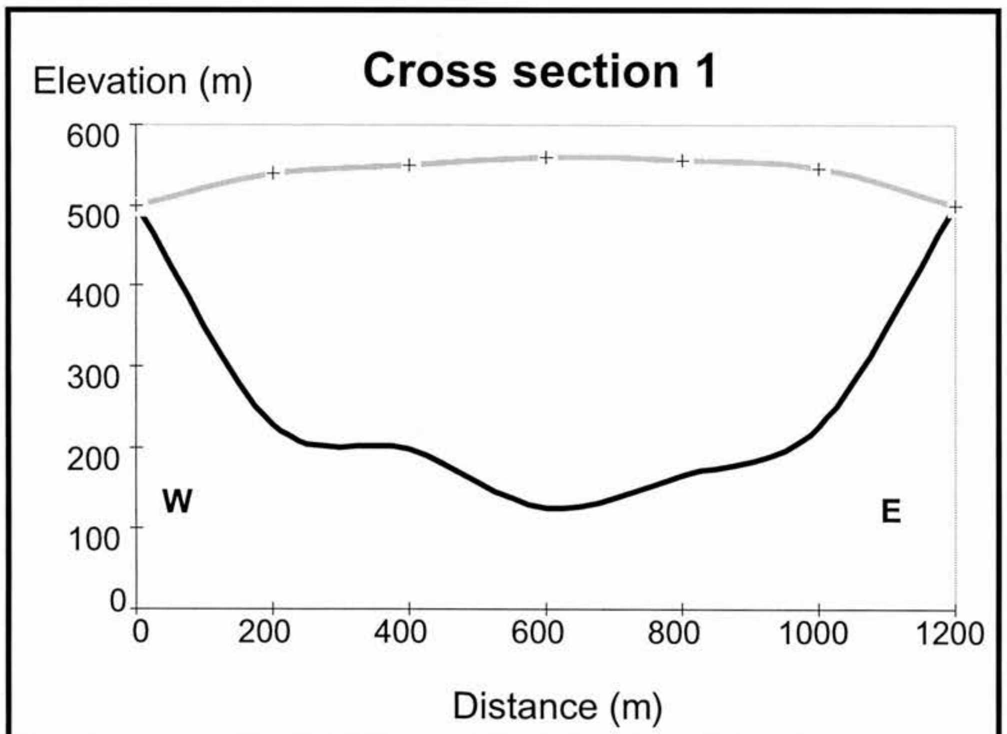
Our surveyed ice elevation (1996/1997) versus DMA map 1812 (pre-1980?). Note the difference between the profiles, especially near the snout, where the glacier advanced by 482 m between 1970 and 1995.

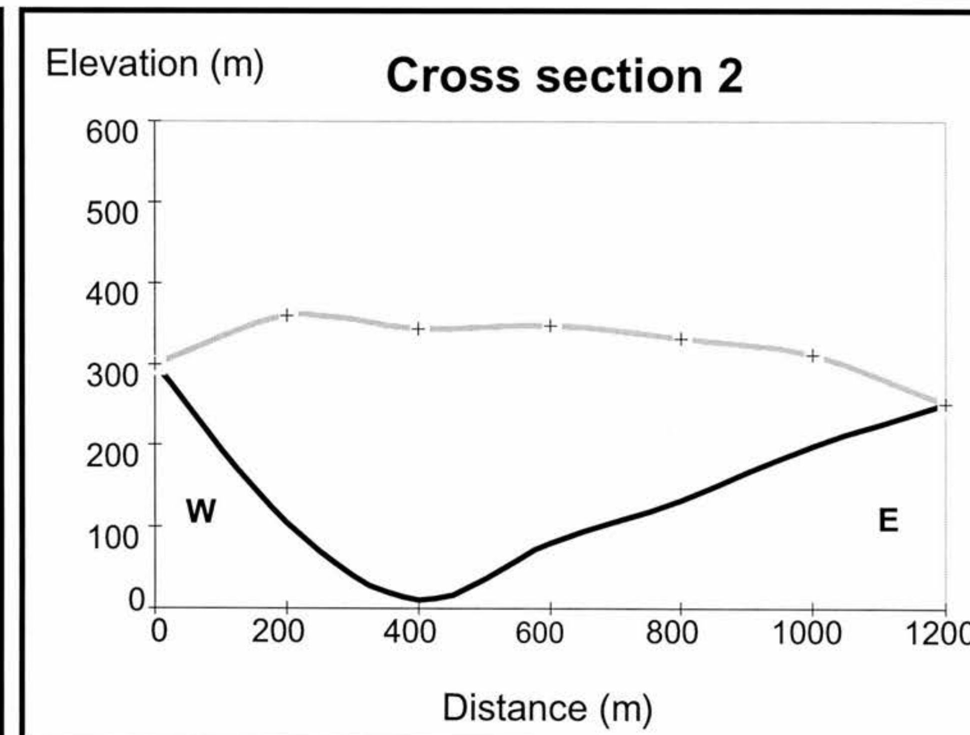


ICELAND
66°N
64°
100km
2°W
20°
18°
16°
glacier margin
ice divide
survey location
contours, in metres
1500
1400
1380
1360
1340
1320
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
37'
63°35' N
33'
0
1
2 km
Sólheimajökull
Klifurárjökull
Jökulhaus
Jökulsá
Myrdalsjökull
10km
20'
19°15' W
10'


↑ 5.00mV → 1.00µs
15 ↑ 5.00mV → 1.00µs
NOTRIG
MANUAL
AVG256
A
15
Cursor readings on waveform A
dt
3.16 µs
Frequency
--- Hz
CONTR
MORE RECALL
RECALL SCREEN
RECALL WAVEFORM
RECALL SETUP
REMOVE WAVEFORM


Sólheimajökull: Outlet glacier long profile
Elevation (m)
1400
1200
1000
800
600
400
200
0
N
S
0
1
2
3
4
5
6
7
8
9
Distance (km)


Cross section 1
Elevation (m)
600
500
400
300
200
100
0
W
E
0
200
400
600
800
1000
1200
Distance (m)


Cross section 2
Elevation (m)
600
500
400
300
200
100
0
W
E
0
200
400
600
800
1000
1200
Distance (m)

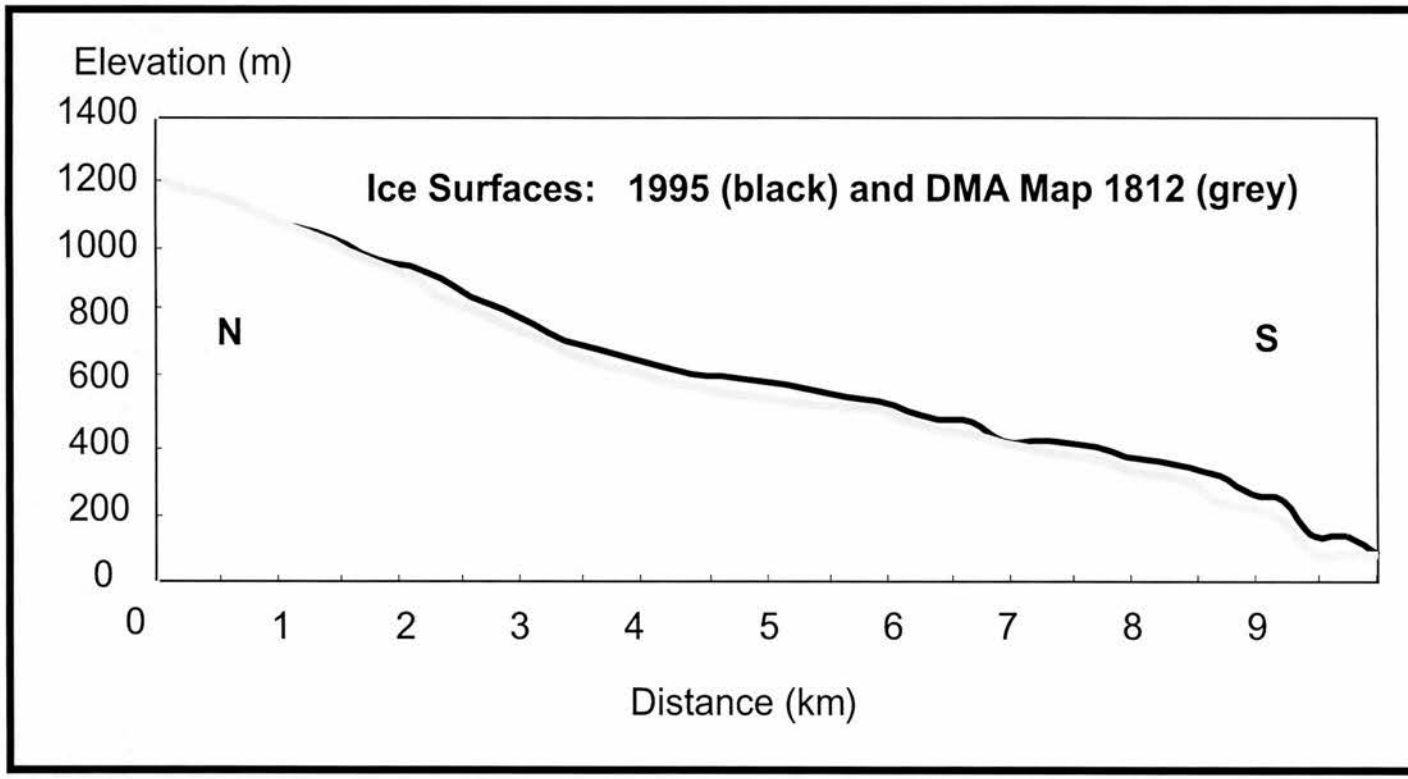

Elevation (m)
1400
1200
1000
800
600
400
200
0
Ice Surfaces: 1995 (black) and DMA Map 1812 (grey)
N
S
0
1
2
3
4
5
6
7
8
9
Distance (km)

Table 1: Sólheimajökull Ice Radar Survey

Point	Measured Time (micro sec.)	Ice Thickness (metres)	Location UTM, Zone 27 5,70	Surface Elevation m.a.s.l	Bed Elevation m.a.s.l
Outlet Glacier					
1	1.96	177	88 200, 55 280	1220	1043
2	3.16	278	88 080, 54 800	1180	902
3	3.28	288	87 760, 54 600	1160	872
4	4.08	356	87 455, 54 415	1120	764
5	4.12	359	86 946, 54 137	1080	721
6	4.52	393	86 552, 53 950	1055	662
7	4.36	379	86 262, 53 723	1020	641
8	3.48	305	85 790, 53 494	975	670
9	3.64	319	85 517, 53 245	955	636
10	3.44	302	85 225, 52 958	915	613
11	3.4	299	84 965, 52 682	860	561
12	3.52	309	84 990, 52 343	820	511
13	3.76	329	84 998, 51 926	780	451
14	3.8	332	85 026, 51 516	730	398
15	3.64	319	84 842, 51 169	705	386
16	3.72	326	84 892, 50 577	680	354
17	3.92	342	84 600, 50 456	650	308
18	4.08	356	84 587, 50 200	630	274
19	4.44	386	84 488, 49 997	620	234
20	4.76	413	84 490, 49 796	610	197
21	4.54	395	84 459, 49 576	595	200
22	5	433	84 457, 49 234	580	147
23	5	433	84,331, 49 077	559	126
24	4.88	423	84 312, 48 770	549	126
25	4.56	396	83 919, 48 106	512	116
26	4.96	430	83 860, 48 142	490	60
27	4.8	416	83 375, 47 966	482	66
28	4.4	383	83 583, 47 707	428	45
29	4.08	356	83 342, 47 560	428	72
30	4.28	373	83 164, 47 373	419	46
31	4.2	366	82 971, 47 152	405	39
32	4.04	353	82 717, 47 030	379	26
33	3.78	331	82 491, 46 877	362	31
34	3.8	332	82 245, 46 845	343	11
35	2.72	241	81 676, 46 117	322	81
36	1.54	141	81 442, 46 663	266	125
37	1.42	131	81 210, 46 512	246	115
38	0.68	67	80 919, 46 119	140	73
39	0.5	50	80 851, 46 154	135	85

Point	Measured Time (micro sec.)	Ice Thickness (metres)	Location UTM, Zone 27 5,70	Surface Elevation m.a.s.l	Bed Elevation m.a.s.l
X-Section 1					
1	3.56	312	84 086, 48 830	540	228
2	4.04	353	84 385, 49 890	550	197
3	5	433	84 331, 49 077	559	126
4	4.48	390	84 540, 49 042	555	165
5	3.64	319	84 701, 49 009	545	226
X-Section 2					
1	2.88	255	82 271, 47 046	360	105
2	3.8	332	82 246, 46 845	343	11
3	3.02	267	82 368, 46 689	347	80
4	2.22	199	82 379, 46 539	332	133
5	1.22	114	82 360, 46 360	312	198
Ice Cap					
1	2.96	261	90 610, 53 135	1370	1109
2	2	180	90 477, 53 322	1368	1188
3	2.76	245	90 294, 53 508	1360	1115
4	2.8	248	90 132, 53 640	1359	1111
5	3	265	89 896, 53 817	1358	1093
6	3.32	292	89 946, 53 804	1350	1058
7	3.12	275	89 549, 53 903	1348	1073
8	3	265	89 287, 53 896	1343	1078
9	3.2	282	89 032, 53 940	1338	1056
10	3.24	285	88 699, 54 094	1336	1051
11	2	180	88 492, 53 971	1332	1152
12	3.08	272	88 699, 54 094	1328	1056
13	3.12	275	88 443, 54 702	1320	1045
14	3.52	309	90 057, 55 982	1369	1060
15	3.3	290	90 167, 54 722	1366	1076
16	3.28	288	89 300, 54 912	1335	1047
17	3.52	309	89 270, 54950	1330	1021
18	3.4	299	88 620, 55 186	1320	1021
19	3.4	299	88 032, 54 719	1316	1017
20	3.4	299	88 667, 54 491	1303	1004
21	3.44	302	88 298, 54 733	1313	1011
22	3.4	299	87 765, 54 674	1268	969