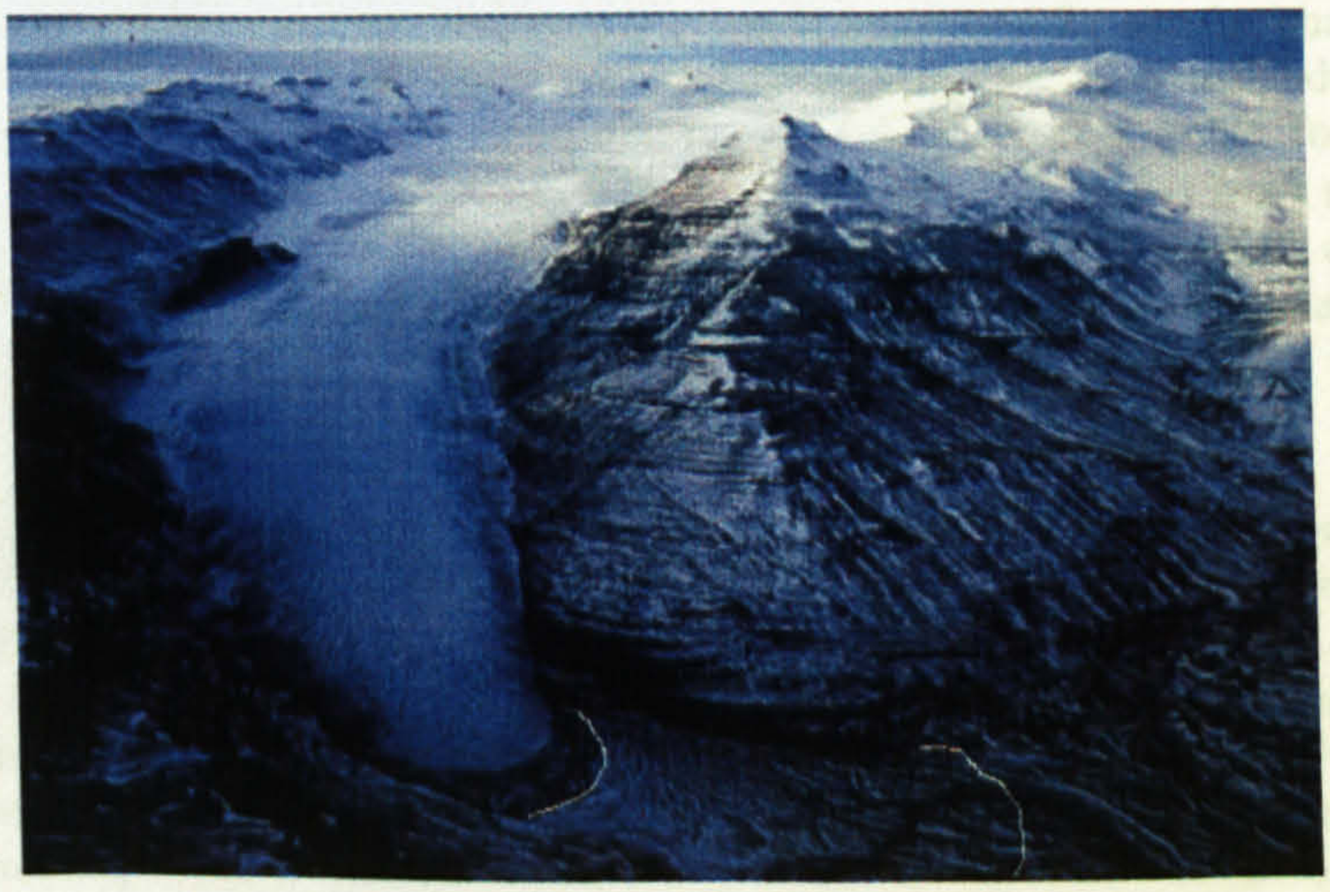
Glacier Fluctuations, Lichenometry and Climatic Change in Iceland

Tom Bradwell



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Abstract

This thesis examines the spatial and temporal expression of Holocene glacier fluctuations in southeast Iceland. The study uses geomorphological evidence to reconstruct the former extent of Lambatungnajökull – a non-surging, valley glacier flowing from the eastern flank of the Vatnajökull ice-cap. Lichenometry is used to date recent glacial landforms and decode the pattern of glacier fluctuations over the last 300 years. Tephrochronology is used to date older features (<10,000 yr). The results show a pattern of fluctuations driven by climatic change. During the Lateglacial–Early Holocene Period the glacier terminus was situated at the present-day coastline. The ice-margin has retreated c. 20 km during the last 10,000 years. At least four periods of glacier re-advance have been identified, at c. 5000, 3000, 1600 and 170 years BP. Overall, the cumulative ice recession since c.10 ka BP represents an ELA rise of c. 400 m which equates to an increase in mean air temperature of at least 2°C, assuming constant precipitation levels.

Since the late 18th century, Lambatungnajökull has been in overall retreat. Moraines dated using two different lichenometric techniques indicate that the most extensive period of glacier expansion during historical time culminated shortly before c. AD 1795, probably in the 1780s. Recession over the last 200 years has been interrupted by re-advances in the 1850s, 1870s, and c. AD 1890. In the 20th century, most notably in the 1930s and 1940s, Lambatungnajökull receded more rapidly than at any time during the previous 150 years. However, brief cold spells (~5 yrs), centred around the years AD 1918 and AD 1964, temporarily halted glacier recession. Lambatungnajökull has only retreated slightly over the last 20 years. The degree and nature of glacier retreat since 1930 compares well with similar-sized glaciers in southern Iceland. Furthermore, the pattern of glacier fluctuations over the last 150 years reflects the temperature oscillations recorded at nearby meteorological stations. Much of the climatic variation experienced in southern Iceland, and the glacier fluctuations that result, can be explained by secular changes in the North Atlantic Oscillation. A shift to more zonal atmospheric circulation and a weaker Icelandic Low – resulting in a greater frequency of negative NAO anomalies – may have been responsible for the cooling and associated glacier advances of the 18th and 19th centuries. One implication of this work relates to the exact timing of the *Little Ice Age* in the Northeast Atlantic. The advanced position of glaciers during the late 18th century suggests that this period represented the culmination of the *Little Ice Age* in Iceland. This contrasts with the current consensus that the *Little Ice Age* 'glacier maximum' in southern Iceland was during the late 19th century.

Other implications concern lichen-dating and its wider applications. Firstly, this research shows that the 'growth' curve of yellow-green *Rhizocarpon* lichens over the last 270 years in southeast Iceland is not linear. Although growth rates appear constant for periods of several decades, the growth 'curve' is exponential overall, with larger (older) lichens apparently growing more slowly than smaller lichens. Secondly, growth rates of *Rhizocarpon* Section *Rhizocarpon* in Iceland vary as a function of climate, with growth in the moist, maritime, southeast being c.40% faster than in the cooler and drier northwest. Thirdly, this growth rate relationship – across the Northeast Atlantic region as a whole – can best be expressed in terms of climatic 'oceanicity' ($r^2 = 0.95$). This latter relationship could be used to estimate lichen growth rates in areas where dating curves cannot be constructed. Finally, these findings suggest that lichen growth rates are likely to have varied in response to climatic change. In Iceland, slow-growing lichens, such as *Rhizocarpon*, probably grow more rapidly now – since the climatic amelioration of the 1920s and 30s – than they did in the cooler and drier periods of the 18th and 19th centuries.

Declaration

I made this thesis. All the work included within these pages – text, maps and figures – are original and my own, unless indicated otherwise.

Signed

Tom Bradwell May 8th 2001

"It is man's nature to wish to see and experience the things that he has heard about and thus to learn whether the facts are as told or not."

The King's Mirror (p142)
– 13th century Icelandic text

"Where the Yokul has pushed forward in one direction, and again receded, large heaps of clay, sand and turf are thrown up so as to form a catenation of small hills round its base; but where its progress is continuing, no such hills are seen; only furrows are laid open in the sand, by the sharp projecting pieces of ice..."

Moraines formed by Breiðamerkurjökull described by Ebenezer Henderson on his travels around Iceland in 1815.

Henderson (1819 : 199)

Acknowledgements

Aspects of this research have been funded by several organisations: *The Department of Geography* at the University of Edinburgh, *The Carnegie Trust for the Universities of Scotland*, *The Royal Geographical Society* (Dudley Stamp Memorial Fund), *Oikos*, *The National Science Foundation* and the *North Atlantic Biocultural Organization*. I would like to thank them all.

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In Edinburgh, numerous people have added to the quality of my life over the last 3 years. In the Department of Geography, Sarah Davies, Anthony Newton, Nick Hulton, Ross Purves, Donna Easterlow and Kerry Watt (my favourite librarian), as well as fellow postgraduates – Chris, Fran, Lorraine, Anja-Maaike, Ruth, Pete, Kate, Clare, Nick, Darcey, Guðrun, Rafael and Javier – have been good friends. Mr. E. W. J. Murphy continues to be an extremely welcome distraction after 10 years of acquaintance.

Ann Breen has been a constant source of encouragement during the final stages of writing this thesis. Not only did she take on the daunting task of 'managing' my reference list but she has also tolerated exquisitely her nail-bitten, hard-up, rather pre-occupied boyfriend over the last 6 months!

Final thanks must go to my mother and father for their unwavering support throughout my university career.

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Chapter 1 : Aims, Introduction and Scientific Rationale

Aim

The aim of this research is to use geomorphological evidence, coupled with lichenometrical and tephrochronological dating methods, to decode the sequence of glacier fluctuations in southeast Iceland and add to current knowledge of Holocene climatic change.

There are three overall objectives:

- Delimit, spatially and temporally, the former extent of a previously unstudied glacier in Iceland.
- Synthesise these findings with existing knowledge of glacier fluctuations in Iceland.
- Reconstruct the climate associated with the regional pattern of Holocene glacier fluctuations in Iceland.

This chapter aims to provide background to these specific objectives and address the scientific rationale for undertaking this research.

Why study Climatic Change?

Recent studies have highlighted the warming of the Earth's climate that has taken place over the past century (eg. Jones *et al.*, 1986; Folland *et al.*, 1990; Houghton *et al.*, 2000). Many scientists have argued that this warming is a direct result of increased anthropogenic activity and the high concentration of 'greenhouse gases' in the atmosphere. Climatic models predict that global temperatures will rise as 'greenhouse gas' levels continue to increase, with the greatest degree of temperature change being experienced in polar regions.

Clearly the climate has changed considerably over the last 100,000 years. Presently we live in an Interglacial Period, a warm phase between two Ice Ages. Climatic cycles on this timescale are relatively well understood (eg. Ruddiman and McIntyre, 1979; Imbrie, 1982). However, the causes of changes on millennial, centennial, and decadal scales are more elusive. The role of solar variability and volcanic eruptions on climatic change may be of particular importance (Lamb, 1977; Zielinski, *et al.*, 1994; Lean *et al.*, 1995; Karlén and Kuylenstiernia, 1996; Van Geel *et al.*, 2000). A greater knowledge of climatic change over the last 10,000 years, particularly from high latitudes, will help scientists to make clearer judgements on the relative magnitude of the present climatic oscillation. As stressed by Houghton *et al.*, (2000), the warming of recent times needs to be put in some sort of context, particularly if the information is used to make global economic policies. Scientific studies that attempt to reconstruct the climate of the past several centuries and millennia are therefore of the highest priority.

Why study Glacier fluctuations?

Glaciers are climatically controlled features, being the direct result of particular temperature and precipitation conditions. It has been shown that temperate glaciers and small ice-caps are sensitive to climatic change. Both the direction and magnitude of climatic change directly affects the mass balance of glaciers. Fluctuations of glacier volume and length reflect prevailing climatic conditions closely, particularly in more maritime locations. Observed rates of glacier retreat or advance over many years have been related primarily to variations in air temperature and precipitation (eg. Hoinkes, 1968; Nesje *et al.*, 1989; Sigurdsson, 1998). However, the response of a glacier to climatic change may not be immediate due to the mass of ice involved and the particular glacier dynamics. Furthermore, the speed of response depends on the climatic location and the rate at which the mass balance of the glacier changes with altitude. Glaciers in high precipitation areas have steep *mass balance profiles* and consequently higher turnover rates. Ice-caps in sub-arctic, high precipitation zones such as Iceland, western Norway and New Zealand should be more sensitive to temperature change than glaciers

in drier, higher latitude locations. Working in southern Iceland, Mackintosh (2000), has partly confirmed this hypothesis using a numerical model.

Many scientists have used movements of glacier fronts to make inferences about climatic change (eg. Ahlmann, 1948; Andrews, 1972; Denton and Karlén, 1973; Rodbell, 1990; Oerlemans, 1998; etc). Landforms such as moraines, indicating a more advanced position of the glacier, tell of a different climate than that of today. However, the exact relationship between meteorology and the growth or decay of glaciers is complex. For instance, a once larger glacier does not simply imply a cooler climate rather, that conditions were more conducive to ice accumulation than ablation. For this reason the use of glaciers as key climatic proxies should be employed alongside other palaeo-environmental sources to develop a clearer picture.

Why Iceland?

Iceland is a uniquely important location for the study of climatic change, lying within the Convergence Zone between temperate-maritime and Polar air masses. Slight migrations of this atmospheric front largely determine the relative mildness or severity of the Icelandic climate. The island also currently lies astride the boundary between warm Atlantic Ocean water, of sub-tropical origin, and cool Arctic Ocean water. Movements of this boundary, the Oceanic Polar Front, throughout the Holocene have been associated with climatic fluctuations on a decadal to millennial scale (Lamb, 1979; Kelly *et al.*, 1987; Koç *et al.*, 1993).

High latitude regions have experienced a greater degree of warming over the course of the 20th century than middle and low latitude regions (Houghton *et al.*, 2000). Therefore, marked climatic changes throughout history should be more evident in Iceland compared with other European sites. This factor should allow relatively high-resolution records of past variations to be identified.

In Iceland, particularly in the southeast, there are many polythermal (temperate) glaciers with steep mass-balance gradients that are particularly sensitive to climatic fluctuations. Studies have shown that glaciers confined within long, narrow, topographic depressions, experience a greater degree of advance or retreat, for a given change in mass balance, than glaciers terminating on broad plains (Mercer, 1961; Furbish and Andrews, 1984; Oerlemans, 1989). A further practical reason for studying such glaciers regards the potential moraine record. Moraine preservation is more likely in front of narrow glaciers due to the topographic constriction of meltwater. In the case of wide, lobate glaciers meltwater is free to roam laterally, eroding moraines and hence rendering the geomorphological record incomplete.

Approach

It is clear from work previously conducted in the North Atlantic that the climate of the region has changed considerably over the last 10,000 years (Dansgaard *et al.*, 1984; Meese *et al.*, 1994; Veum *et al.*, 1992; Mann *et al.*, 1999; Bradley, 2000). However, until relatively recently, studies of glacier fluctuations conducted in Iceland revealed little evidence of this Holocene climatic variability, for various reasons. Dugmore's (1987, 1989) work on Sólheimajökull showed that glaciers were larger during the mid-Holocene (4-7000 years BP) and, consequently, that the record of North Atlantic climatic change was more complicated than hitherto thought. These issues are examined in Chapter 3.

Between c. AD 1600-1900 the climate of much of the Northern Hemisphere was unusually harsh (Bradley and Jones, 1993). This period, often called the *Little Ice Age*, represents an interesting climatic anomaly. The weather in Iceland during this time is thought to have been particularly cold, inclement and changeable (eg. Thorarinsson, 1960; Bergthorsson, 1969; Ogilvie, 1984, 1992). Glaciers grew, reaching their greatest

extents in the 18th and 19th centuries (Thorarinsson, 1943; Grove, 1988). Average temperatures were 0.5-1.5°C lower than the AD 1931-1960 mean (Bradley and Jones, 1993).

This thesis aims to reconstruct the former extent of glaciers in southeast Iceland using geomorphological, tephrochronological and lichenometrical means. A similar pattern of glacier fluctuations to that observed by Dugmore (1989) would add further support to the hypothesis of multi-phase Holocene *Neoglaciation.* Furthermore, a high resolution chronological study – focusing on the exact timing and magnitude of *Little Ice Age* glacier advances – would shed more light on the climatic deterioration of recent centuries. The palaeo-environmental implications of Holocene glacier fluctuations in Iceland, and the wider North Atlantic region, will be addressed on three timescales:– annual and multi annual, decadal and centennial.

Thesis Structure

This thesis is divided into eleven chapters. Each of the chapters attempts to fulfil a specific objective, as laid out below. The rationale behind each chapter will be discussed briefly at this point.

In order to place the results of this research into a climatic context, the degree of climatic change in the Northeast Atlantic over the last few centuries is summarised. This is done in Chapter 2. Furthermore, existing knowledge regarding the behaviour of Iceland's glaciers since the last Ice Age is reviewed in Chapter 3. Once the location of the study site and the research methods have been outlined, in Chapter 4, a large section of the thesis is dedicated to lichenometric dating (Chapters 5, 8 and 9). The detailed review of research on lichen growth processes and growth rates, in Chapter 5, is central to the debate surrounding the use of lichenometry as a dating tool. A sound understanding of the processes in operation will aid the interpretation of any lichenometric results. Chapter 6 describes the geomorphology of the study area in relation to former glacier fluctuations, whilst Chapter 7 presents the tephrastratigraphic findings. Lichenometric dating curves are established in Chapter 8. These are then used to date features in the vicinity of Lambatungnajökull. At the end of this chapter, two lichenometric-dating methods are critically compared. Chapter 9 highlights the role of environment on lichen growth rates. A model is derived for lichen growth rates in the Northeast Atlantic region based on climatic parameters. The model is critically tested using empirical data and the implications of the model are discussed. Chapter 10 synthesises the findings of Chapters 6, 7 and 8 and places the reconstructed glacier fluctuations into a climatic context. Finally, Chapter 11 highlights the main contributions of this thesis and suggests avenues for further research.

Chapter 2 : The Climate of Iceland and the Northeast Atlantic

A : The climate of the Northeast Atlantic

Introduction

The Atlantic Ocean exerts a strong influence over the climate of Northwest Europe. In particular, the climate of coastal regions such as the British Isles, Norway, Iceland and Greenland is fundamentally determined by the temperament of the North Atlantic Ocean.

The mild conditions enjoyed by the countries of Western Europe represent one of the World's most interesting climatic anomalies. The reason for the relative warmth experienced at such high latitudes (50-65°N) is primarily due to the warm waters that originate at tropical latitudes and flow northeast, delivering heat to the western coastal fringes of Europe. This surface ocean current, known as the Gulf Stream and continuing northwards (>50°N) as the North Atlantic Current, maintains relatively high winter temperatures in the British Isles and prevents sea ice from forming around Iceland and in the Norwegian Sea. The presence of mild sea-surface conditions acts to regulate air temperature in coastal regions of Northern Europe in all but the most exceptional years. Underpinning these surface currents is a deep-water oceanic 'conveyor' which is driven by density contrasts within the water column. The constant production of deep water in the sea between Iceland and Norway is believed to keep Northern Europe 5 - 8°C warmer than it would be were the 'conveyor' to shut down (Broecker, 1990).

The atmospheric circulation

The weather over mid-latitudes is related to variations in the position and strength of fast-flowing air currents or *jet streams* in the upper troposhere (Rossby, 1949). These jet streams, deflected by the rotation of the Earth, are sometimes referred to as the

upper *Westerlies*. The Polar jet stream is continuous in the Northern Hemisphere. Where it crosses prominent topographic barriers such as the Rocky Mountains the air flow is disturbed setting up an oscillation known as a *Rossby Wave*. Changes in the amplitude and position of these high-altitude Rossby Waves influence the weather experienced at low levels across the North Atlantic and Western Europe.

Research has shown that the position of the atmospheric (Arctic) Polar Front is determined to a greater or lesser degree by the strength, location and amplitude of Rossby waves within the westerly jet stream (Lamb, 1977; Bryant, 1997). The eastern North Atlantic is presently characterised by strong meridional air flow as a result of large amplitude waves in the upper Westerlies transporting more air from tropical to polar regions. During periods when Rossby wave amplitude is small, air flow becomes more zonal – flowing from west to east. The position of the main 'weather-generating' atmospheric systems, such as the *Icelandic Low* and *Greenland High*, are consequently linked to the strength and amplitude of the high altitude jet stream. Some authors dispute this connection stating that a clear relationship between high altitude westerly winds and surface weather conditions has yet to be established (Bryant, 1997).

The North Atlantic Oscillation

The weather of the eastern North Atlantic (ie. the region between 50 and 70°N and 20°E to 50°W) is presently dominated by maritime, northeastward-moving air masses. Depressions normally form in the western North Atlantic off Newfoundland and move slowly east and north before reaching their maximum intensity in the vicinity of Iceland. The strength of this *Icelandic Low* is known to fluctuate in relation to the overall vigour of the general atmospheric circulation. Although not a permanent feature, the *Icelandic Low* appears to be quasi-stationary with only a slight oscillatory migration observed on a decadal time-scale (Maksimov *et al.*, 1967). This high latitude pressure system is only half of an atmospheric dipole operating in the North Atlantic. The presence of an anticyclonic cell over the subtropical North Atlantic, the *Azores High*, sets up a strong meridional surface air pressure contrast. This pressure 'see-saw' and the climatic changes associated with it are known as the *North Atlantic Oscillation* (NAO) (van Loon and Rogers, 1978). The variations in the NAO from year to year are usually defined by the difference between the normalised sea-level

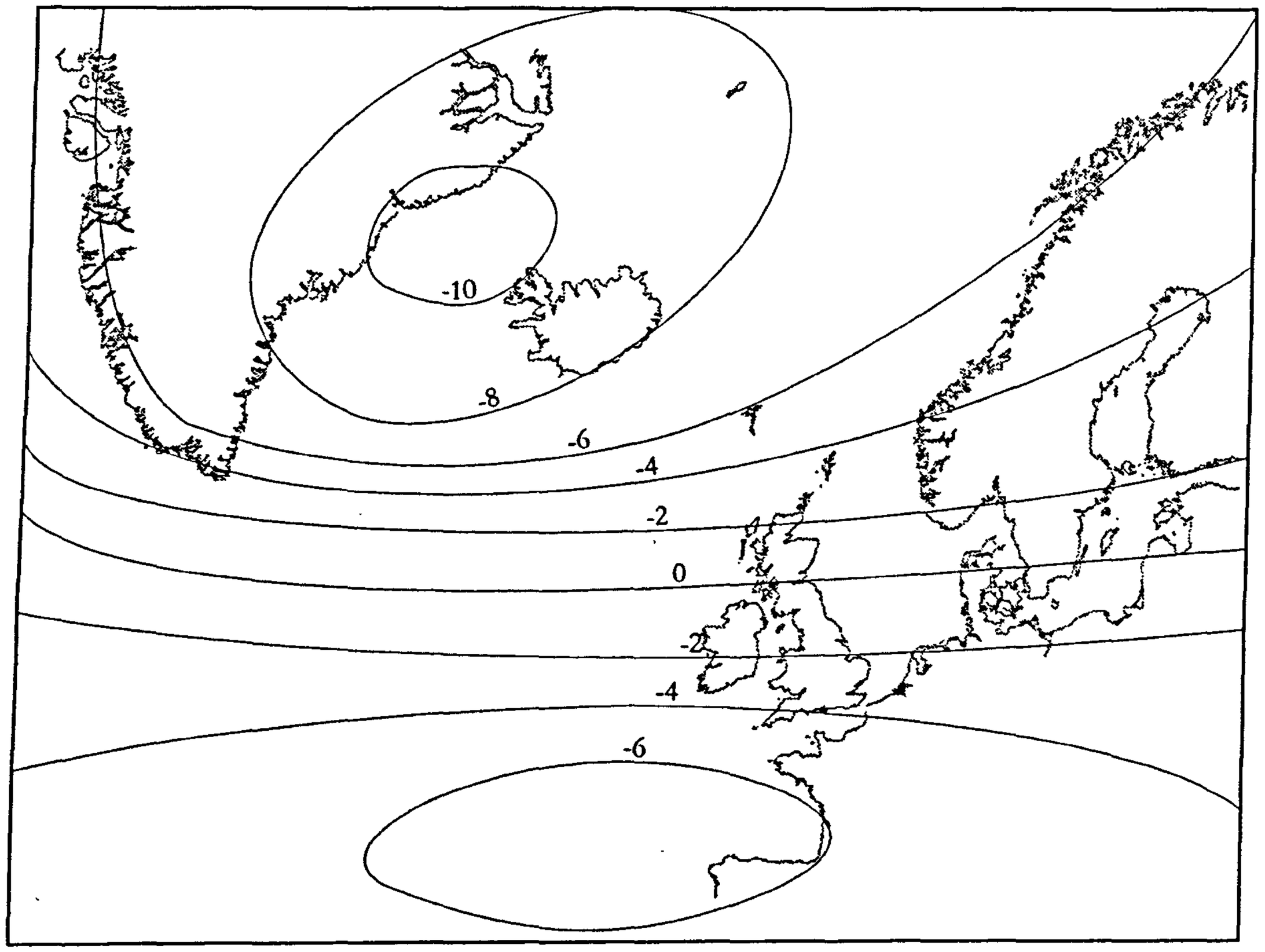
pressure anomaly at Lisbon, Portugal and Stykkisholmur in Iceland (Hurrell, 1995) (Fig. 2.1). More recently, Jones *et al.* (1997) have extended the NAO index back to AD 1823 using data from southwest Iceland and Gibraltar. A highly positive NAO index indicates stronger than average surface westerly winds over mid-latitudes whilst negative values are due to an increased frequency of mobile, polar high pressure systems in the vicinity of Iceland resulting in a blocked situation.

The NAO is believed to be an expression of the Northern Hemisphere Annular Mode, otherwise known as the *Arctic Oscillation* (Wallace, 2000). Being a hemisphere-wide atmospheric phenomenon, the Arctic Oscillation may communicate weather patterns to the Southern Hemisphere. Links between the NAO and El Nino–Southern Oscillation (ENSO) are still uncertain although there is growing evidence of inter-hemispheric synchroneity (Rogers, 1984; Mysak *et al.*, 1996; Huang *et al.*, 1998). Ultimately, the driving mechanism remains elusive – the strength of high-altitude Rossby waves being the most likely source of external perturbations on a multi-annual timescale (Rogers, 1984; Perlwitz and Graf, 1995).

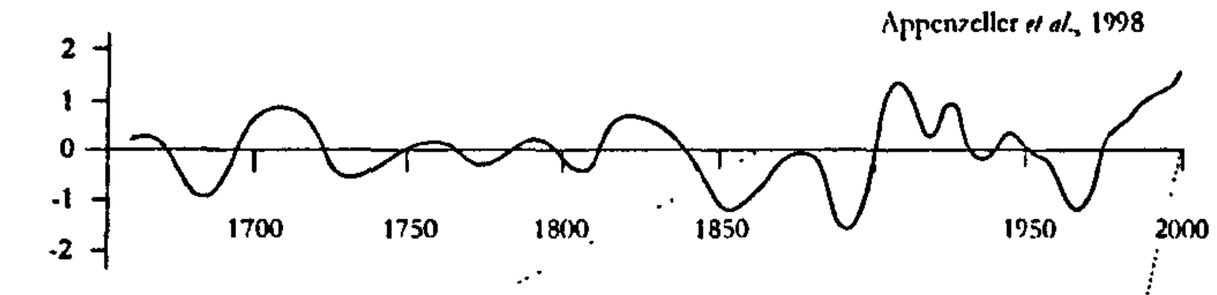
Much of the interannual climate variability experienced in the North Atlantic region has been attributed to the NAO. Hurrell (1995) has shown convincingly that increased wintertime air temperatures over much of Northern Europe are associated with the positive phase of the NAO. Similarly, the NAO has been linked with other atmospheric and oceanic phenomena including: – the path and intensity of the North Atlantic storm track (Rogers, 1984; Hurrell, 1995), evaporation and precipitation patterns (Hurrell, 1995), sea surface temperatures (Deser and Blackmon, 1993; Sutton and Allen, 1997), the extent of Arctic sea-ice (Chapman and Walsh, 1993) and ocean wave height (Bacon and Carter, 1993). Furthermore, the NAO has recently been held responsible for co-ordinating and modulating the intensity of deepwater convection in the ocean at the three main North Atlantic sites (Norwegian Sea, Labrador Sea and Sargasso Sea) (Dickson *et al.*, 1996). The extremely negative values of the NAO in the 1960s may also have been a key factor in generating the largest known recent disruption of the freshwater balance, during the so-called 'Great Salinity Anomaly' (Mysak and Power, 1996).

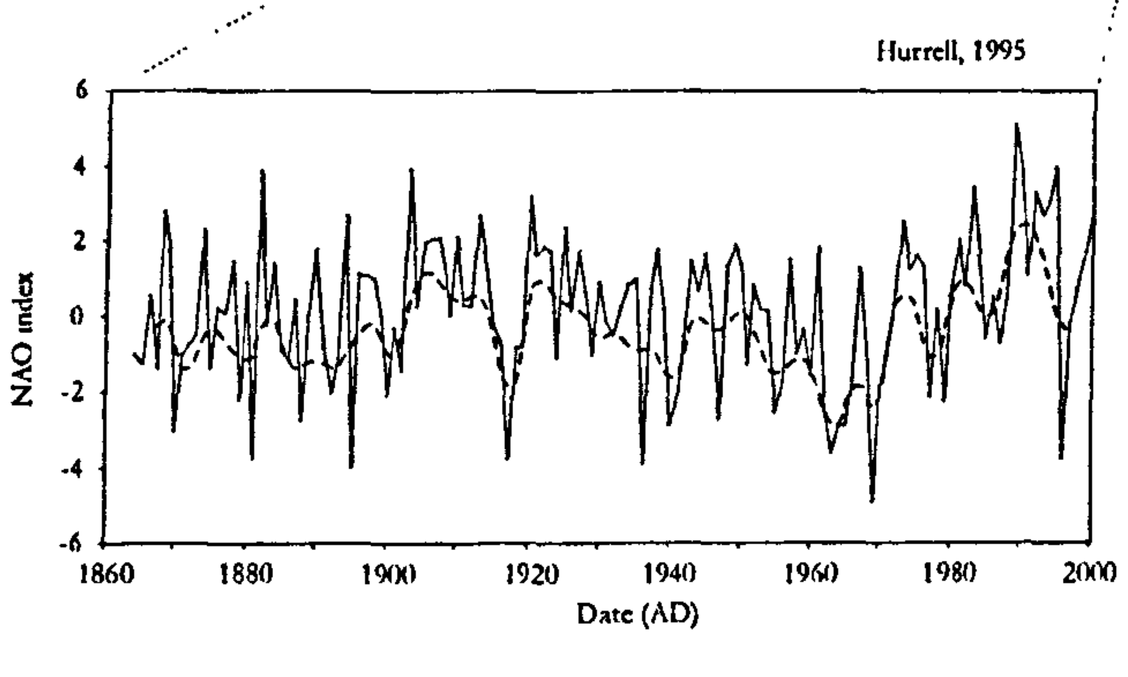
Figure 2.1 The North Atlantic Oscillation (NAO) – a meridional oscillation of atmospheric mass with centres of action near Iceland and the Azores.

Acting like a see-saw, the NAO is thought to explain a significant degree of interannual climate variability in the North Atlantic region (Kushnir and Wallace, 1989). The NAO is believed to be an expression of the hemisphere-wide Arctic Oscillation (Wallace, 2000).



(above) The difference in sea-level pressure between years with an NAO index >1.0 and those with an index <-1.0 since 1899. The contour increment is 2 mbar (*modified from Hurrell, 1995*).





The NAO index is usually based on the difference between the normalised sea-level pressure at Lisbon, Portugal and Stykkisholmur, Iceland (after Hurrell, 1995). An index of -1 indicates that pressure in Iceland is higher than in Lisbon. Since ~1980 the NAO index has been in a strongly positive phase with sea-level pressure in Iceland remaining low, whilst over the sub-tropical Atlantic pressure has been persistently high.

The uppermost graph shows a normalised NAO index since AD 1650. The index has been reconstructed from ice accumulation rates on the Greenland ice-sheet over the last 350 years (from Appenzeller et al., 1998). The data has been smoothed to remove high-frequency fluctuations.

The lower graph is the annual NAO (winter) index since 1865 based on sea-level measurements in Iceland and Portugal The dashed line represents the data smoothed with a low-pass filter (taken from Hurrell, 1995: Fig.1)

Periods of persistently low NAO index occurred between 1675-1690, 1800-1820 and 1850-1900 (according to Appenzeller et al., 1998). More recently, the 1960s were characterised by a low NAO index (Hurrell, 1995). Various workers have suggested that the NAO exhibits a periodicity operating at a frequency of around 8-14 years (eg. Rogers, 1984; Hurrell and Van Loon, 1997; Appenzeller *et al.*, 1998).

The North Atlantic Oscillation index has been reconstructed using various proxies (D'Arrigo *et al.*, 1994; Cook *et al.*, 1998; Appenzeller *et al.*, 1998). Most notably, ice-core data from Greenland (Barlow *et al.*, 1997) has been used to extend the NAO index back over 350 years (Fig. 2.1). Using carefully selected data from western Greenland, Appenzeller *et al.* (1998) have detected apparent periodicities in the NAO index of 5-7 and 12-14 years since AD 1650, with an 80–90-year cycle possibly in operation since the mid-19th century. These periodicities have all become more pronounced over the last 150 years. Other workers have also found evidence of periodicity in the NAO index, most notably operating at a frequency of 8-14 years (Rogers, 1984; Burroughs, 1994; Cook *et al.*, 1998; Hurrell and van Loon; 1997). Such behaviour, if supported, could have implications for predicting future hemispheric climatic change on a decadal scale.

Oceanic conditions

Surface ocean circulation

Ocean conditions largely determine the weather experienced in northwestern Europe, particularly the North Atlantic islands and the western fringe of Norway. Figure 2.2, modified from Krauss (1986), provides a schematic overview of sea circulation patterns within the North Atlantic Ocean. The numbers indicate the volumes of water transported in sverdrups (when, 1 sv = 10^6 m^3 s^{-1}). At around 40°N the *Gulf Stream* undergoes an abrupt change and bifurcates (Worthington, 1970; Rossby, 1996). Approximately half of the warm waters travel eastwards as the *Azores Current* and eventually re-circulate, whilst the remaining current flows northeast towards the mid-Atlantic ridge as the *North Atlantic Current.* At around 50°N this current bifurcates again sending an arm of warm water (c. 20 Sv) towards western Iceland whilst the remainder (c. 10 Sv) of the North Atlantic Current proceeds north and east, passing between northwest Scotland and the Faeroe Islands, before reaching the Norwegian Sea. Approximately one-tenth of the volume of warm Gulf Stream water that passes Cape Hatteras (35°N) reaches the coast of Norway (65°N) (Rossby, 1996; Sutton and Allen, 1997) (Fig. 2.2). These warm sub-tropical waters take between 9 and 12 months to travel from the coast of Florida to the west coast of Ireland, as surface buoy observations have confirmed (Bacon, 1998; Schmitz and

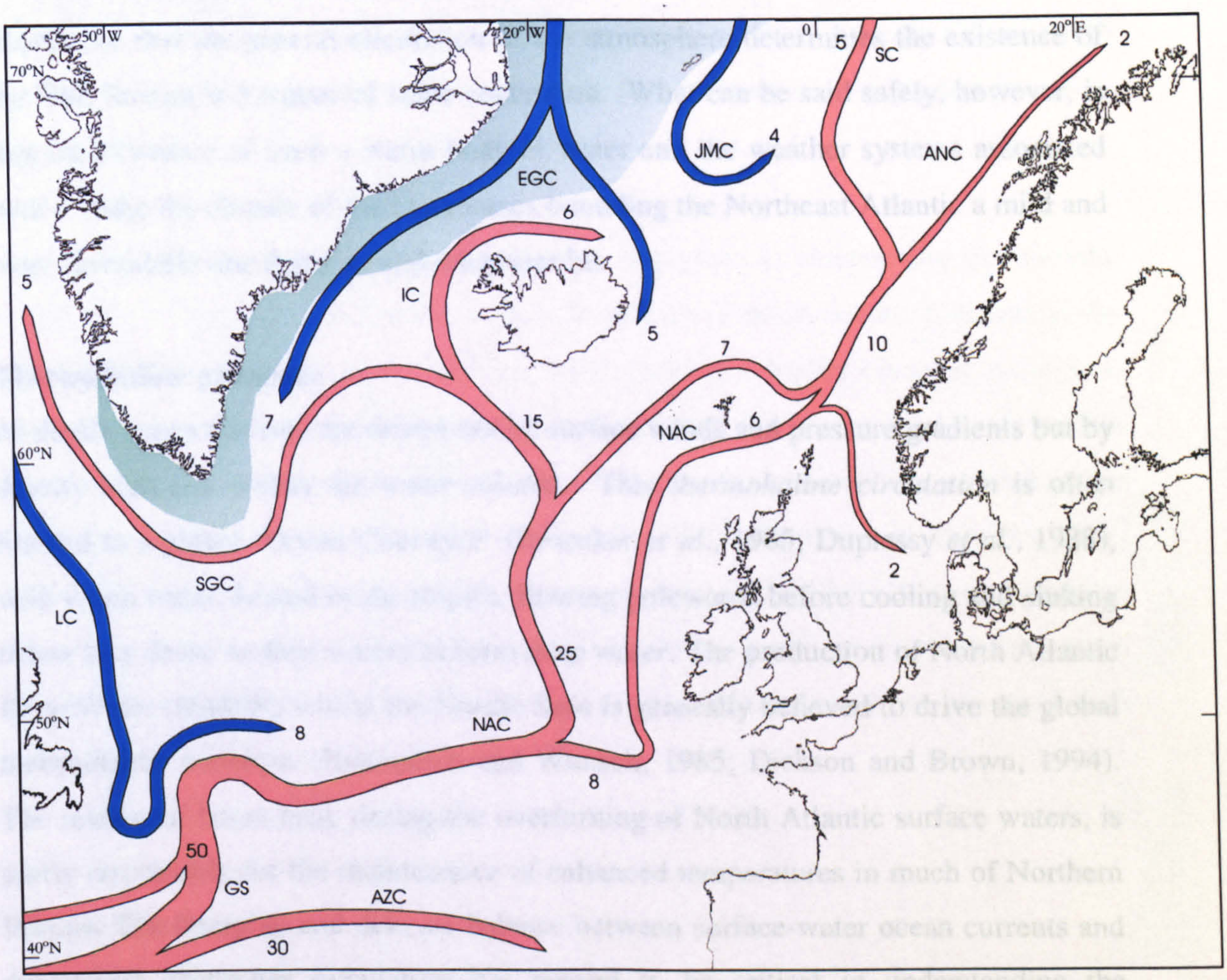


Figure 2.2 Major North Atlantic surface ocean currents. Numbers relate to the approximate flow in Sverdrups (1sv = 10^6 ms^{-1}). The average winter extent of Arctic sea-ice (1930-1960) is shaded pale blue. (Modified from Krauss; Fig.1., 1986.)

Key to ocean currents

Waters of mid-latitude origin (ie. Warm)
- GS - Gulf Stream
- AZC - Azores Current
- NAC - North Atlantic Current
- IC - Irminger Current
- SGC - South Greenland Current
- ANC - Atlantic Norwegian Current
- SC - Svalbard Current

Waters of high-latitude origin (ie. Cool)
- EGC - East Greenland Current
- JMC - Jan Mayen Current
- LC - Labrador Current

McCartney, 1993). The heat transported by this current however may take considerably longer (12-14 years) as the energy is believed to propagate at depth by advection (Sutton and Allen, 1997).

Whether the presence of the Gulf Stream determines the climate of Northwest Europe, or that the general circulation of the atmosphere determines the existence of the Gulf Stream is a matter of some contention. What can be said safely, however, is that the existence of such a warm body of water and the weather systems associated with it make the climate of the landmasses bounding the Northeast Atlantic a mild and more favourable one that it would otherwise be.

Thermohaline processes

At depth, ocean currents are driven not by surface winds and pressure gradients but by density contrasts within the water column. This *thermohaline circulation* is often likened to a global 'Ocean Conveyor' (Broecker *et al.*, 1985; Duplessy *et al.*, 1988), with warm water, heated in the tropics, flowing polewards before cooling and sinking below less dense surface waters to form *deep water*. The production of North Atlantic Deep Water (NADW) within the Nordic Seas is generally believed to drive the global thermohaline conveyer (Roemmich and Wunsch, 1985; Dickson and Brown, 1994). The release of latent heat, during the overturning of North Atlantic surface waters, is partly responsible for the maintenance of enhanced temperatures in much of Northern Europe. The interplay and delicate balance between surface-water ocean currents and convective deepwater circulation has proved to be critical in understanding the physical oceanography and climatology of the Northeast Atlantic.

Certain cooling events, which commenced abruptly approximately 13,500, 12,000, 11,000 and 8200 years BP, have been attributed to the partial or total cessation of the North Atlantic thermohaline circulation (Broecker; 1990; Bond *et al.*, 1997; Curry and Oppo, 1997; Arz *et al.*, 2001). The theory gained popularity when thermohaline 'shutdown' was linked with the massive freshwater discharge resulting from the melting of the Laurentide ice sheet and the subsequent drainage of several, large, North American pro-glacial lakes (eg. Rooth, 1982; Broecker and Kennet, 1989; Broecker, 1990; Vidal *et al.*, 1997).

Until recently it was commonly believed that the North Atlantic Ocean had only one stable mode of operation, all other hypothetical scenarios being quasi-stable and prone to rapid breakdown. Workers have since shown multiple stable modes of ocean circulation have occurred during the last 20,000 years (Bryan, 1986; Manabe and Stouffer, 1988), with research now suggesting that the North Atlantic Ocean could have as many as three or four stable, yet markedly, different circulation patterns (Koç *et al.*, 1993).

Whether the ocean or the atmosphere is most important in determining the climatic conditions of Northwestern Europe is still the subject of much debate. Ultimately, the relationship between the oceans and the atmosphere is a highly complex symbiosis. The two are inextricably linked. In order to understand the processes at work models must incorporate both realms. These complex, coupled mathematical models have so far yielded divergent results (eg. Delworth *et al.*, 1993; Manabe and Stouffer, 1995, 2000; Winton, 1997). No consensus has been reached on the likely degree of climatic change in a world with warmer oceans. These problems highlight the need for a more detailed empirical record of past climatic and oceanic fluctuations from key areas such as the North Atlantic.

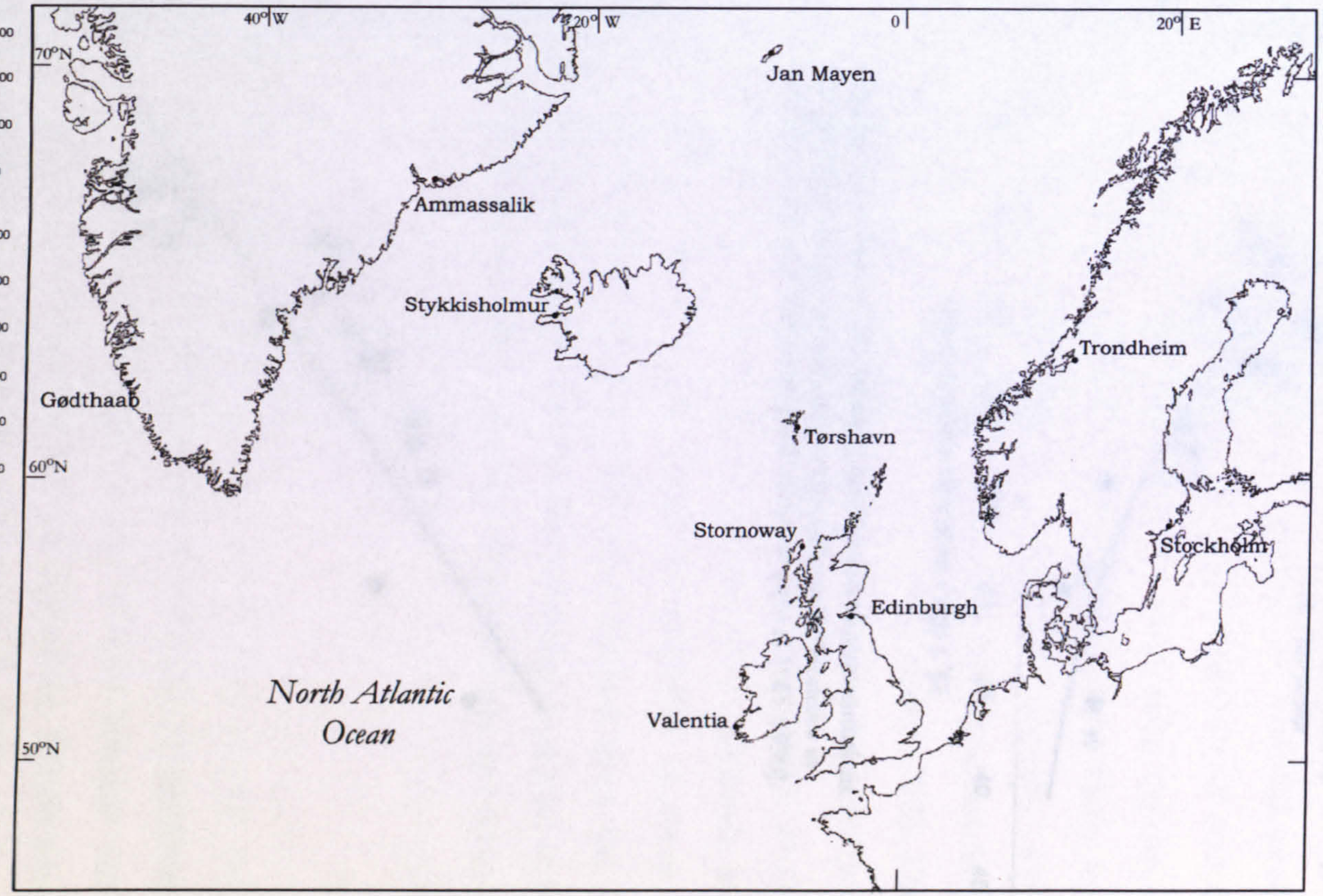
The present-day climate

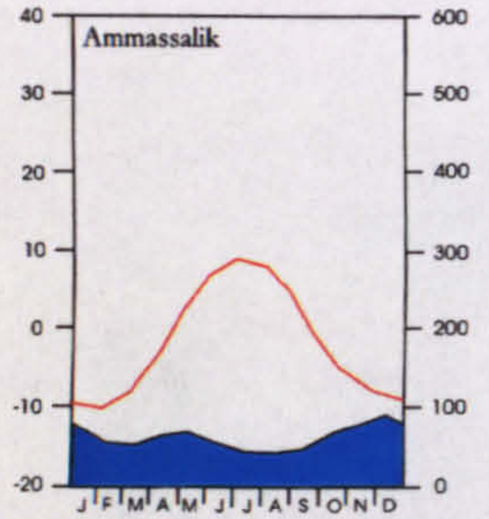
Figure 2.3 shows the typical climatic conditions (air temperature and precipitation) currently experienced at various localities around the Northeast Atlantic region. The data refer to the 30-year reference period from 1961 to 1990.

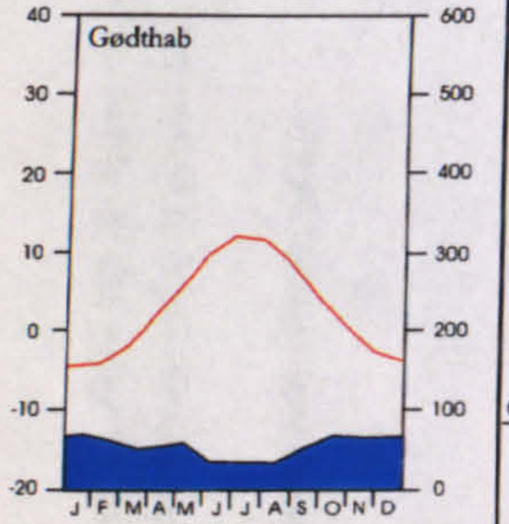
Continentality and Oceanicity

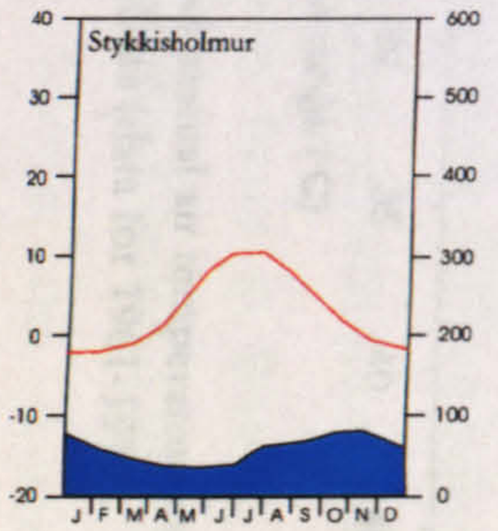
A feature of oceanic climates is a mild year-round climate, often coupled with relatively high amounts of precipitation. This is well demonstrated in the climate graphs from Valentia, Tørshavn and Stornoway (Fig. 2.3). In order to explore the relationship between annual temperature range and rainfall more fully, data from nine of the ten reference stations (excluding Jan Mayen due to uncertainties surrounding the precipitation record), plus a further ten from around Northern Europe and

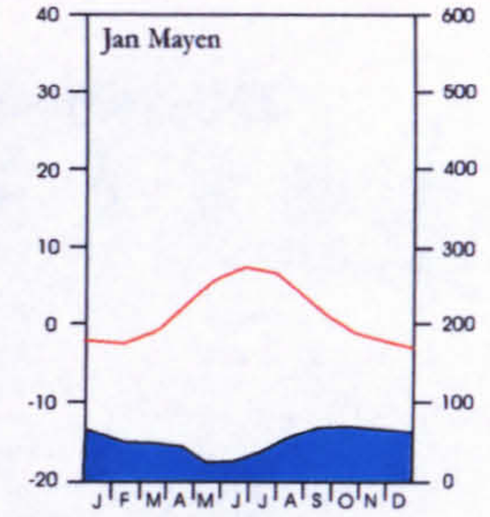
Figure 2.3 The present-day climate of the Northeast Atlantic (1961-1990)

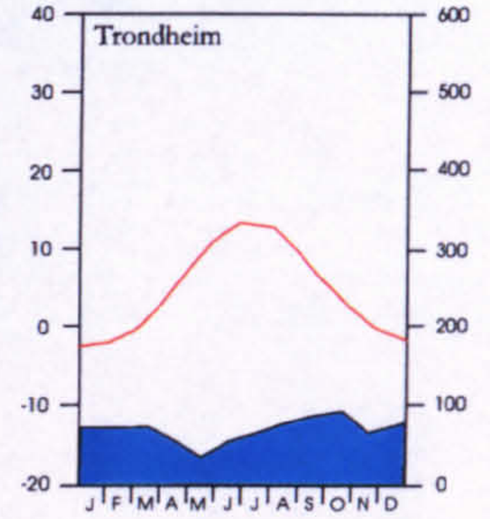


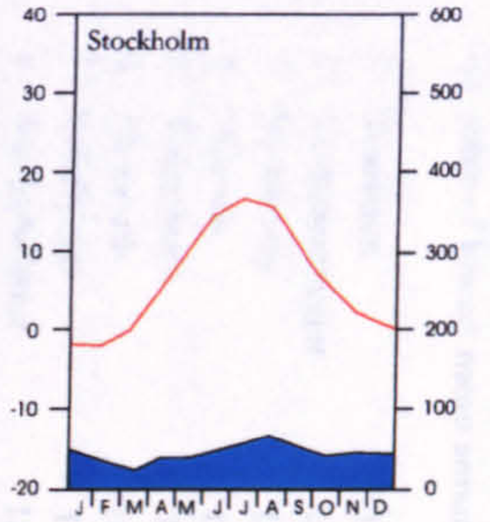


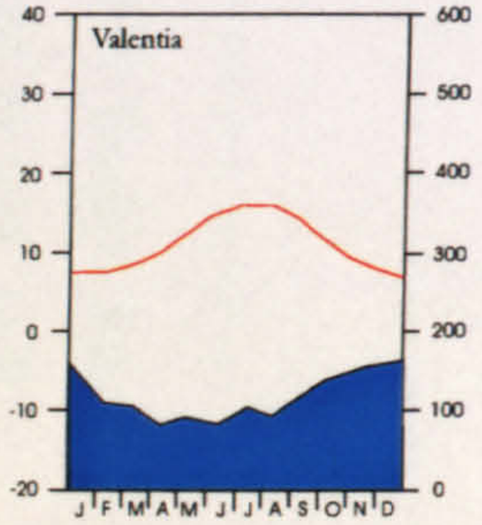


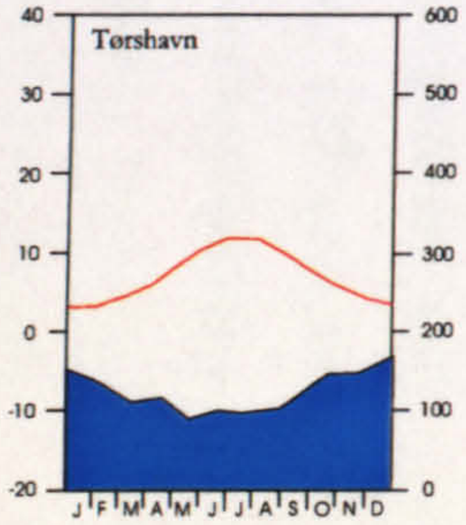


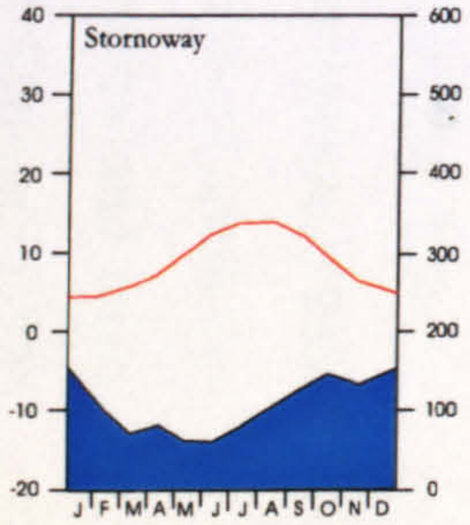


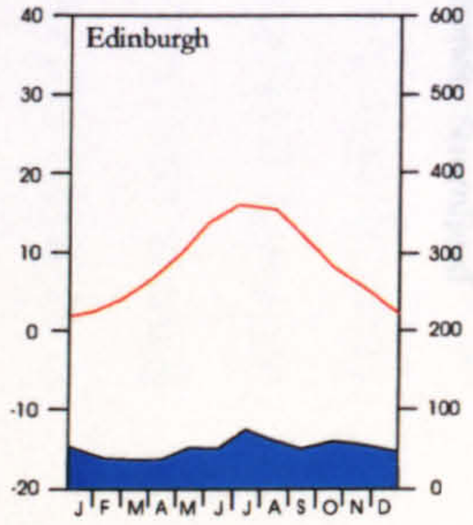












Average climatic conditions at 10 localities in the Northeast Atlantic region. Mean monthly air temperatures (red) in °C. Precipitation in mm (blue). Figures are for the 30-year reference period (1961-1990).

Data taken from the NOAA climatic database.

Key to weather stations:
(in order of lowest mean annual temperature range)

1 Torshavn
2 Vestmannaeyjar
3 Stornoway
4 Valentia
5 Teigarhorn
6 Plymouth
7 Edinburgh
8 Stykkisholmur
9 Aberdeen (Dyce)
10 Grimsstadir
11 Trondheim
12 Ammassalik
13 Godthaab
14 Hamburg
15 Warszawa
16 Stockholm
17 Jakobshavn
18 Danmarkshavn
19 Sondre Stromfjord

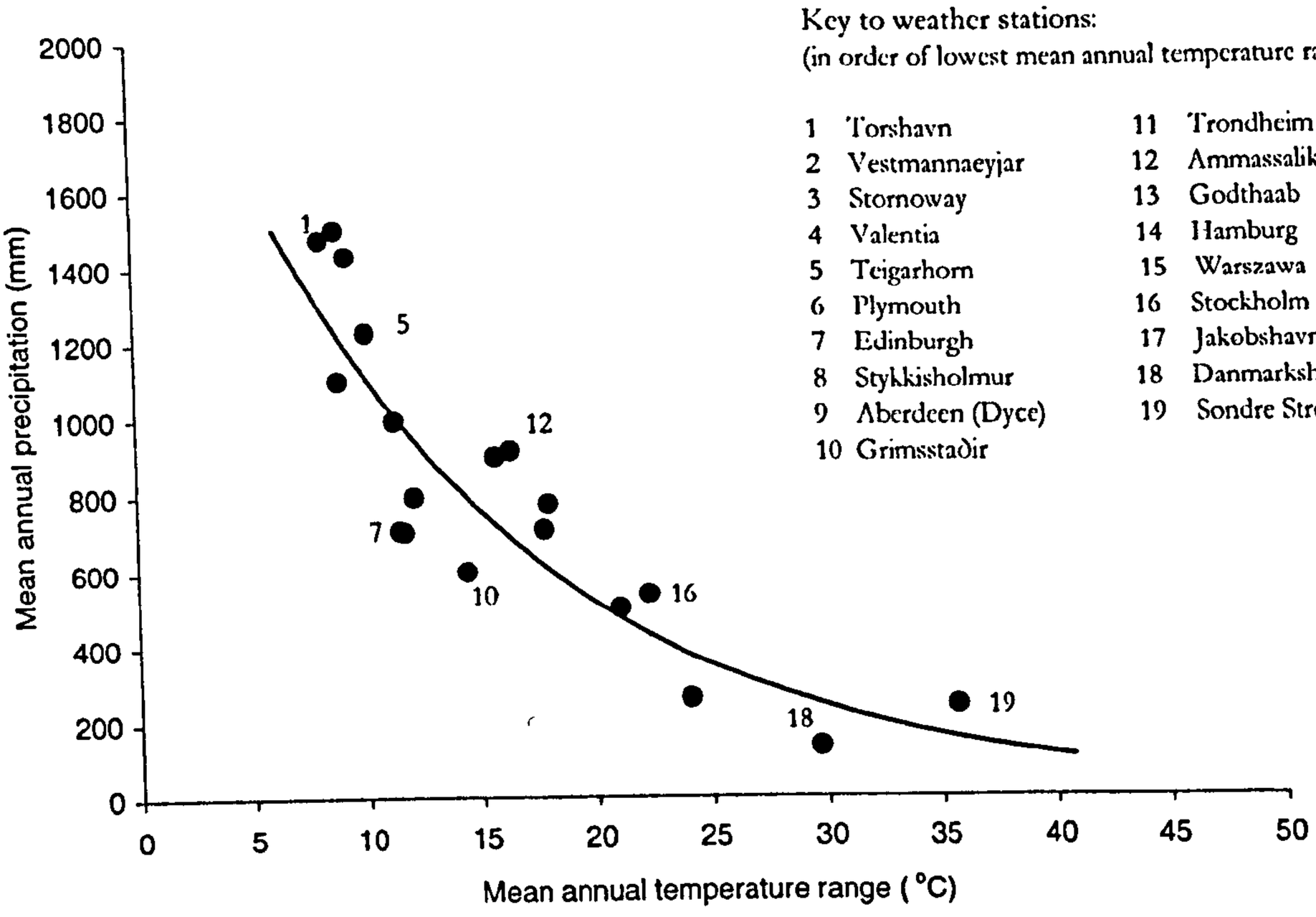


Figure 2.4 (above) The relationship between mean annual air temperature range and annual precipitation at 19 localities in Northern Europe and Greenland (data for period 1961-1990).

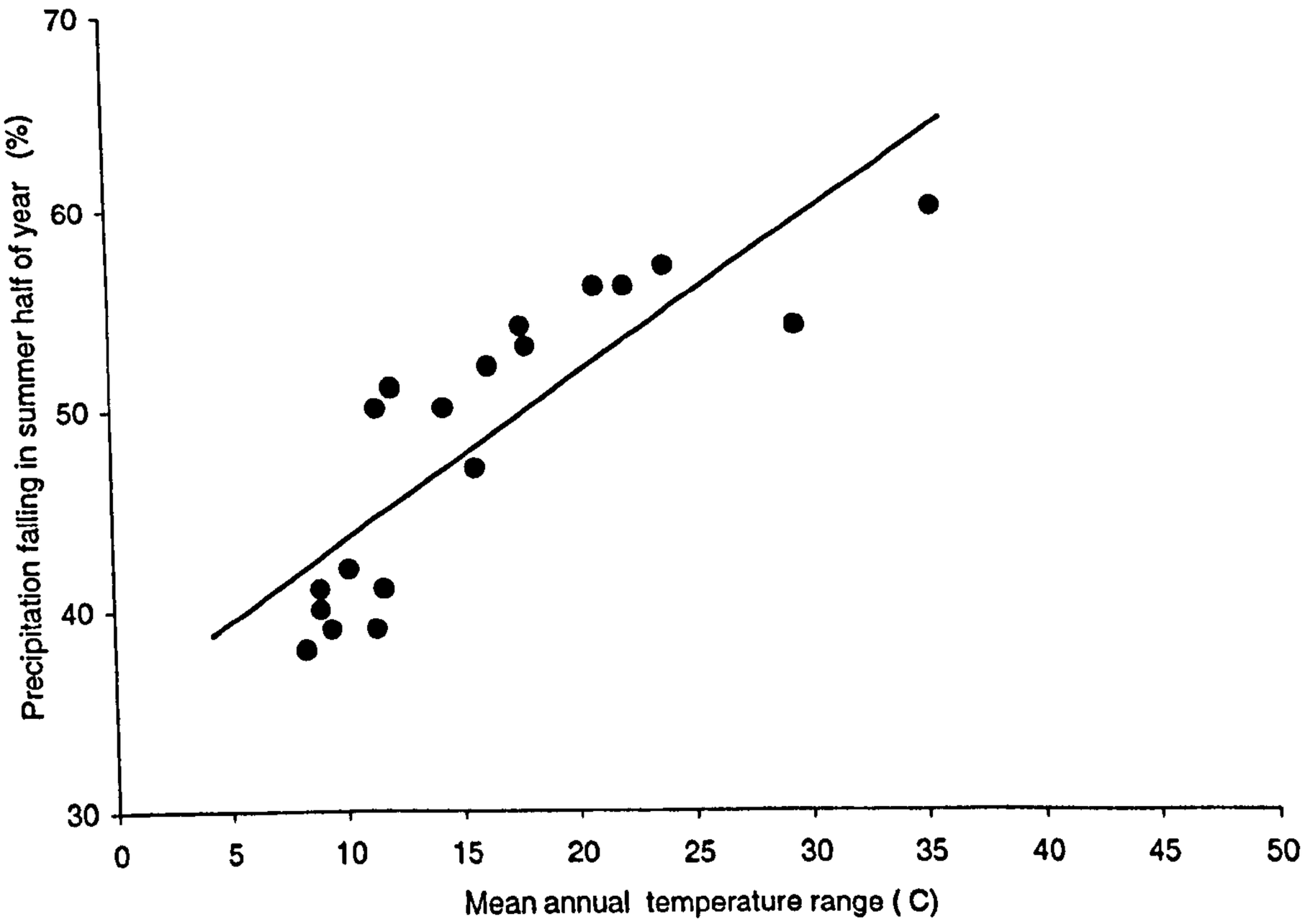


Figure 2.5 The relationship between mean annual air temperature range and rainfall seasonality at the same 19 localities (data for 1961-1990).

Greenland, were analysed. The resulting graph (Fig. 2.4) shows that coastal localities, experiencing maritime climates, are characterised by a low mean annual temperature range and relatively high rainfall. Inland sites, at some distance from the open Atlantic Ocean, such as Stockholm, Warsaw and Søndre Strømfjørd in Greenland, experience a much greater annual temperature range and receive notably less rainfall. Figure 2.4 demonstrates that at Northeast Atlantic sites where the annual temperature range is high (>20°C), precipitation values are generally low (<500 mm/yr), whereas at sites with a low annual temperature range (<11.5°C) rainfall totals are significantly higher, typically over 1000 mm/yr. A further relationship is also observed between temperature range and rainfall seasonality. In continental regions, where the annual temperature range is high (>20°C), precipitation is more likely to be confined to a certain season (ie. summer), whilst in maritime regions, with a low annual temperature range, rainfall is more evenly distributed throughout the year (Fig. 2.5).

These relationships have long been known (eg. Koppen, 1923, Thornthwaite, 1933; Trewartha, 1954; Barry and Chorley, 1992). However the classification of spatially-variable climate in terms of 'continentality' and 'oceanicity' is complex and lacks standard definition (eg. Conrad, 1946; Oliver, 1970; McBoyle and Steiner, 1972; Barry and Chorley, 1992). The temporal variability of these temperature and precipitation relationships, however, has been studied since the late 18th century. Knowledge of the weather over the last few centuries is crucial in order to identify – and ultimately better understand – climatic change over longer timescales.

The climate since AD 1750

During the last two-and-a-half centuries the climate of northwestern Europe has undergone a change. Arguably, the cause of this climatic change is less well understood than those at the end of the last glaciation, 10,000 years ago. Although detailed meteorological measurements exist from the 18th century they are rare and restricted geographically. Of all the meteorological records from before AD 1750 only a handful are found outside Europe – namely those in Boston, New Haven and Philadelphia on the eastern coast of North America. Consequently, owing to the Euro-centric distribution of long instrumental records, the potential for studying

hemispheric or global climatic change over more than the last 100 years is somewhat limited. The relative wealth of historical data from Northern Europe and the Northeast Atlantic region allow the greatest potential for the study of instrumentally recorded climate change anywhere on Earth.

It has been suggested that the constantly fluctuating weather of the European climate is more than partly responsible for the origination of meteorology as a science. The invention of the thermometer by Galileo in 1611 was obviously fundamental to the establishment of a climatological network. Without properly calibrated equipment weather monitoring was effectively meaningless. Several Commissions were set up in Europe during the 17th and 18th centuries which encouraged the collection of daily weather observations: *Academia del Cimento* (1654) in Florence, *The Royal Society of London* (1690), The Astronomical Observatory in St. Petersburg (1751) and *The Societé Royale de Medécine* in Paris (1778). It was not until 1781, however, that the first international network was instigated with meteorological observations being published by the *Society of Mannheim*, in Germany. Although the network collapsed ten years later, many other major European cities established permanent meteorological observatories around this time (eg. Edinburgh, 1764, Prague 1775, Vienna 1775, and Warsaw, 1779), with instrumental temperature and precipitation records remaining relatively uninterrupted to the present day. The continuity and duration of these long records enabled Wilhelm Koppen in 1873 to assess whether temperatures had changed since 1750. Similarly, twenty-first century climatologists are now able to examine the evidence for climatic change from over 250 years of instrumental records.

Figure 2.6 presents homogenised temperature records for the ten weather stations featured in Figure 2.3. These localities, chosen for their geographical diversity (ranging from 71° to 52°N and 19°E to 51°W) also span the whole of the last 250 years, with three stations capturing observations from the second half of the 18th century (Stockholm 1756–, Trondheim 1762– and Edinburgh 1764–).

The selected weather stations:

Location	*Lat., Long.*	*Altitude (m)*	*Country*	*First observation*
Stockholm	59°21'N, 18°04'W	45	Sweden	1756
Trondheim	63°25'N, 10°27'W	133	Norway	1761
Edinburgh	55°55'N, 3°11'W	76	Scotland	1784
Stornoway	58°11'N, 6°21'W	10	Scotland	1823
Stykkisholmur	65°05'N, 22°46'W	25	Iceland	1845
Valentia	51°56'N, 10°15'W	9	Ireland	1861
Tørshavn	62°02'N, 6°45'W	24	Faeroe	1867
Gødthaab	64°12'N, 51°45W	50	Greenland	1866
Ammassalik	65°37'N, 37°33'W	32	Greenland	1895
Jan Mayen	71°01'N, 8°25'W	12	Norway	1912

There are several important features regarding the instrumental temperature records from around the eastern North Atlantic. If factors such as increasing urbanisation and weather station relocations are ignored – both known to affect the long-term homogeneity of instrumental records (Bradley *et al.*, 1985; Jones *et al.*, 1986) – certain common trends can be identified. Firstly, all those series spanning more than 100 years show an overall warming. The degree of warming since ~1870 is most pronounced at Gødthaab and Stykkisholmur and least noticeable at Trondheim, Tørshavn and Valentia. The coldest periods occurred during the 1780s, between 1800-1815, 1860-1890 and around 1915-1920, with temperatures at high latitudes falling most significantly. Secondly, all stations (with the exception of Jan Mayen) recorded a sharp rise in temperature during the period 1920-1940. Again, this phenomenon is most marked at higher latitudes. Further cold spells occurred in the 1960s and the 1980s. Finally, since the 1950s temperatures have fallen generally across the Northeast Atlantic, although there is some suggestion of renewed warming after ~1990. This data, gathered from calibrated meteorological instruments, highlights the 'natural', decadal-scale, variability within the climate of the Northeast Atlantic over the last three centuries.

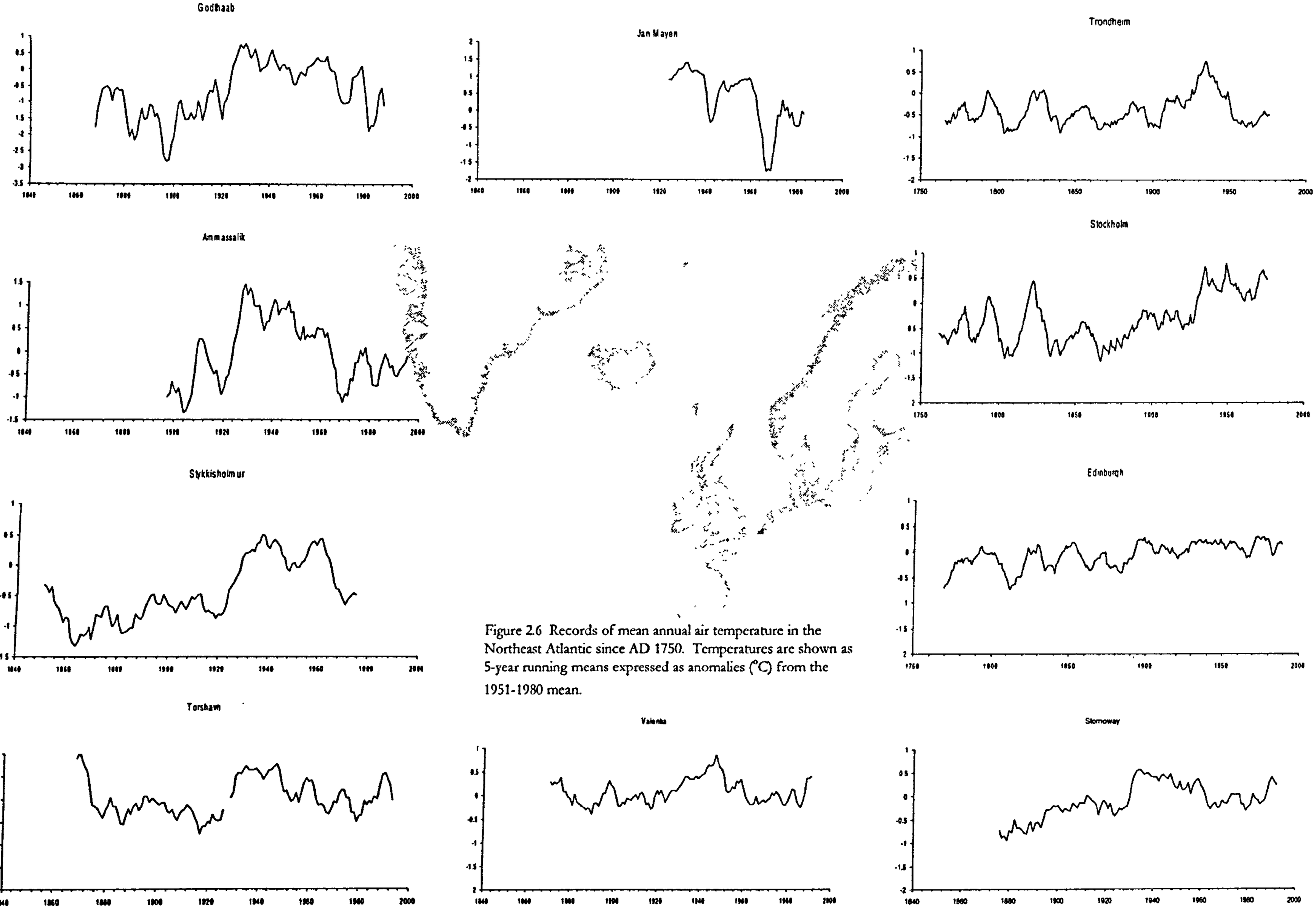


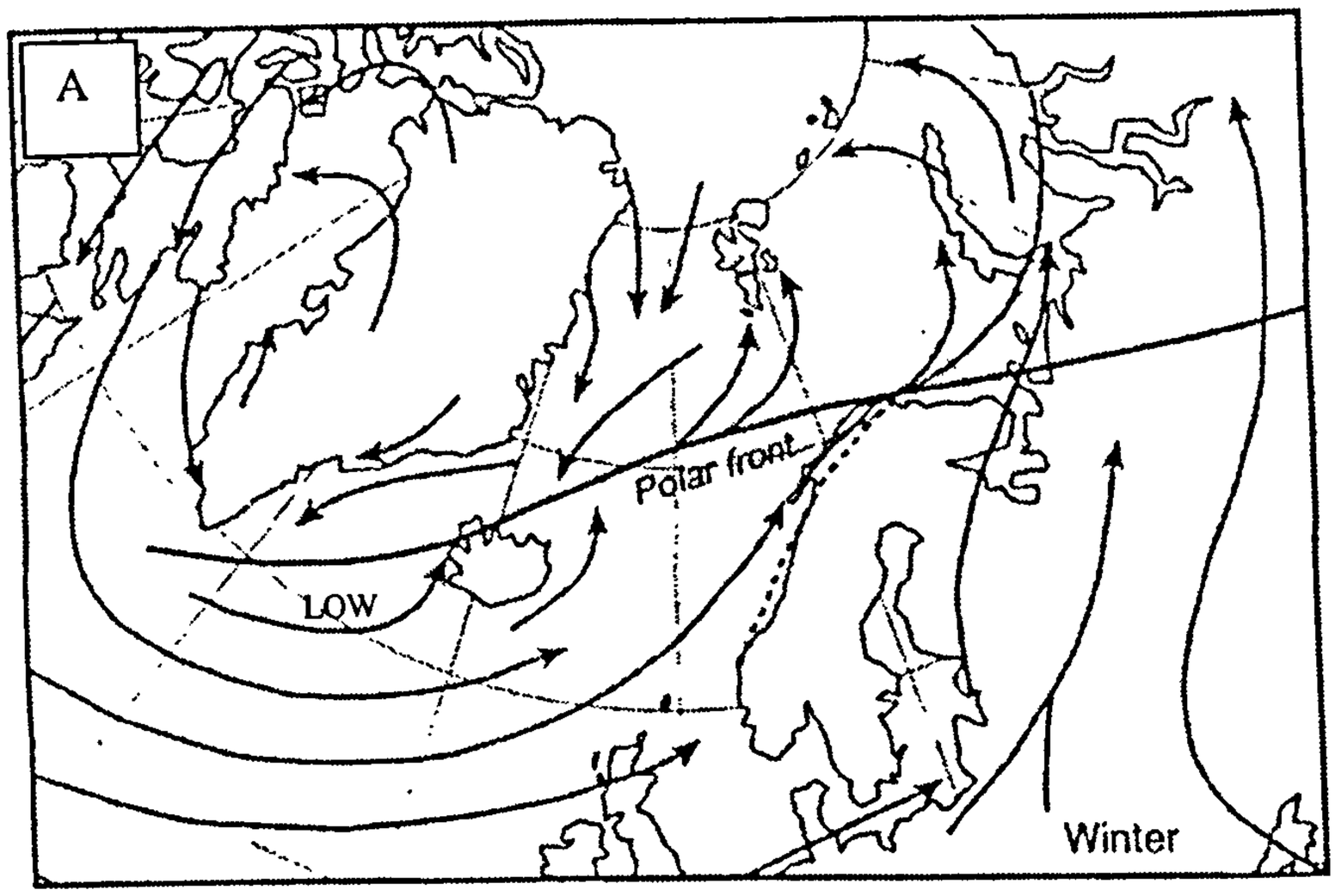
Figure 2.6 Records of mean annual air temperature in the Northeast Atlantic since AD 1750. Temperatures are shown as 5-year running means expressed as anomalies (°C) from the 1951-1980 mean.

B : The climate of Iceland

Atmospheric circulation and the present-day climate of Iceland

Iceland's weather is fundamentally determined by the island's location in the eastern North Atlantic. Situated 64° to 66.5°N, and approximately midway between the land masses of Europe and Greenland, Iceland lies within the convergence zone between cold polar and warm temperate air masses (Fig. 2.7). The Greenland ice-sheet normally sustains a high-pressure system which is often associated with the presence of a strong cyclonic system, the Icelandic Low. This pressure gradient drives the winds across Iceland and, in part, determines the position of the convergence zone known as the Arctic Front. Fluctuations of the Arctic Front occur on a seasonal, annual and decadal scale (Lamb, 1979). Associated changes in the strength and location of the convergence zone are thought to be a primary cause of short-term (<20 years) climatic oscillations seen over Iceland and the Northeast Atlantic region as a whole.

The weather in Iceland is cool, maritime and extremely variable. The Arctic Front marks the main path of North Atlantic storms; hence Iceland is frequently subjected to cyclonic weather systems. Rapid changes in temperature are common as warm and cold air masses are in constant competition. The island can be sub-divided into two climatic provinces: north and central Iceland, and south and western Iceland. The central highlands of Iceland rising 500-700 masl act as a barrier to weather moving across the island. As the bulk of Iceland's precipitation originates in the North Atlantic to the south of the island, the northern parts of Iceland receive much less. The precipitation-shadow effect is most notable in the northeast of the island where totals are normally only one-tenth that of the ice-capped southern districts (Fig. 2.8). In the south of Iceland along the coasts, temperatures remain above freezing all year round. The mildest winter temperatures are on the Westman Islands and at Vík. In the north, the mean January temperature at the coast ranges from -1°C in Þorustaðir, in the northwest, to -2°C in Akureyri (Fig. 2.8). Summers are normally 2°C warmer along the south coast (11°C) than the north. Mean annual temperatures in the south-coast



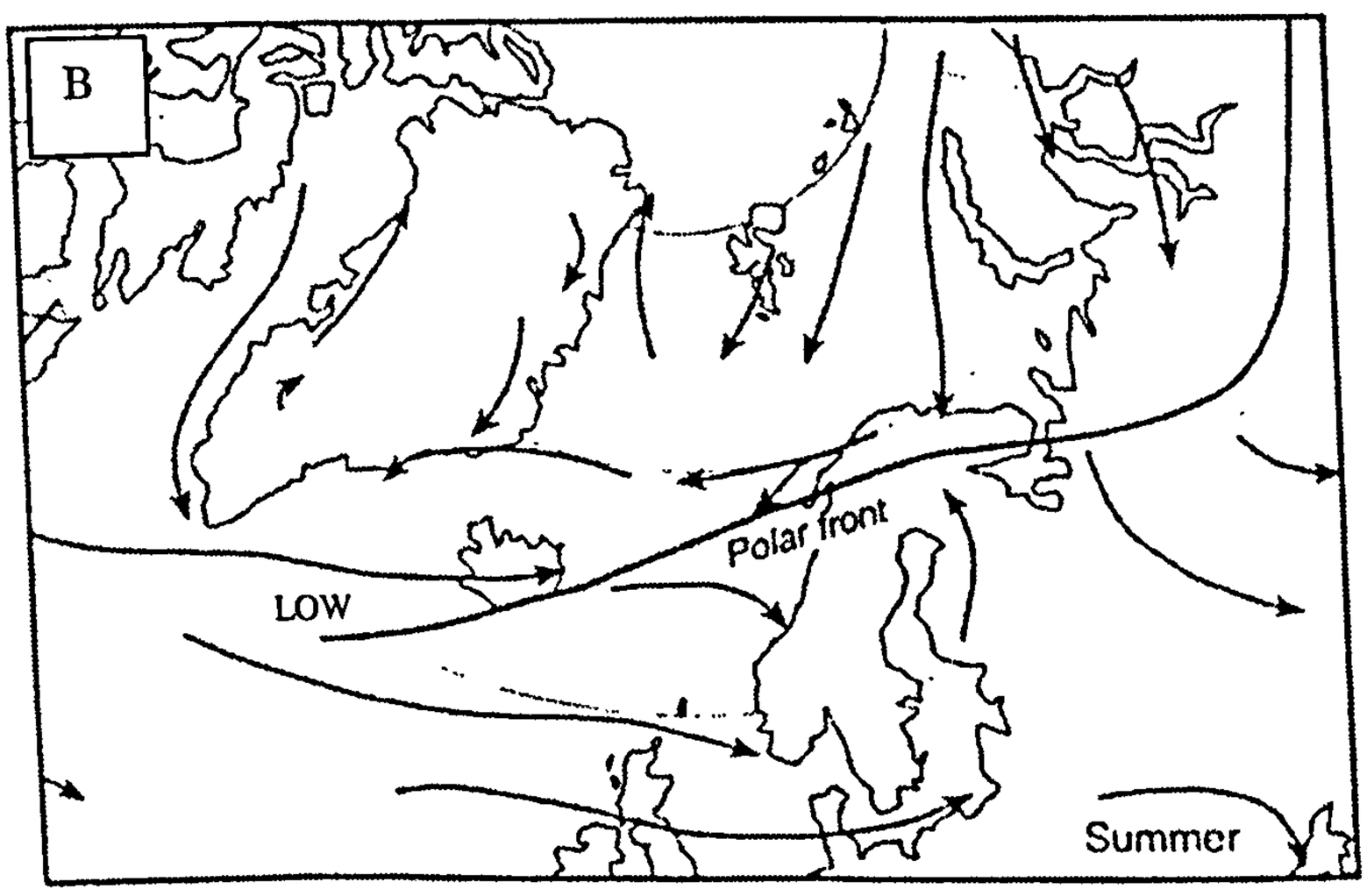


Figure 2.7 The position of the Atmospheric Polar Front in the North Atlantic.

(a) The Polar Front marks the convergence zone between warm air of low-latitude origin and cool air from the Arctic.

(b) During summer, the Polar Front normally migrates southwards across Iceland as the strength of the Polar vortex increases. The quasi-stationary *Icelandic Low* is located SW of Iceland (marked LOW). During positive phases of the North Atlantic Oscillation this depression is more vigorous and storms are more common.

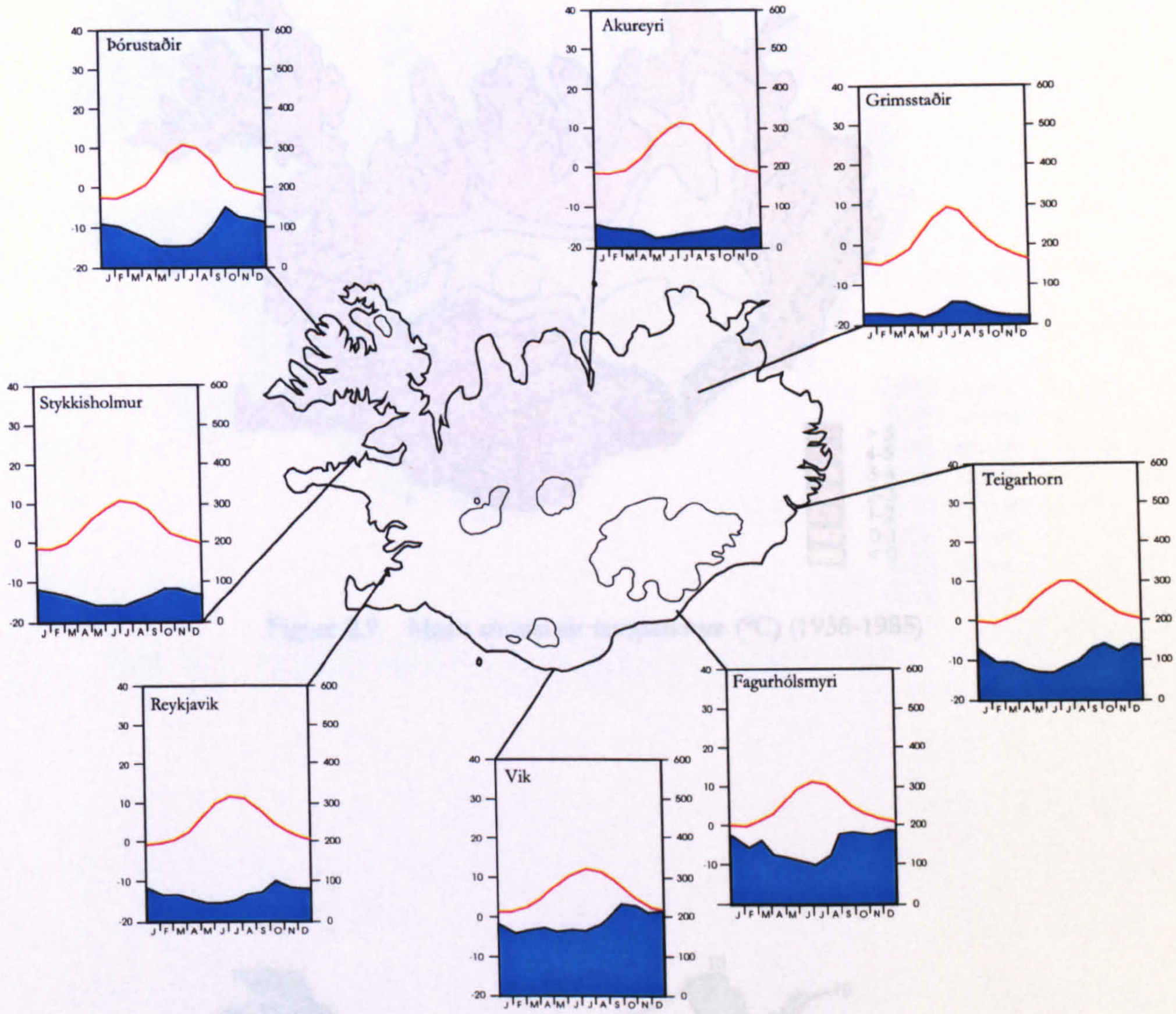


Figure 2.8 Climatic conditions in Iceland (1931-1960).

Mean monthly temperature in degrees Celsius (red line) and precipitation in millimetres (blue area), for eight stations in Iceland during the period 1930-1961.

Seven of the eight weather stations are located on the coast and are less than 50 m above sea level. The contrast between the maritime, mild, wetter conditions in the south and east and the cooler and drier north and west can be seen most clearly in the records of Vík and Akureyri.

Grimsstaðir (385 m asl) is more representative of climatic conditions inland with little precipitation (350 mm) and a greater mean annual temperature range (15°C).

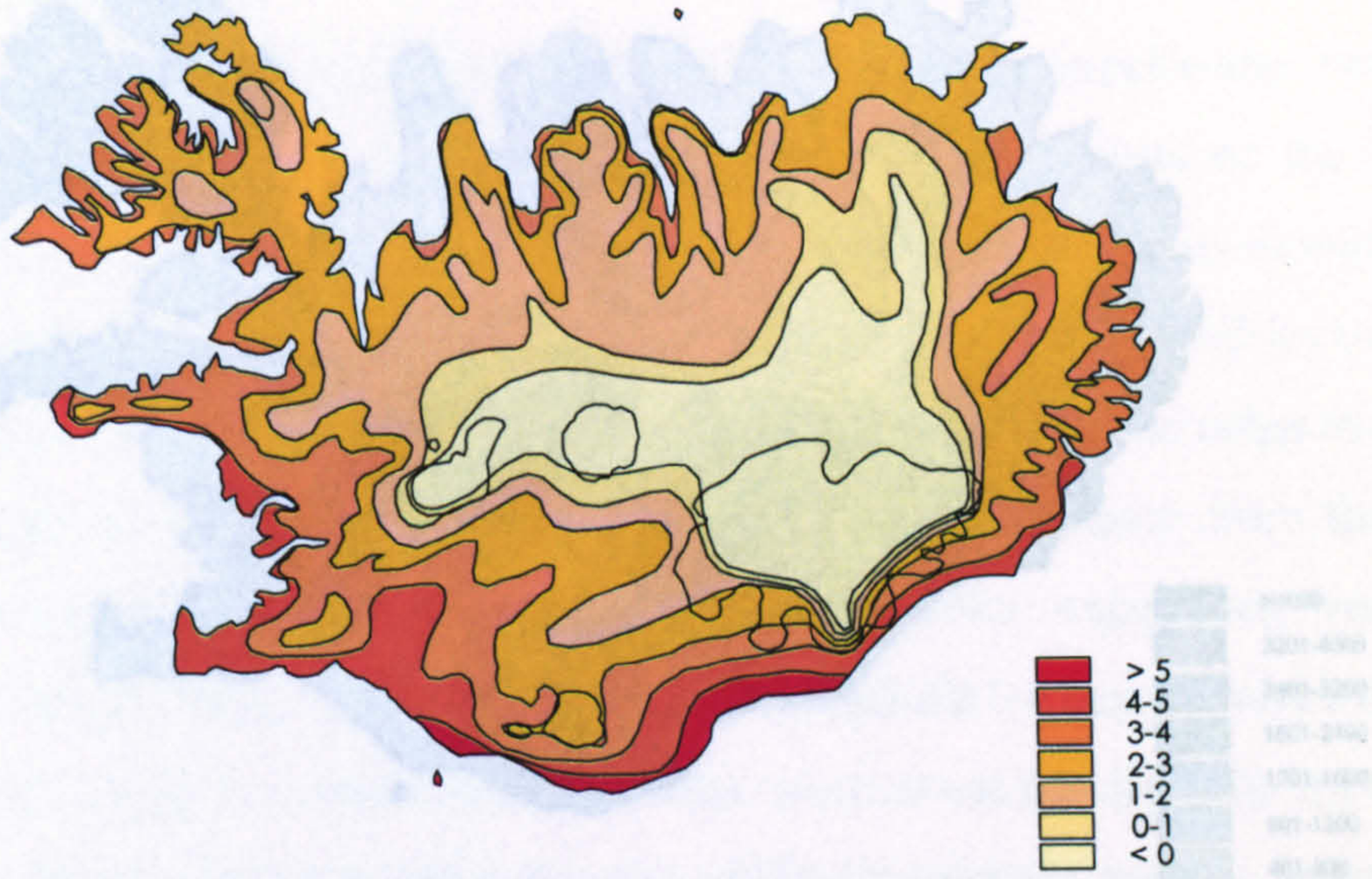


Figure 2.9 Mean annual air temperature (°C) (1936-1985)

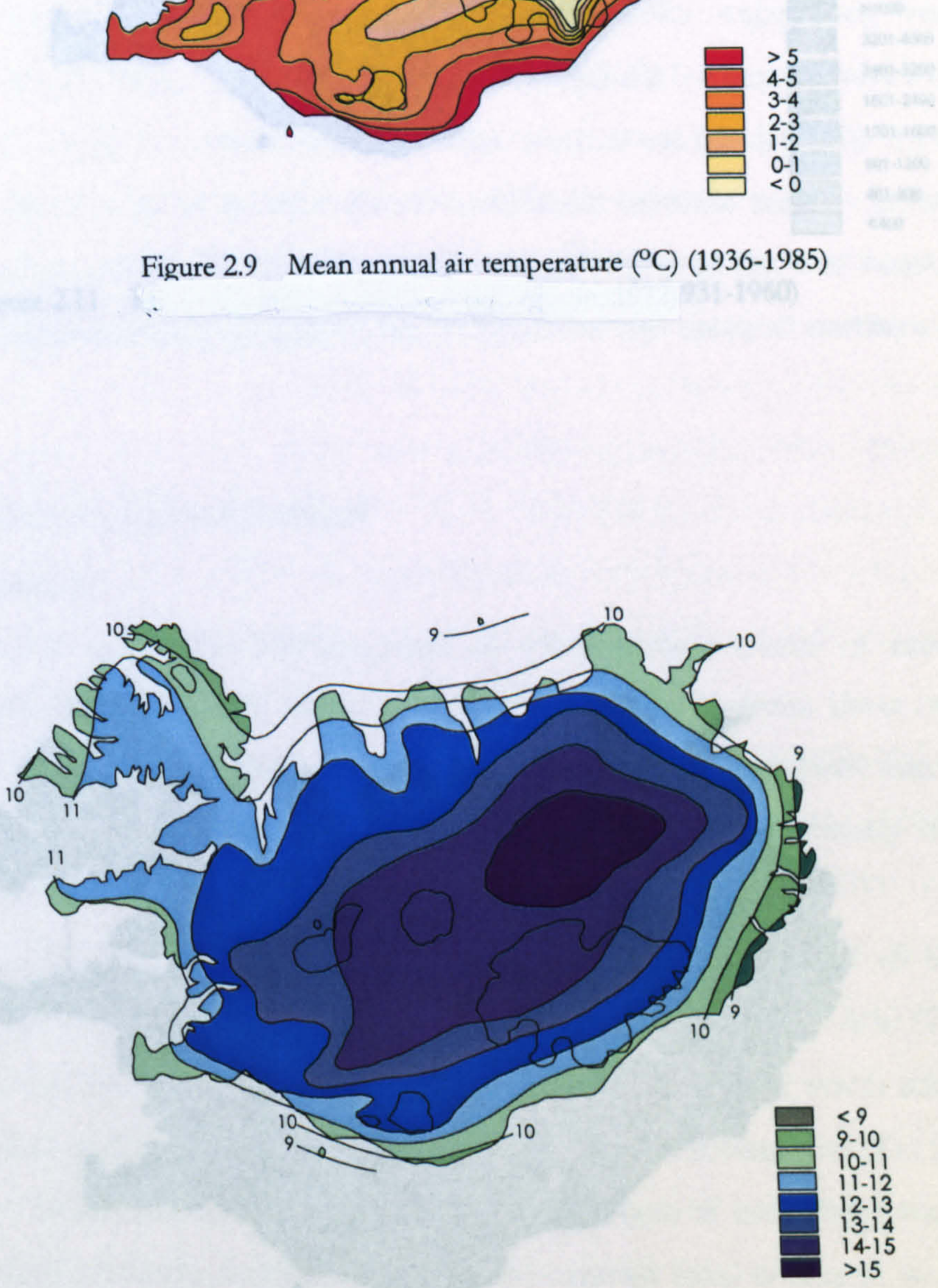


Figure 2.10 Mean annual air temperature range (°C) (1936-1985)

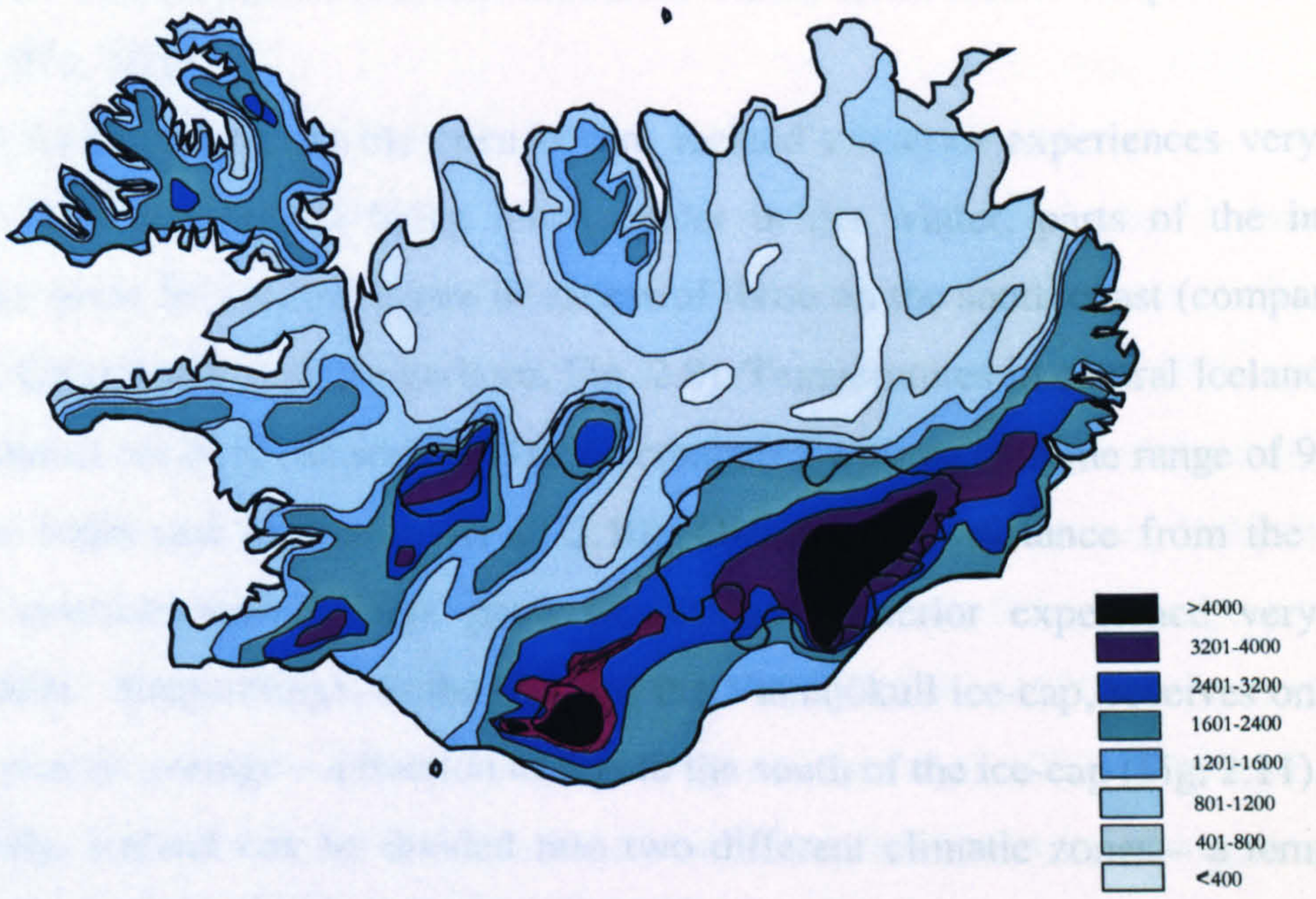


Figure 2.11 Precipitation distribution in Iceland (mm a^{-1}) (1931-1960)

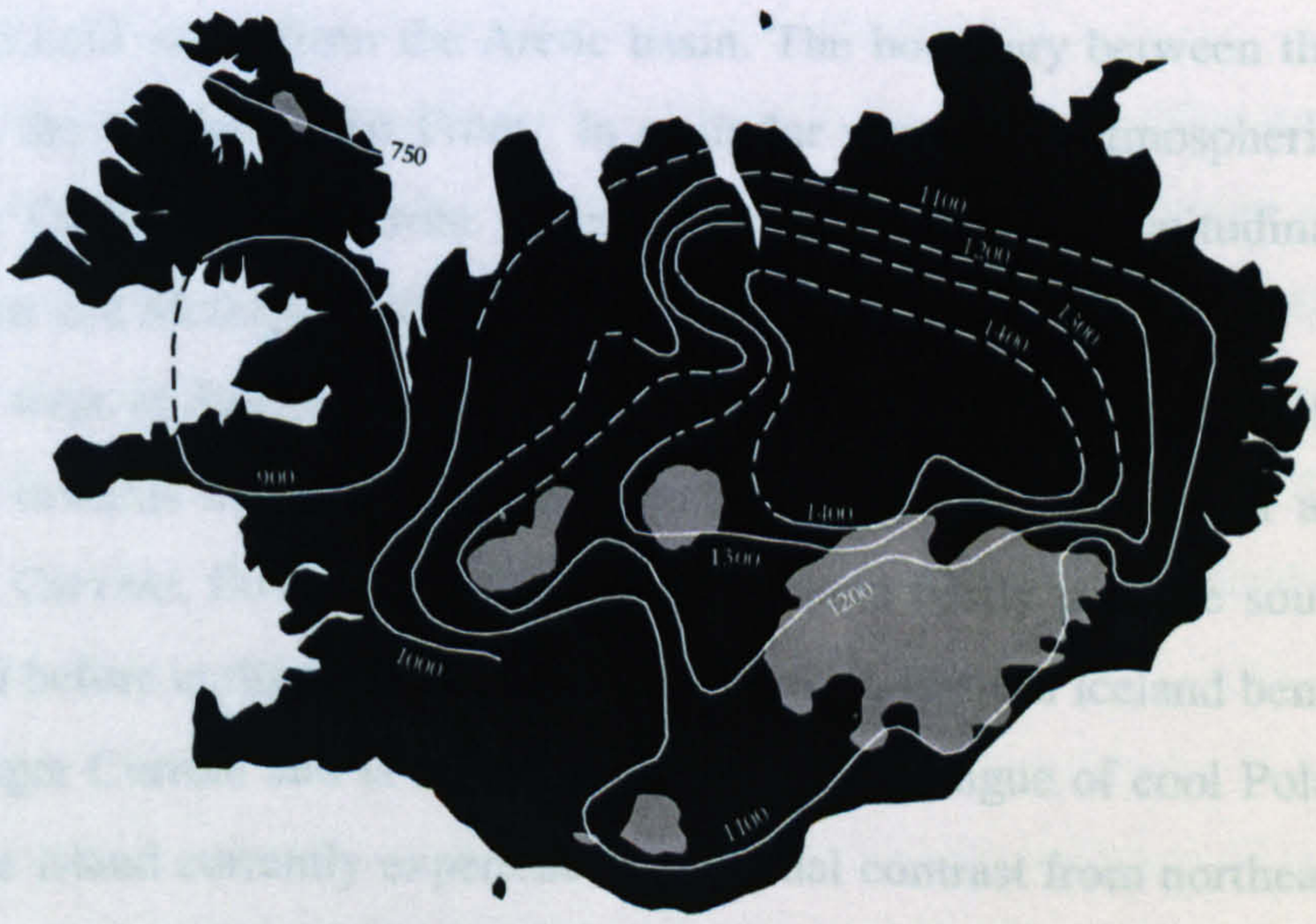


Figure 2.12 Snowline altitudes in Iceland (m asl) (modified from Ahlmann and Thorarinsson, 1937)

lowlands are around 5°C – similar to those in Shetland – whilst in the extreme northeast of Iceland, 20 km from the Arctic Circle, the mean annual temperature is less than 3°C (Fig. 2.9).

Due to its distance from the open ocean, Iceland's interior experiences very little maritime influence. Whilst being much cooler in the winter, parts of the interior experience mean July temperatures in excess of those on the south coast (compare, for example, Grimsstaðir with Teigarhorn, Fig. 2.8). Temperatures in central Iceland have a mean annual range of between 13-16 °C, compared with an extreme range of 9-10°C along the south and east coasts (Fig. 2.10). Owing to the distance from the North Atlantic moisture-source, large parts of Iceland's interior experience very little precipitation. Krepputunga, to the north of the Vatnajökull ice-cap, receives only 300 mm per year on average – a fraction of that to the south of the ice-cap (Fig. 2.11).

Generally, Iceland can be divided into two different climatic zones:– a temperate, wet, maritime zone stretching from the northwest peninsula to the east coast; and an arid, sub-polar zone encompassing the central highlands and much of northern Iceland.

Ocean circulation around Iceland

Surface conditions

Presently, Iceland marks the meeting point of warm surface waters of sub-tropical origin and cold water from the Arctic basin. The boundary between these two water masses is the Oceanic Polar Front. In a similar way to the atmospheric convergence zone, the Oceanic Polar Front is known to have migrated latitudinally over time (Ruddiman and McIntyre, 1979)

To the west of Iceland the North Atlantic Current splits, with half (6 Sv) being deflected towards the east coast of Greenland (Fig. 2.2). The other arm (7 Sv), the *Irminger Current*, flows clockwise around Iceland firstly past the south and west of the island before curling around the northern coast. Eastern Iceland benefits least from the Irminger Current and is normally affected by a tongue of cool Polar waters. As a result, the island currently experiences a thermal contrast from northeast to southwest with the south and west coasts being warmed by North Atlantic waters, whilst the north and east coasts are cooled by a branch of the *East Greenland Current*. The

position of this cold water current around Iceland's eastern shores influences the weather conditions experienced in nearby coastal regions.

Deepwater conditions

The flow of warm, saline Atlantic waters past south and western Iceland is balanced by a return flow of cool, dense, deepwater. The process of surface water cooling and overturning sets up the global thermohaline circulation (Broecker, 1990) with the Nordic Seas (Greenland-Iceland-Norwegian) being often cited as one of the main sites of Atlantic deepwater formation (eg. Veum *et al.*, 1992; Dickson *et al.*, 1996; Bacon, 1998). As with surface water masses, deepwater fluxes are thought to have varied significantly over the last 10,000 years. Bianchi and McCave (1999) have found that deepwater flow over the Iceland-Scotland ridge was greatest 1000, 2000 and 4500-5000 calendar years BP. These periods of increased flow intensity correspond with warmer climatic spells (the so-called 'Medieval Warmth', 'Roman Warmth' and Boreal period). Their work also suggests that thermohaline circulation may be quasi-periodic, with slower phases of deepwater flow recurring every c. 1500 years. More recent changes in deepwater outflow between Iceland and the Faeroe Islands have occurred during the 20th century (Bacon, 1998). Deepwater current data suggests that the rate of flow during the 1980s was approximately twice that of the 1960s. Bacon (1998) believes this variability to be driven by high-latitude atmospheric processes that translate to the ocean via the North Atlantic Oscillation.

Few oceanic records are sufficiently detailed to be able to resolve the hydrographic fluctuations that have occurred throughout the Holocene. To better understand the centennial and decadal-scale climate variability that has been a feature of the last 10,000 years, more high-resolution palaeo-environmental studies are required. Work currently in progress (Rosell-Mélé, *pers. comm., 2000*) is focusing on the past oceanic conditions around Iceland, with emphasis on unravelling the Holocene sequence of events.

Changes in ocean conditions

During recent history the surface ocean circulation in the vicinity of Iceland has been known to change (Lamb, 1977, 1979; Björnsson, 1979; Einarsson, 1991). Although

detailed hydrographic observations have only been carried out since the 1940s, continuous sea-surface temperature measurements for the Northeast Atlantic exist for the last 130 years. Records dating back to AD 1867 from Tørshavn, in the Faeroe Islands, show that significant fluctuations in sea-surface temperature have occurred in recent times. A difference of over 2.0°C is observed between the water temperatures in 1867-1869, 1888-1891 (6.5°C) and those of the warmest period between 1936 and 1951 (8.6°C). These records imply that cold Polar waters have occasionally penetrated as far south as 62°N during the last 130 years. This fact is further supported by observations of sea ice to the east and west of the Faeroe Islands during the severe winter and spring of 1888 (Lamb, 1979; Humlum, 1998a). More recently, in 1966-1970, sea ice was observed in the East Iceland Current as far south as the 64° latitude. This caused sea-surface temperatures at Tørshavn to drop to levels not seen since the late 19th century. Using airborne measurements, Pickett and Athey (1968) recorded a thermal contrast of 5°C between waters to the north and south of the 64° latitude in April 1968. This southward excursion of the oceanographic Polar Front lasted only briefly and was associated with an imbalance of fresher waters in the Arctic Ocean, often referred to as the 'Great Salinity Anomaly' (Mysak and Power, 1996).

Prior to this time, the conditions of the North Atlantic between Iceland and Britain can only be inferred from proxy data. Various historical archives yield important information relating to sea-surface temperatures before the 19th century. Fishing records, charting the relative abundance of temperature-sensitive sea fish, indicate that the periods AD 1675-1700 and 1750-1790 were particularly bad for cod in the vicinity of the Faeroe Islands. It was not until after 1828 that fish numbers improved, only to be curtailed again during the late 19th century (1860-1895) (Lamb, 1979). In the most extreme years, during the 1690s and the 1780s, the cod stock became sparse as far south as Shetland and across the entire Norwegian Sea (Lamb, 1979). Whaling too suffered, presumably as a result of lowered sea temperatures in the Northeast Atlantic. Humlum (1998a) associated periods of relatively few whale catches with incursions of Polar water around the Faeroe Islands. Historical records suggest that whaling around Spitsbergen (75-78°N) was more intense in the early 17th century and decreased notably around 1660 when sea-ice conditions became unfavourable for sailing (Hacquebord, 1984).

A few early Northeast Atlantic sea-temperature measurements exist from the 18th century. The most detailed of which were made by the crew of the Stanley Expedition on their way to Faeroe and Iceland from Edinburgh. The data compiled from the diaries of James Wright allowed Lamb (1979) to reconstruct approximate sea-surface temperatures, expressed as departures from the modern average, for the summer of 1789 (Fig 2.13). The measurements indicate that the Polar waters of the East Iceland current may have continued around the south coast of Iceland, cooling the sea between Iceland and Faeroe by approximately 1.0°C. Meanwhile, warm North Atlantic waters raised temperatures slightly on the west coast resulting in a steep sea-surface temperature gradient.

Changes in sea-ice conditions

Sea-ice conditions around Iceland have varied considerably since the island was settled 1100 years ago (Fig. 2.14 and Fig. 2.15). Sea-ice incidence at the coast of Iceland is closely related to air temperature in coastal regions (Schell, 1961; Bergthorsson, 1969; Ogilvie, 1981; 1984). Thanks to the Icelanders literary tradition, written accounts referring to sea ice are remarkably numerous, detailed and continuous, covering much of the last 1000 years (Ogilvie, 1984, 1990).

Sea-ice observations have been made routinely only since 1920 after the foundation of the Icelandic Meteorological Office (*Veðurstofa Islands*). There have been several attempts to reconstruct the sea-ice record back to the ninth century. Thoroddsen (1916-17) drew together a vast amount of information from various annals, sagas and manuscripts. Koch (1945) produced a chart of sea-ice variations, based largely on Thoroddsen's work (Fig. 2.15). Unlike Thoroddsen, Koch discounted those records that did not specifically mention sea ice. Koch's sea-ice index was expressed in 10-year intervals and showed a considerable increase in ice over the last several centuries, culminating in the period from 1750 to 1900. Koch (1945) concluded that great climatic changes had occurred over the last millennium leading to the relative abundance of ice in the waters off Iceland. Thorarinsson (1956) supplemented Koch's index with accounts that Koch had not discovered. This revised annual sea-ice index from 1601 to 1975 is available in Lamb's *Climate and History* (1977). Sigtryggson

(1972) also published a revised, slightly different, yearly index for the same period (1601-1972) (Fig. 2.15).

Ogilvie (1981, 1984) critically evaluated the sea-ice index derived by Koch (1945) using a qualitative 'thermal index'. She established the relative faithfulness of the sources used in the original compilation and found that, although the bulk of the material was reliable, errors, omissions and factual distortions had occurred in certain years. Consequently, she advised caution when analysing the sea-ice records prior to the 15th century. With reference to unpublished annual reports (Sheriffs' letters) and various, hitherto untapped, Icelandic archives, Ogilvie (1992) presented a new interpretation of the sea-ice record from AD 1650-1850 (Fig. 2.15), which led to a refinement of the work of Thoroddsen (1916-17), Koch (1945) and Bergthorsson (1969). Although more thorough, this new record does not differ greatly from the original index presented by Koch (1945) (Fig. 2.15).

Overall, the Icelandic sea-ice record is a very valuable climate proxy spanning approximately the last 1100 years, and is only interrupted between c. AD 1430 and c. 1550 (Ogilvie, 1981). Bergthorsson (1969) has shown how the average course of air temperature in Iceland since c. AD 900 can be estimated based on the fairly strong correlation ($r = -0.68$) between sea-ice severity and temperature measurements during recent times (AD 1846-1919). More recently, however, Ogilvie (1981, 1984) has questioned the exact validity of reconstructing climatic conditions from documentary, often anecdotal, evidence. Clearly, the use of sea-ice conditions to reconstruct air temperatures is not entirely satisfactory as the degree of sea ice reflects a variety of atmospheric and oceanographic factors – including wind speed and direction, sea-surface salinity and temperature, and ocean circulation patterns. Rather, sea-ice indices should be used more appropriately to approximate the relative degree of Arctic air and ocean currents around Iceland. However, in combination with other proxies, human observations of sea ice from Iceland show clearly how variations in climate have affected the Northeast Atlantic region over the last ten centuries or so.

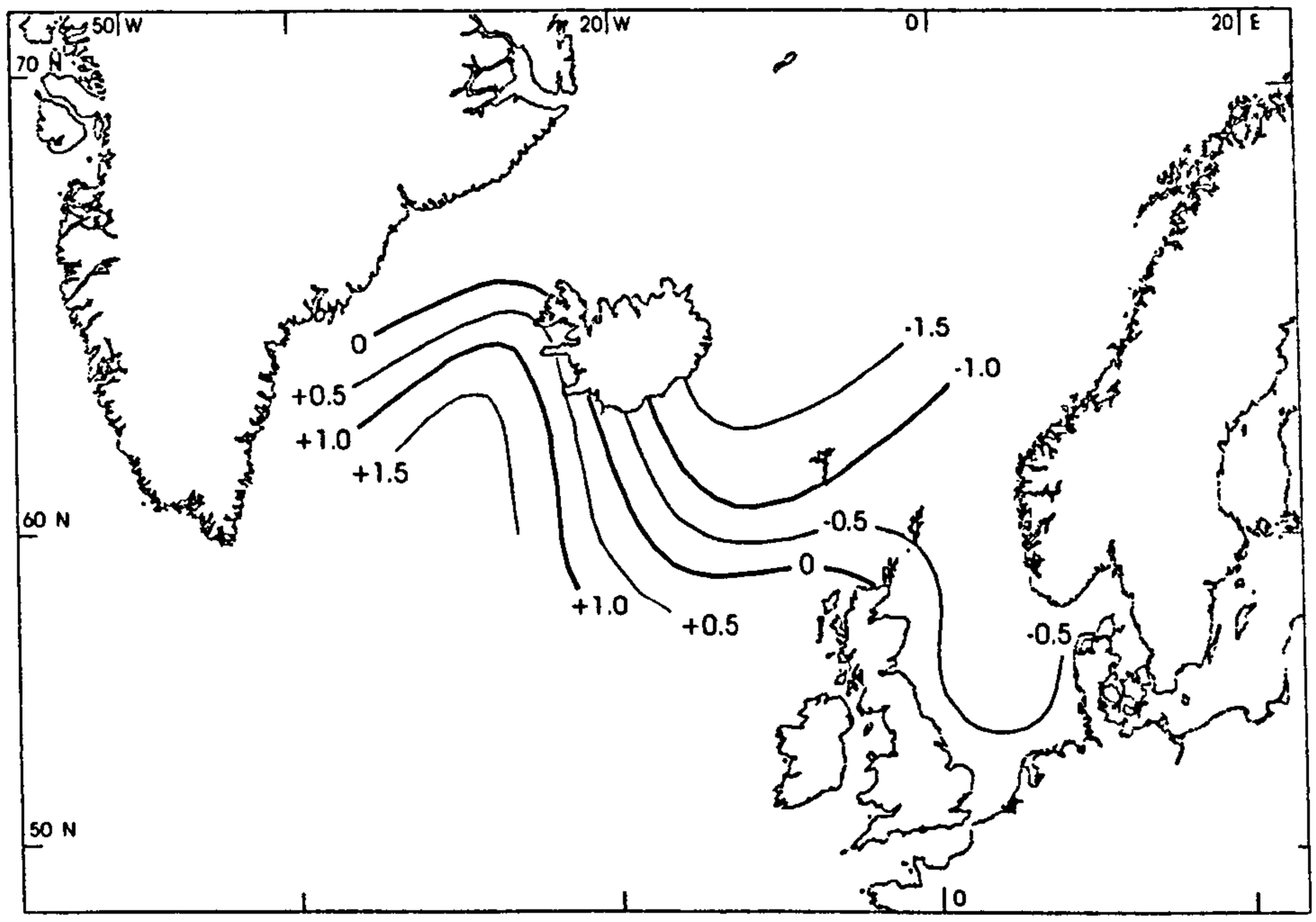


Figure 2.13 Sea-surface temperatures measured in the summer of 1789 (Stanley Expedition). Isotherms are departures (°C) from modern average values (1931-1960) (taken from Lamb, 1979)

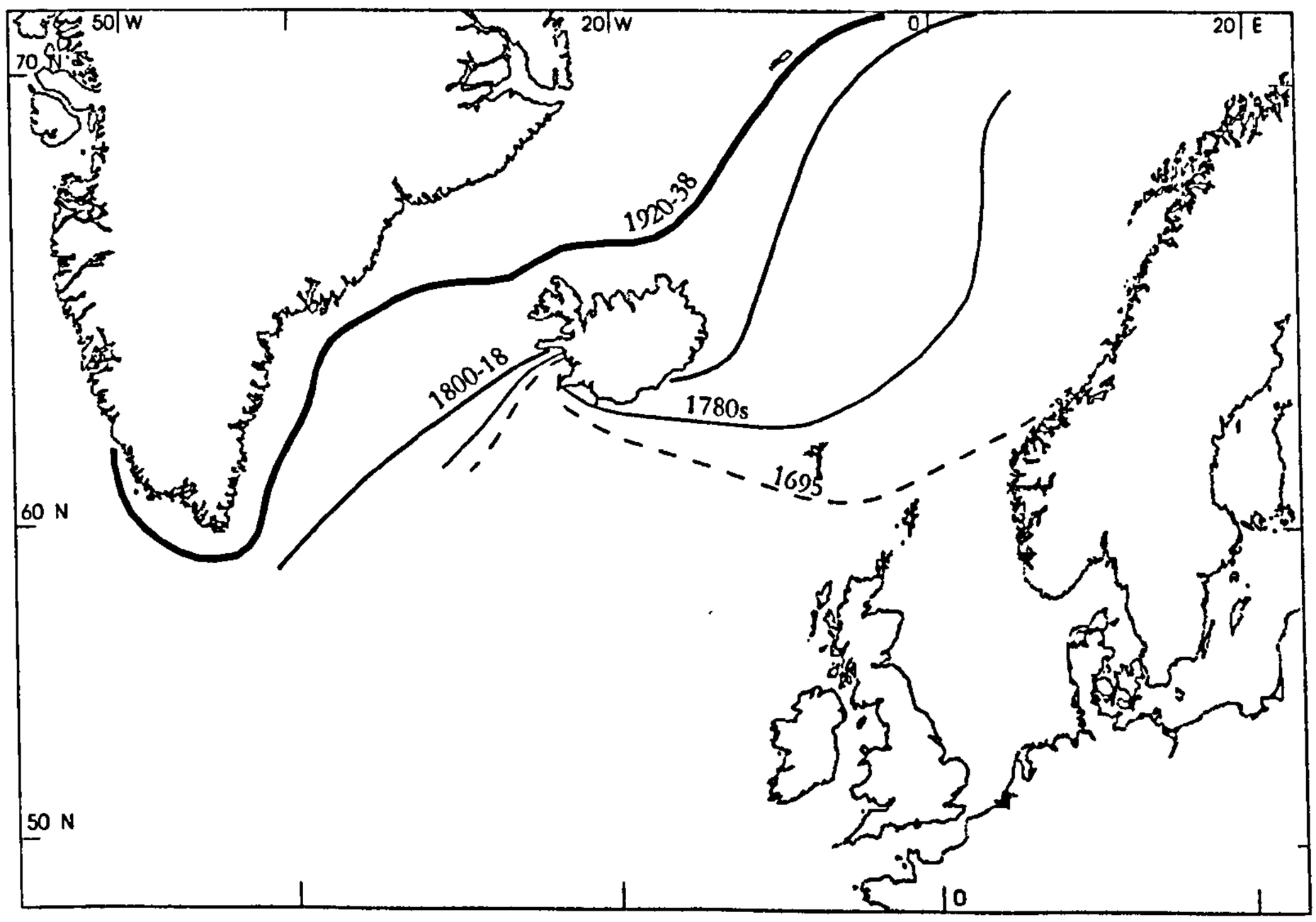


Figure 2.14 Limits of the Arctic sea ice at approximately maximum extent (March-May) during various periods (based on Lamb, 1979)

——— 1920-1938 - 20th century minimum (Veðurstofa Islands)
——— 1800-1818 - after William Scoresby
——— 1780s - approximate maximum extent during extreme years between 1770 and 1820 (most notably from 1782-1795)
– – – 1695 - approximate maximum extent in 1695, a particularly severe year (Sea-ice may have encroached to this limit)

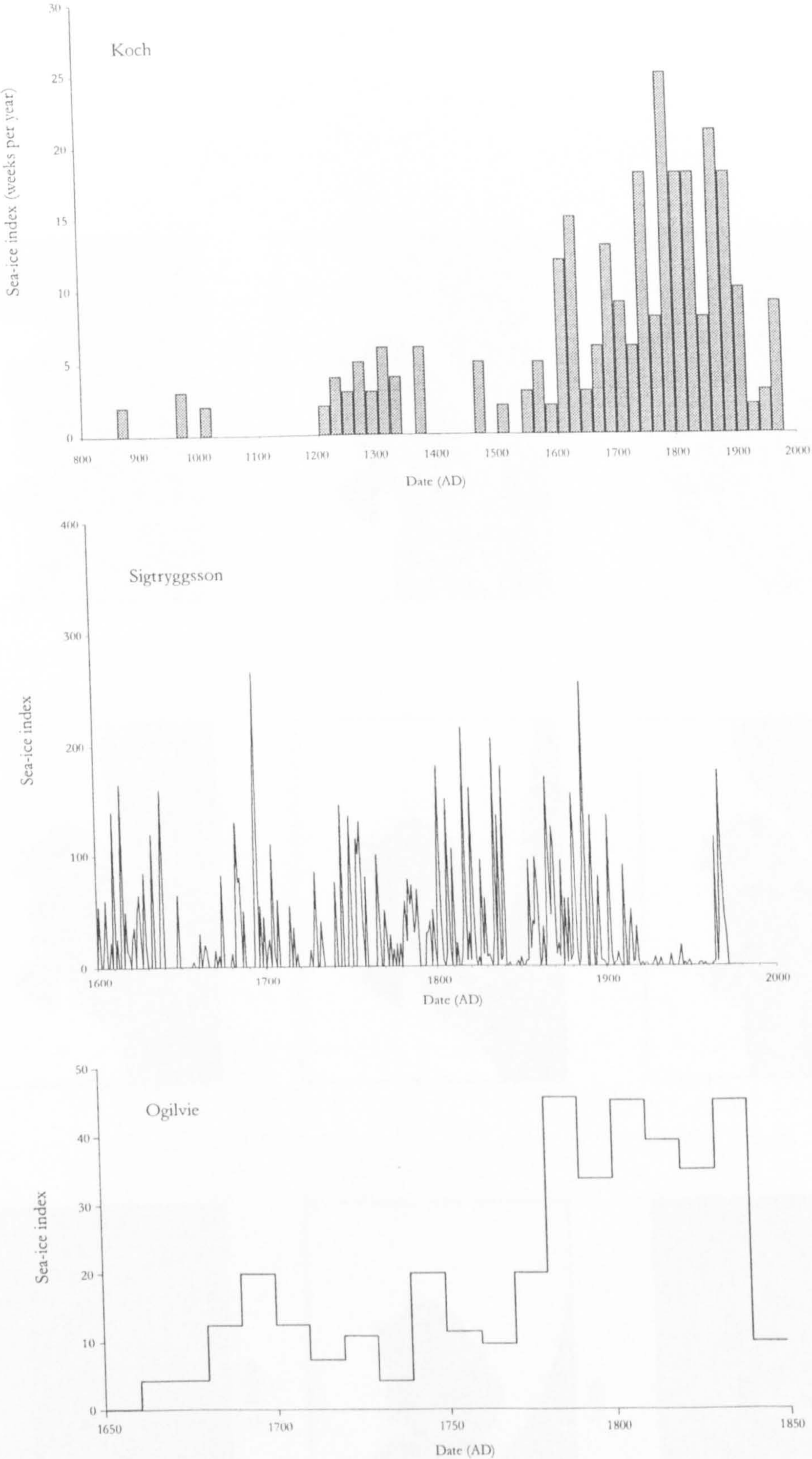
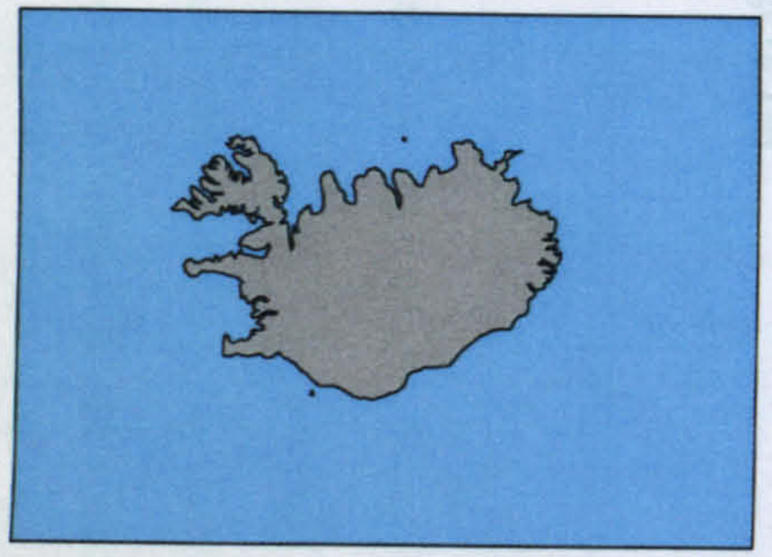
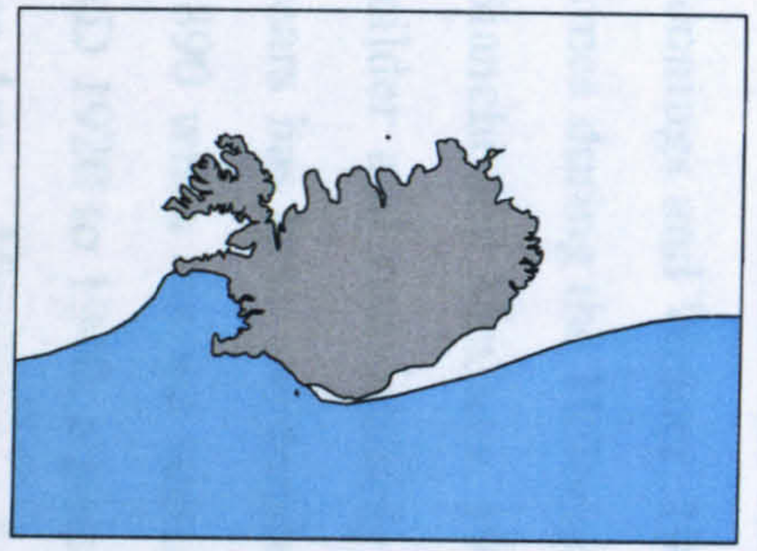


Figure 2.15 Sea-ice conditions around Iceland according to Koch (1945), Sigtryggsson (1972) and Ogilvie (1992). Note both the horizontal and vertical scales differ in all three graphs.

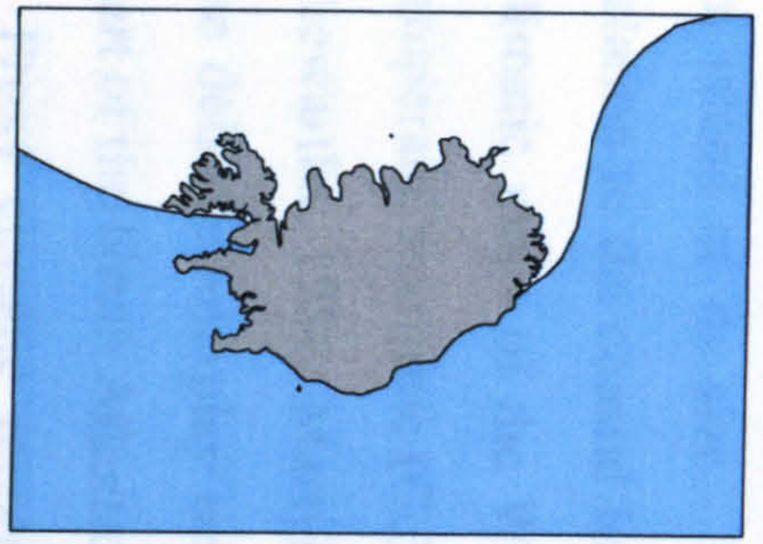
– Koch's (1945) [updated] index represents twenty-year averages for the whole 'settlement' period – AD 870 to 1980.
– Sigtryggsson's (1972) index is annual values for the period AD 1601-1971. Figures since 1919 are taken from the Icelandic sea-ice reports published in *Veðrattan*.
– Ogilvie's (1992) index is decadal values covering the period AD 1650-1850 and is believed to be the most accurate representation of sea-ice conditions during the 17th and 18th centuries.



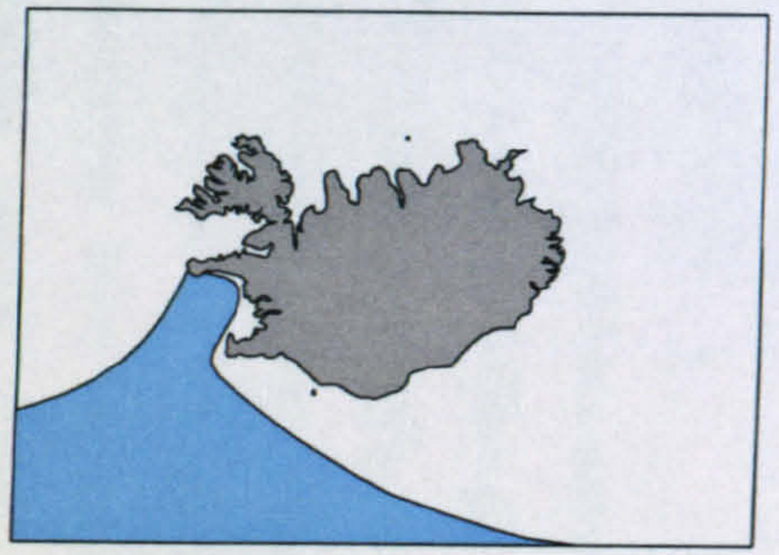
c. 1000



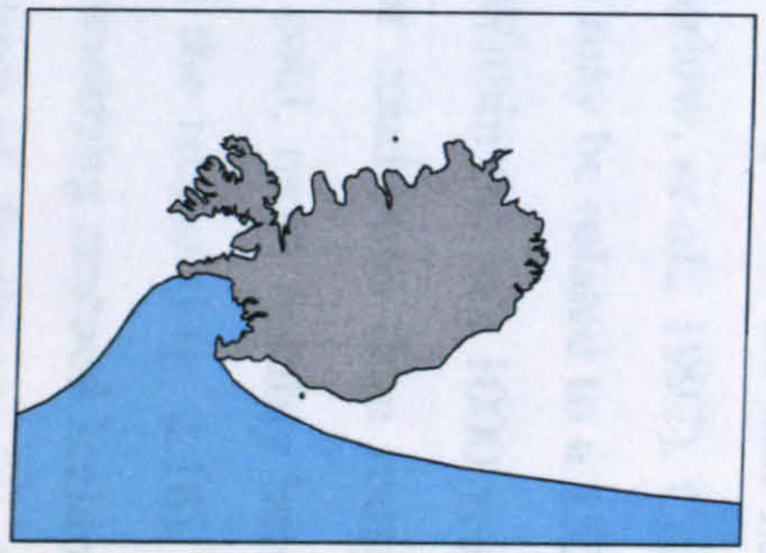
c. 1280



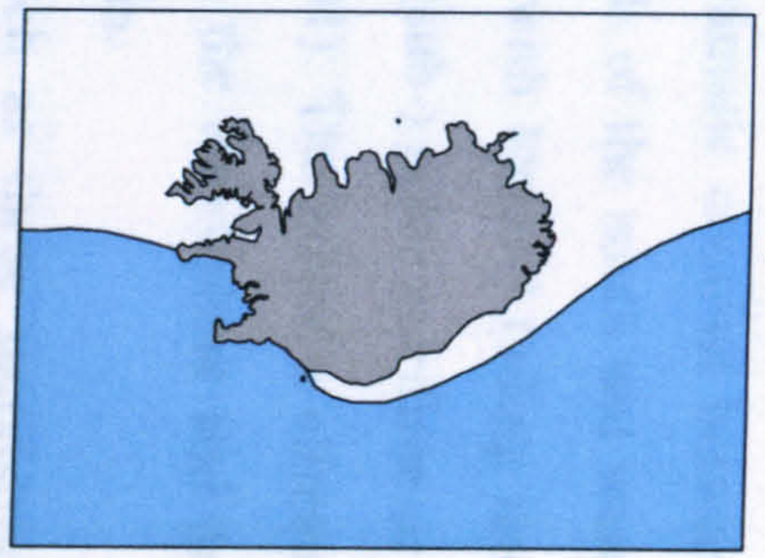
c. 1600



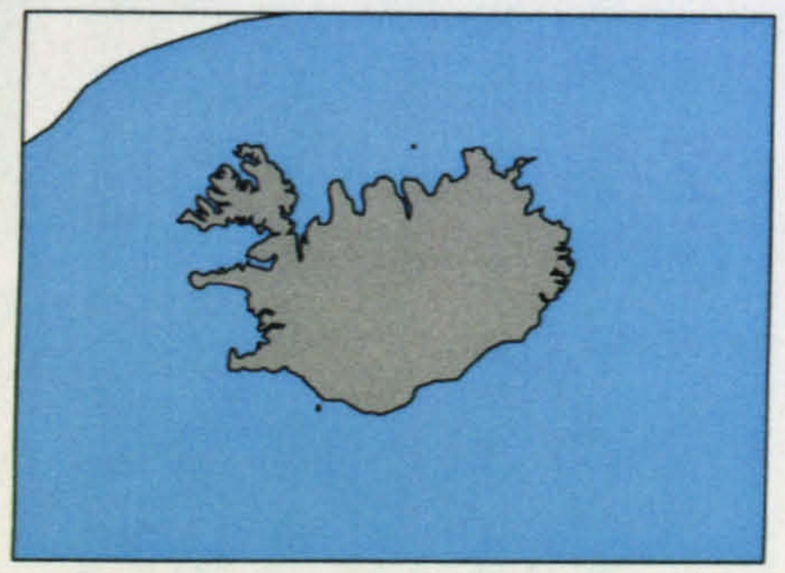
1695



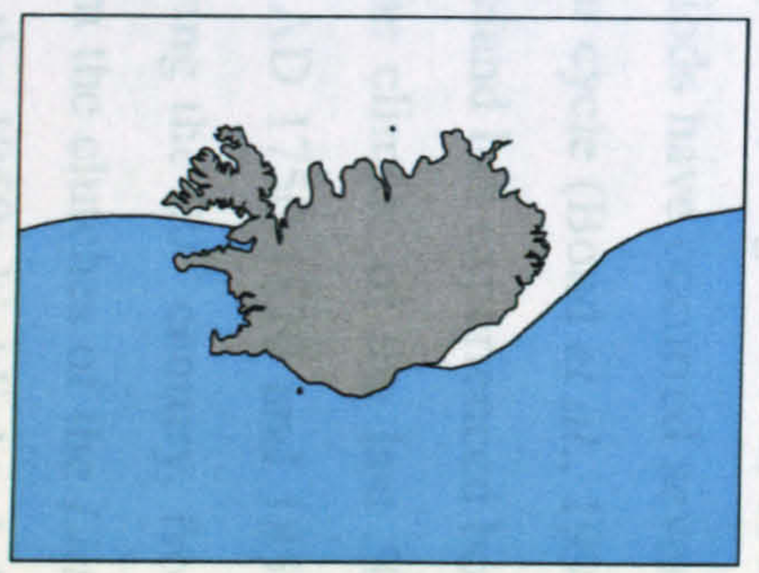
1785



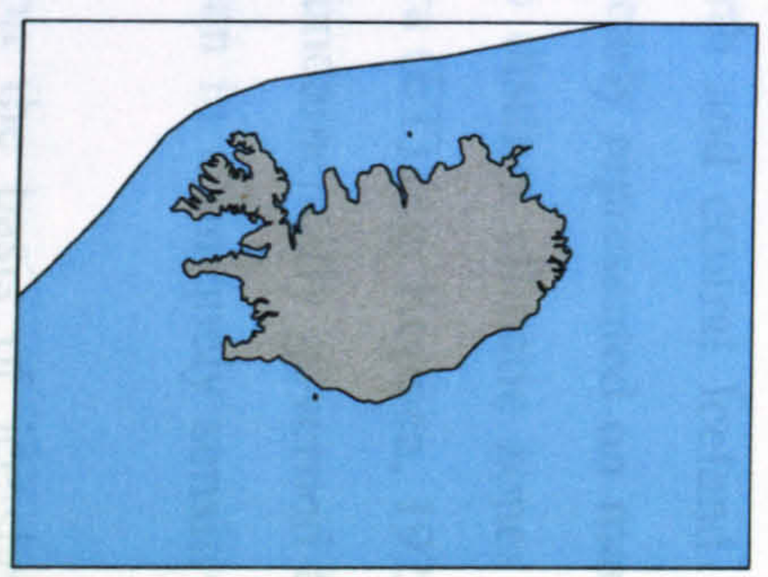
1888



1938



1968



1998

Figure 2.16 Maximum extent of sea ice around Iceland at various times in the last 1000 years. The maps have been compiled on the basis of work by Koch (1945), Sigtryggsson (1972), Lamb (1979) and Ogilvie (1981, 1984, 1992). Note the severe conditions in AD 1695 and 1785 compared to those of around AD 1000 – the time of Norse settlement.

Summary

To summarise, Iceland's climate is presently a cool, maritime one. Considerably warmer than other landmasses at the same latitude (eg. Baltic Sweden and southern Baffin Island), Iceland experiences a remarkably wide range of weather conditions for an island of its size. The climatic contrast between north and central Iceland in relation to the coastal lowlands of the south and west is clearly represented on many climatic maps of the World, with Iceland falling into two distinct climatic zones – temperate, maritime (Cc) and (sub-) polar maritime or tundra (ET) (eg. Koppen, 1923; Trewartha, 1954; Martyn, 1991). The island's climate is fundamentally governed by the delicate interplay between the atmosphere and the ocean in a uniquely sensitive part of the North Atlantic Ocean.

Proxy climatic records, such as those reconstructed on the basis of ice-sheet accumulation rates, ocean floor sediments and sea-ice observations, suggest that Iceland's climate has changed notably over the last 10,000 years (Bergthorsson, 1969; Jennings and Weiner, 1996; Barlow, *et al.*, 1997). Cold periods have occurred several times during the Holocene and may be related to a 1500-year cycle (Bond *et al.*, 1997; Bianchi and McCave, 1999). Within the last 1000 years, Iceland has experienced both milder and considerably cooler conditions than today. The climate of the last 500 years has been predominantly cold, most notably between AD 1750-1820 and 1850-1890 when sea-ice surrounded the island (Fig. 2.16). During the 20th century, from AD 1920 to 1940, a period of warming rescued Iceland from the clutches of the *Little Ice Age*. However, a brief return of Arctic sea-ice during the 1960s highlighted the remarkable variability of the climate in Iceland and of the Northeast Atlantic in general.

Chapter 3 : Present Knowledge of Glacier Fluctuations in Iceland

Introduction

Numerous scientists have concerned themselves with the former extent of Iceland's glaciers. In recent times research has concentrated on the relationship between glacier fluctuations and climatic change.

Arguably, Icelanders have been studying the nature of their landscape since the first settlers arrived from Norway in the late 9th century. The first scientific description of a glacier was probably written around AD 1200 by Saxo Grammaticus (Thorarinsson, 1960). This earliest of accounts mentions rotational slipping and terminal moraines some 600 years before Louis Agassiz proposed the theory of glaciation. The history of glacial knowledge in Iceland is a large subject beyond the scope of this research. The reader is referred instead to detailed reviews by Thorodssen (1905-06), Bardarsson (1934) and Thorarinsson (1960). This chapter aims to summarise the present knowledge regarding glacier fluctuations in Iceland with reference to four specific time periods: – the last 70 years, the last 1000 years, the period from 1000 to 9000 years BP and the period from 9000 to ~13,000 years BP.

The 'observational' period (AD 1930 –)

In 1930, Jon Eythorsson, a meteorologist and keen natural scientist, began monitoring the frontal positions of 44 glaciers in Iceland. Although measurements on some of these glaciers were later abandoned, 28 of those originally measured are still attended annually and a further 13 glaciers have been added in recent decades. Measurements have been performed mostly by farmers living close to the glaciers in question. Each year the surveyor records the distance in metres from the ice-front to a benchmark of known position. However, many glaciers remain unobserved, particularly those in less

accessible areas. No ice-front measurements exist for northwest and easternmost Vatnajökull, northern Mýrdalsjökull, eastern Hofsjökull and western Langjökull.

Summaries of glacier fluctuations (*Jöklarbreytingar*) have been published in *Jökull*, the Icelandic Earth Science journal, since 1932. Longer-term trends have been analysed 10, 30 and 65 years after the first measurements were made (Eythorsson, 1942, 1963; Sigurdsson, 1998). The complete dataset has been computerised recently and critically examined to investigate the relationship between glacier fluctuations and climatic change over the last seven decades (Johannesson and Sigurdsson, 1998).

The frontal movements of 12 glaciers chosen from the Icelandic Glaciological Society database are shown in Figure 3.1. The choice of glaciers highlights the spatial variability of glacier fluctuations in Iceland as well as the full temporal range of observations.

The 'historical' period (AD 870 – 1930)

It has long been known that the degree of glacial recession in Iceland and, more generally, around the North Atlantic, particularly during the first two-thirds of the 20th century, has been quite considerable (Thorarinsson, 1943, 1956; Ahlmann, 1948; Johannesson and Sigurdsson, 1998). The large amount of ice-front recession over the last 100 years is chiefly due to the advanced stage occupied by most glaciers during the 18th and 19th centuries. Eythorsson stated in 1935 that the outlet glaciers of Drangajökull in Northwest Iceland were receding from an advanced stage, probably their greatest extent in historical time. Thorarinsson (1936, 1943, 1956) echoed Eythorsson's feelings stating that many of the recent terminal moraines in Iceland (ie. 17th, 18th and 19th century) mark the maximum extension of glaciers in "historical and probably even Postglacial Times" (Thorarinsson, 1956: 1). Although Thorarinsson (1956) later found outlying moraines dating from the pre-settlement period (which probably formed c. 1500-2500 years BP), he stressed that these were few in number and that the position of these moraines implied that any prehistoric advance only slightly exceeded that of recent centuries.

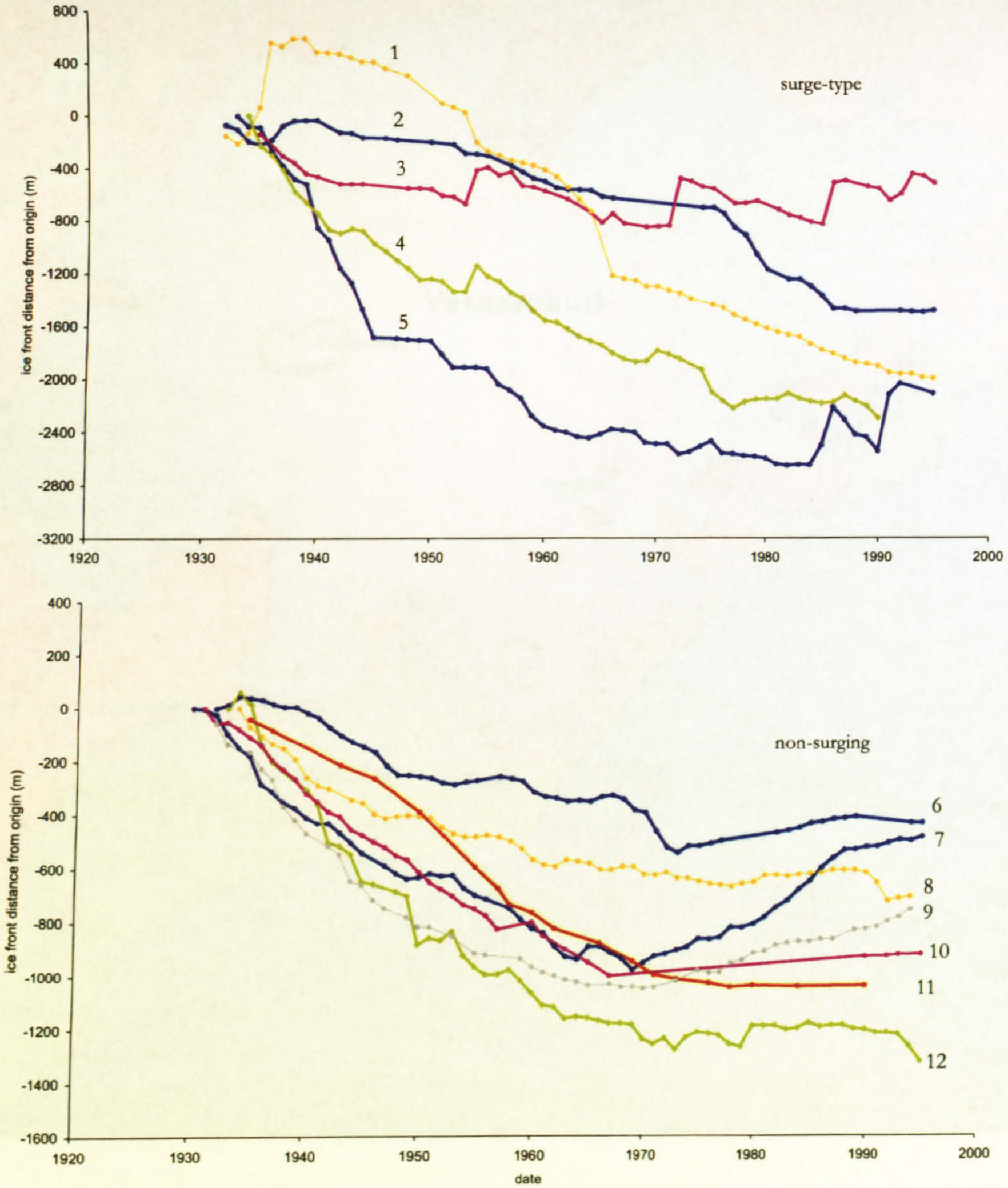
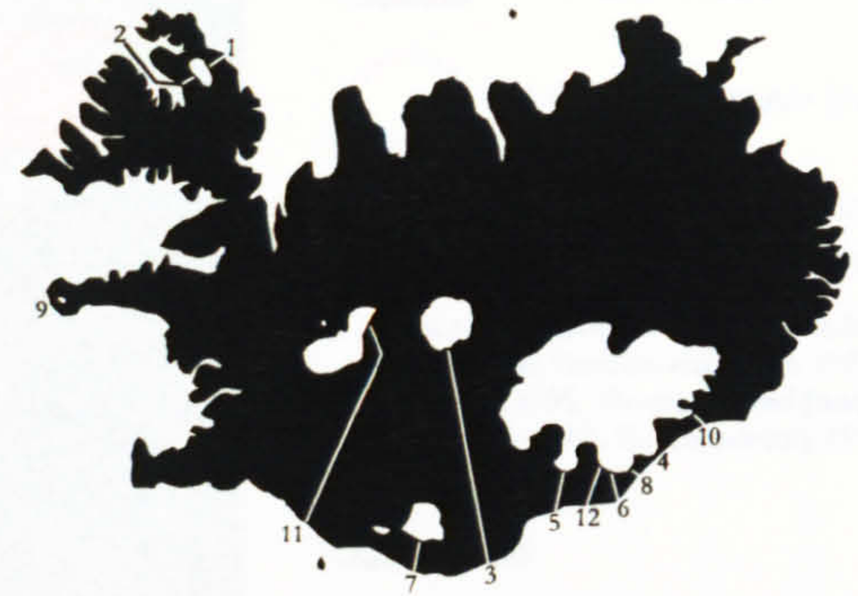


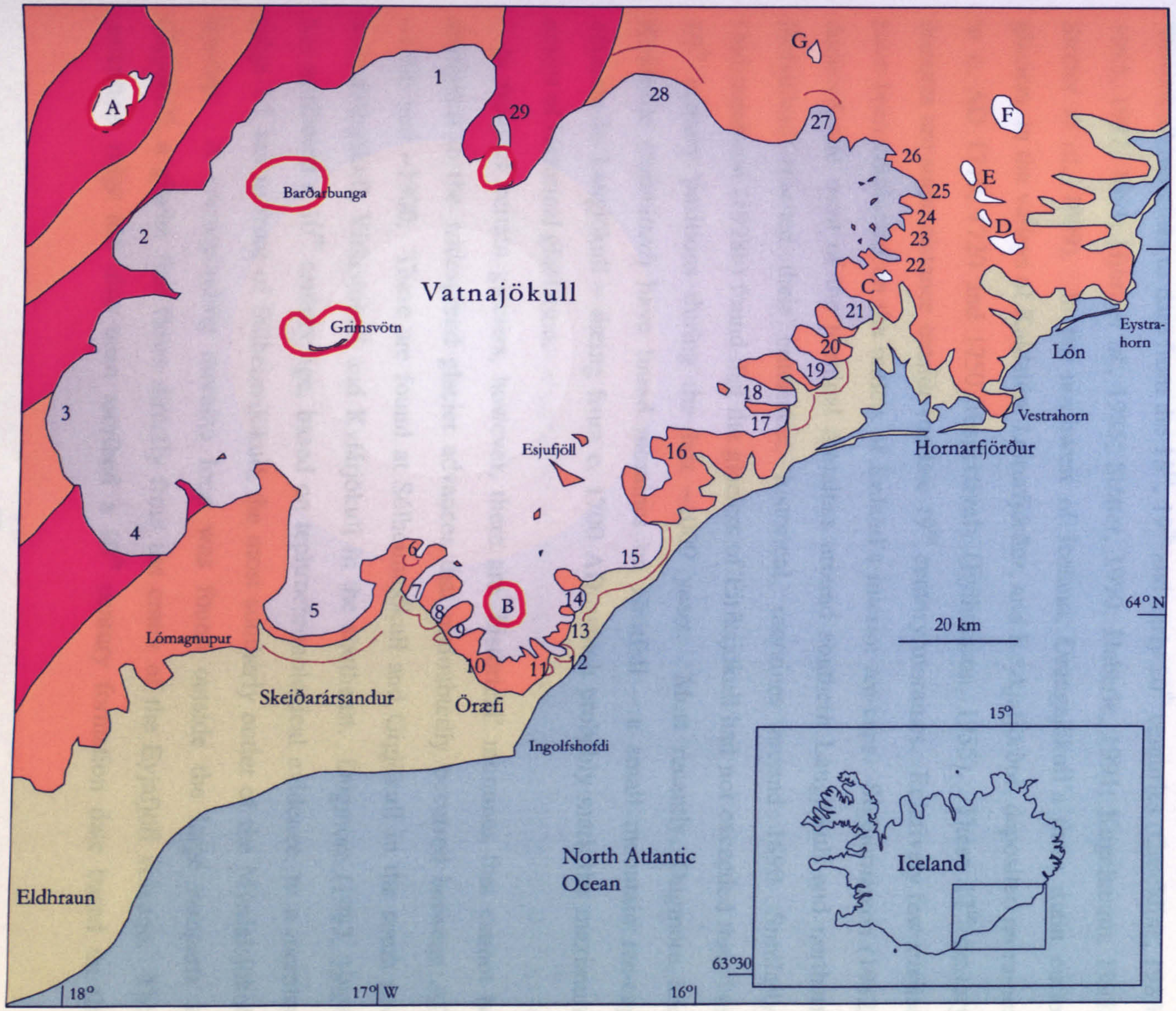
Figure 3.1 Fluctuations of glaciers in Iceland 1930-1995 (taken from the database of the Iceland Glaciological Society; Sigurðsson, 1998).
12 glaciers have been chosen for comparison – 5 surging and 7 non-surging outlets. Note the vertical scale on the upper graph is twice that of the lower graph.



Key to glaciers

No.	*Glacier*	*Ice-cap*	*Length*
1	Reykjafjarðurjökull*	Drangajökull	7
2	Kaldalónsjökull*	Drangajökull	6
3	Múlajökull*	Hofsjökull	20
4	Breiðamerkurjökull*	Vatnajökull	40
5	Skeiðarárjökull*	Vatnajökull	50
6	Virkisjökull	Öræfajökull	6
7	Sólheimajökull	Mýrdalsjökull	15
8	Fjallsjökull	Öræfajökull	15
9	Hyrningsjökull	Snæfellsjökull	2
10	Skálafellsjökull	Vatnajökull	25
11	Jökulkrókur	Langjökull	11
12	Skaftafellsjökull	Vatnajökull	18

Notes:
Length is the distance in km along the flow-line in 1990.
* denotes a surge-type glacier.



Key

- Land below 100 m asl
- Land above 100 m asl
- Glacier ice
- Active volcanic rift zone
- Active volcano (caldera)
- Terminal moraines dating from the late 19th century

(According to Ives and King, 1955; Ives, 1956; Jaksch, 1970; Gordon and Sharp, 1983; Sharp and Dugmore, 1985; Thompson and Jones, 1986; Thompson, 1988; Gudmundsson, 1998a; Evans et al., 1999)

Ice masses

A Tungnafellsjökull
B Öræfajökull
C Grasgiljatindur
D Jökultindur
E Hofsjökull í Lóni
F Thrandarjökull
G Snæfell

Glaciers

1 Dyngjujökull
2 Koldukvíslarjökull
3 Tungnaárjökull
4 Síðujökull
5 Skeiðarárjökull
6 Morsárjökull
7 Skaftafellsjökull
8 Svínafellsjökull
9 Virkisjökull
10 Kotájokull
11 Hólárjökull
12 Kvíárjökull
13 Hrútarjökull
14 Fjallsájökull
15 Breiðamerkurjökull
16 Brókarjökull
17 Skálafellsjökull
18 Heinabergsjökull
19 Flaájökull
20 Viðbordsjökull
21 Hoffellsjökull
22 Lambatungnajökull
23 Austurtungnajökull
24 Nordurtungnajökull
25 Öxarfellsjökull
26 Vesturdalsjökull
27 Eyjabakkajökull
28 Brúárjökull
29 Kverkfjöll

Figure 3.2 Map of southeast Iceland showing the present-day glaciers and volcanoes. *Little Ice Age* moraines are also shown, most of which are believed to date from AD 1850-1890.

Geomorphological evidence for glacier fluctuations in Iceland between the time of settlement around AD 870 and the establishment of the glaciological monitoring project in 1930 is abundant. However, most studies of Vatnajökull's glaciers have concluded that the outermost 'historical' moraines date from the 19th century – notably the period 1850-1895 (Fig. 3.2) (Ives and King, 1954, 1955; Ives, 1956; Jaksch, 1970, 1975; Gordon and Sharp, 1983; Sharp, 1984; Sharp and Dugmore, 1985; Thompson and Jones, 1986; Thompson, 1988; Gudmundsson, 1998a; Evans *et al.*, 1999). In the north of Iceland, studies of cirque glaciers in the highlands of Tröllaskagi have found the outermost 'historical' moraines to date from the 18th, 19th and early 20th centuries (Caseldine, 1983, 1985, 1987, 1991; Hjört *et al.*, 1985; Stötter, 1990; Haberle, 1991; Kugelmann, 1991; Stötter *et al.*, 1999). In the northwest of Iceland, Drangajökull's three main outlet glaciers, in the valleys of Kaldalón, Leirurfjörður, and Reykjafjörður, deposited moraines in c. AD 1710, 1720 and 1750, respectively (Eythorsson, 1935). These 17th century features are some distance outside the late 19th century moraines. Relatively few studies have been conducted on the outlets of Iceland's interior ice-caps. Sigbjarnarson (1982) deduced that most of the terminal moraines around southern Langjökull and northern Hofsjökull reached their outermost, historical, moraines around 1890. Similarly, Gudmundsson (1998b) found that the glaciers of Eiríksjökull had not exceeded their late 19th century positions during the last ~1000 years. Most recently, Dugmore and Kirkbride (*submitted*) have found moraines on Hrútfell – a small mountain ice-cap adjacent to Langjökull – dating from c. 1700 AD, which probably mark the maximum extent of historical glaciation.

In front of certain glaciers, however, there are 'historical' moraines that cannot be attributed to the widespread glacier advances that undoubtedly occurred between AD ~1700 and ~1900. These are found at Sólheimajökull and Gigjökull in the south and Svínafellsjökull, Virkisjökull and Kvíárjökull in the southeast. Dugmore (1987, 1989) has attributed a 10th century age, based on tephrochronological evidence, to a moraine ridge 1.5 km in front of Sólheimajökull, the most southerly outlet of the Mýrdalsjökull ice-cap. A corresponding moraine loop was found outside the large ramparts at Gigjökull, a glacier that flows directly from the crater of the Eyjafjöll volcano. This outermost ridge has also been ascribed a 10th century formation date based on the

distribution of early historical-age tephra layers in the soil within and outside the moraines (Dugmore, 1987, 1989).

Tephrochronological studies conducted by Gudmundsson (1998a) have led to a better understanding of historical glacier fluctuations in the Öræfi district. Three steep outlet glaciers of Öræfajökull – Svínafellsjökull, Virkisjökull and Kvíárjökull – possess moraine fragments dated to around the 13th century, suggesting that all three were more extensive than in later centuries. The *Storalda* moraine in front of Svínafellsjökull is actually a complex feature comprising at least four distinct ridge crests (Gudmundsson, 1998a). Although the exact age of the feature is still unclear, the moraine probably marks the extent of a pre-historic advance (c. 2000 BP) as well as advances in the 13th (AD 1200-1362) and early 18th (AD 1727-1755) centuries (Gudmundsson, 1998a: 184).

In summary, there is little evidence to suggest that Iceland's glaciers exceeded their 18th and 19th century positions during the first eight centuries of settlement. Apart from those few moraines dated to the 10th century, by Dugmore (1987) and Stötter (1990), and three 13th-century moraines identified by Gudmundsson (1998a), the vast majority of 'recent' terminal moraines are less than 200 years old, with most apparently dating from the late 19th century. This may be a function of various factors, including the lack of preservation of the geomorphic record (due to high volumes of meltwater), limitations surrounding the dating technique (ie. lichenometry) and the adoption of a rigid paradigm (ie. Thorarinsson, 1943). However, it is widely assumed that glaciers in Iceland were less extensive during the first several centuries of settlement than during the *Little Ice Age* of the 18th and 19th centuries, with the Vatnajökull ice-cap reaching its maximum dimensions in the late 19th century.

The Holocene period (~1000 – 9000 BP)

Published evidence of pre-historic glacier advances in Iceland is surprisingly rare. Eythorsson (1935) has argued that some of the moraines in Northwest Iceland, particularly those in the valleys of Kaldalón and Leirufjörður, are from advances that must have taken place between the last major glaciation and the historical period. The *Jökulgarðar* moraine at Kaldalón, also mentioned by John and Sugden (1962), remains undated although Gudmundsson (1997: 86) has suggested that it may be of Sub-Atlantic age (c.2000 BP). Features of possibly similar age have been identified, but not studied, on Hornstrandir, in the extreme northwest of Iceland (Hjört *et al*, 1985; John, 1975).

Sigurður Thorarinsson (1936: 194; 1946: 260; 1949: 250) was the first to suggest that certain moraines in southeast Iceland probably formed in prehistoric times around 1000 years before human settlement. However, Todtmann (1952) was the first to identify a pre-historic glacier advance in Iceland. She reported finding the distinctive white tephra of the 1362 Öræfajökull eruption in the soil on the conspicuous *Storalda* moraine adjacent to Svínafellsjökull. This feature was later examined by Thorarinsson (1956) in more detail. He found *Storalda* to be a complex landform, resulting from more than one glacial advance, with the oldest part probably dating to the Sub-Atlantic period (Thorarinsson, 1956: 7). Todtmann (1952: 407) discussed the possibility that moraines in front of Kvíárjökull, an outlet of Öræfajökull, and a recently uncovered feature on the glacier foreland at Breiðamerkurjökull, were also of pre-historic age. Thorarinsson (1956) used tephrochronology to show that the great termino-lateral moraines at Kvíárjökull were the product, at least in part, of a pre-historic glaciation, probably also of Sub-Atlantic age.

Two further chronological studies have since been conducted on the moraines at Kvíarjökull. Black (1990) found evidence of two pre-historic advances that he called the *Pre-Settlement* (AD 720 +/-395) and the *Sub-Atlantic* (>2040 +/-80 ^{14}C yrs BP) stages. Meanwhile, Gudmundsson (1998a), as part of a regional tephrochronological investigation, revealed a more complex glacial history with pre-historic moraines dating from c. 1700 BP, 3300 BP and 4600-6000 BP. Gudmundsson (1998a) identified pre-historic moraines at two other outlets of Öræfajökull, Skaftafellsjökull and Virkisjökull,

and possibly at a third, Kotájökull. These moraines represent advances temporally coincident with those identified at Kvíárjökull.

Sharp and Dugmore (1985) dated buried soils relating to former advances of two large outlets of Vatnajökull, Skálafellsjökull and Eyjabakkajökull. They found that both glaciers advanced during the Holocene, at c. 5700 BP and 6900 BP respectively, although these advances were less extensive than those of the late 19th century.

Dugmore (1987, 1989) deduced the pre-historic sequence of events at Sólheimajökull. Lateral moraines, located outside those deposited in recent centuries, were dated using tephrochronology, supported by radiocarbon analyses, to various periods within the Holocene. Dugmore's (1989) findings were profound. He demonstrated that Sólheimajökull was larger during the mid-Holocene (4500-7000 BP) than at any time since. Furthermore, Dugmore (1989) identified two other stages of glacier expansion or stillstand at around 3100 BP and 1200-1400 BP. This evidence suggests that glaciers may not have completely disappeared from Iceland during the warm mid-Holocene era.

The Tröllaskagi highlands of northern Iceland no longer support any large ice-masses. However, research has shown that cirque glaciers at the head of several valleys were more extensive earlier in the Holocene (Stötter, 1990, 1991, 1994; Häberle, 1991, 1994). As in southern Iceland, pre-historic moraines have been dated using ^{14}C analyses combined with tephrochronology. Glacier advances or stillstands occurred at c. 1550 BP, 2000 BP, 2800-3500 BP, 4300 BP and not long before 6000 BP (Stötter, 1990; Häberle, 1991).

Although knowledge regarding the fluctuations of Iceland's interior ice-caps is limited, recent work at Hrútfell – an isolated mountain ice-cap – has identified glacier advances that culminated between c. 2200-2600 BP (Dugmore and Kirkbride, *submitted*). Furthermore, Eystri Loðmundarjökull, an outlet of Langjökull was also more extensive during the mid-Holocene. Tephrochronological evidence has affixed the age of the outermost moraines to c. 4-6000 BP (Kirkbride and Dugmore, 2001a). Gudmundsson (1998b) assigns a glacier advance at Eiríksjökull to the pre-historic period based on the geomorphological relationship between the outermost moraine and a historical lava-flow.

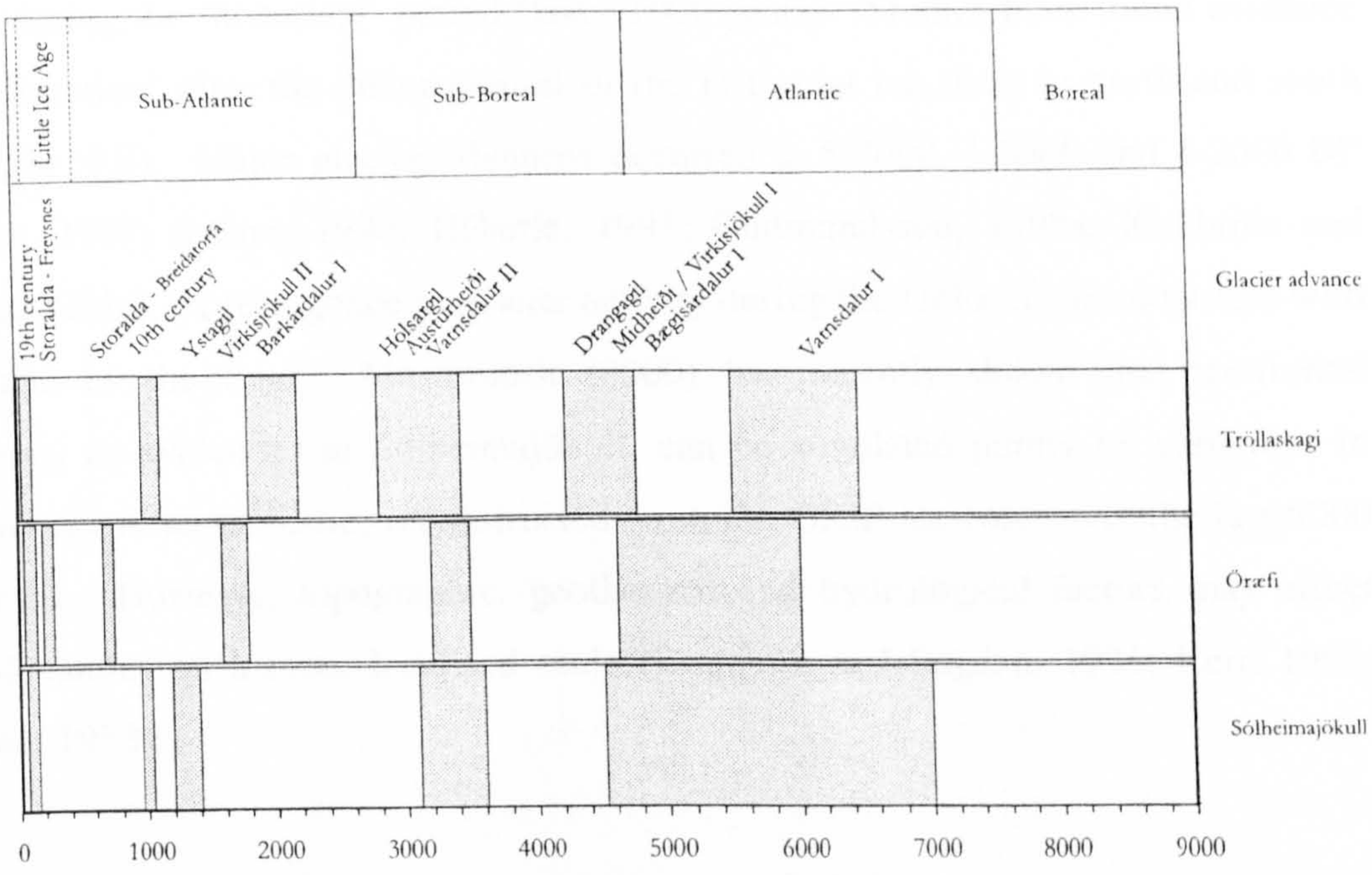


Figure 3.3 A summary of glacier fluctuations in Iceland since ~9 ka BP.

The diagram is based on studies conducted in three different areas:
Dugmore (1987) - Sólheimajökull, south Iceland
Gudmundsson (1998) - Öræfajökull, southeast Iceland
Stötter (1990), Häberle, (1991) - Tröllaskagi highlands, north Iceland

In summary, Iceland's glaciers were considerably larger earlier in the Holocene than at any time during the 'historical' period (last ~1000 years). Studies have found evidence of 'Neoglaciation' after the disintegration of the last great ice sheet in north and south Iceland (Fig. 3.3). Major glacier advances occurred c. 5-7000, 3-3500 and 1-2000 BP (Dugmore, 1987; Stötter, 1990; Häberle, 1991; Gudmundsson, 1998a; Kirkbride and Dugmore, 2001a). A resurgence in glacier activity during the Holocene is consistent with the climate of the time. Mackintosh (2000) has recently shown that ice-frontal fluctuations, as evidenced at Sólheimajökull, can be simulated purely by variations in precipitation and temperature, reconstructed from the GRIP ice-core, over the last 5000 years or so. However, topographic, geothermal and hydrological factors may affect glacier dynamics on a more localised scale (Dugmore and Sugden, 1991; Kerr, 1993; Björnsson, 1988).

The Lateglacial period (~9000 – 13,000 BP)

Geological and geomorphological investigations over the last century have clarified the course of Late Weichselian and Early Holocene deglaciation in Iceland. End moraines are found at great distances outside the presently glaciated areas in southwest, central and northeastern parts of Iceland. These features, first identified by Thorodssen (1905-06) were thought to mark the stationary positions of an ice-sheet that had retreated almost continuously from a maximum at the end of the Weichselian glaciation. Subsequently, Pjeturss (1910) referred to moraines in northeast and southern Iceland that he believed marked a re-advance of the ice-margin. Pjeturss matched these two features with one another, suggesting that they had both formed during a brief return to glacial conditions, which he named the *Langanes Stadial*.

Further evidence of an ice-sheet re-advance during Lateglacial times was presented by Kjartanssan (1943, 1958). He mapped the continuous moraine complex running over 100 km from near Eyjafjallajökull to Efstadalsfjall, referring to it as the Buði moraine (Fig. 3.4). Thorarinsson (1951) identified a similar end-moraine complex near Mývatn in northern Iceland – the Reykjahlíð-Reykjadalur moraine. Einarsson (1964) suggested that

both this moraine and the Buði complex had been formed during the same climatically induced re-advance. This date of the event was uncertain although Thorarinsson (1951) considered it to be of Younger Dryas age.

Einarsson (1967, 1968) revised his initial model of deglaciation when subsequent moraines were found outside these areas. Located in southwest Iceland, the Alftanes, Hvalfjörður and Borgarfjörður end moraines were correlated with the moraines of Eyjafjörður in the north. Einarsson (1968) referred to this oldest advance as the *Alftanes event* but he also correlated it with the *Langanes* event of Pjeturss (1910). Although no dates had been attributed to these Late Weichselian events, Einarsson (1968) suggested that the earlier of the two advances was of Older Dryas (c. 12,000 BP) age whilst the Buði moraine probably represented a Younger Dryas re-advance (c. 10,000 BP).

During the 1980s the model of Weichselian deglaciation became more elaborate. Norðdahl (1979, 1981, 1983), working in the Eyjafjörður region of northern Iceland, proposed a series of ice-sheet fluctuations between the time of maximum glacial extent and the early Holocene (~18 – 9600 BP). A re-investigation of the Borgarfjörður region in southwest Iceland by Ingolfsson, (1985, 1987, 1988) led him to conclude that, at least in certain areas, the Icelandic ice-sheet was almost as extensive in Younger Dryas as Older Dryas times. Petursson (1986, 1991) also found that the extent of the Younger Dryas ice in northeast Iceland was much greater than had been previously believed. In southern Iceland, Johannesson (1985) studied the Buði end-moraine stating that the whole complex, made up of six or seven ridges, could span the entire period between Older Dryas and Younger Dryas (c. 10 - 12,000 BP). More detailed investigations by Hjartarson and Ingolfsson (1988) found marine shells within glacial till which put the maximum age of the Buði moraine between 9745 and 9595 BP (Fig. 3.4).

In eastern Iceland the deglacial picture is complicated. Only two regional studies have been conducted in the eastern fjords. Hjartarson *et al.* (1981) identified two stages of Late-Weichselian re-advance (or stillstand) in the geomorphological record, which they termed the *Valley-glacier* and *Cirque-glacier* stages. They correlated these events with the Older Dryas and Younger Dryas chronozones, respectively. Norðdahl and Einarsson (1988) mapped several end moraines from Reydarfjörður in the north to Hornafjörður in the south (Fig. 3.5). The three successive glacial stages, *Faskrudsfjörður, Breiðdalsvík*

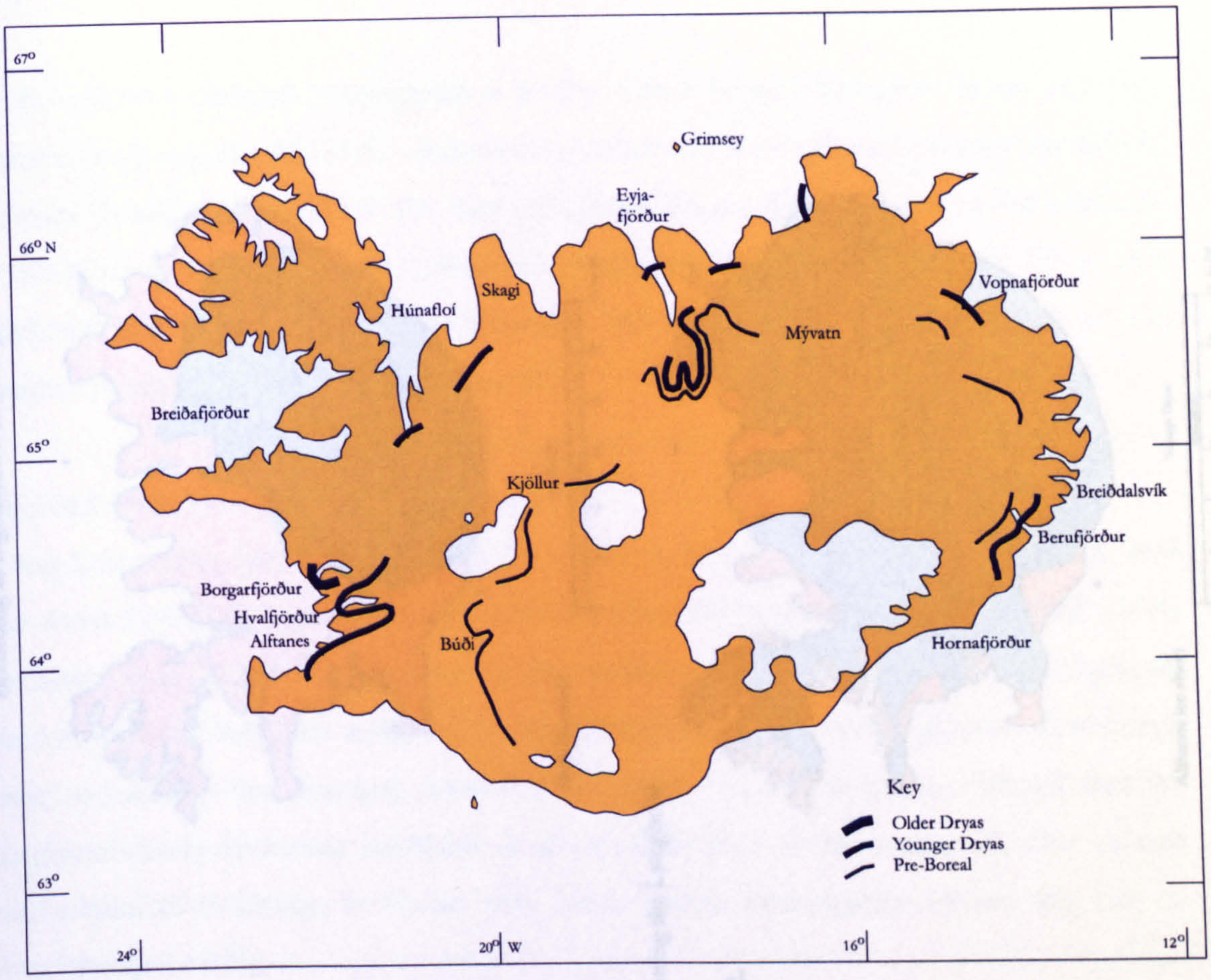


Figure 3.4 Ice marginal positions in Iceland during the Lateglacial and Early Holocene based on moraine evidence (after Ingolfsson and Norðdahl, 1994; and Ingolfson *et al.*,1997).

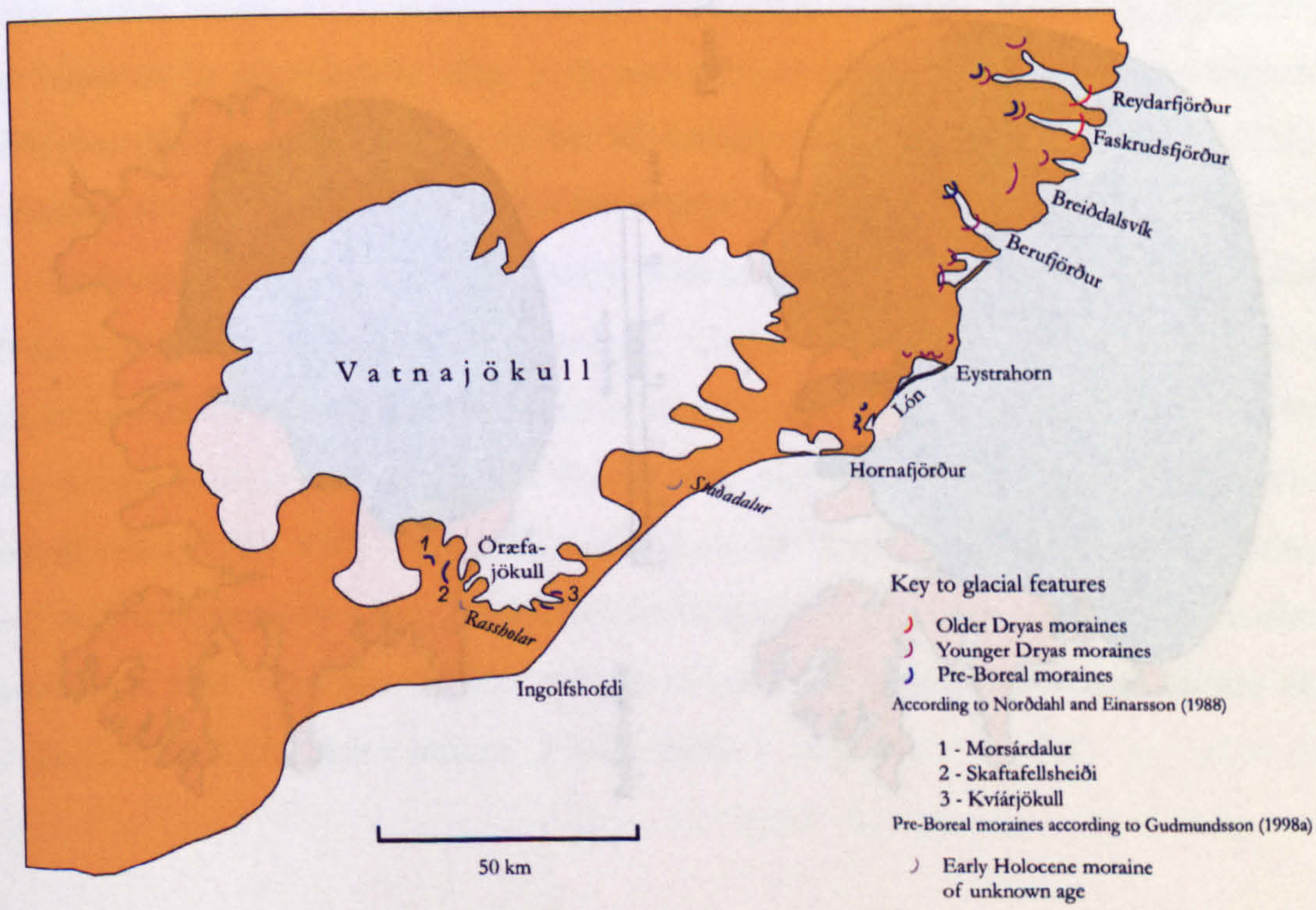
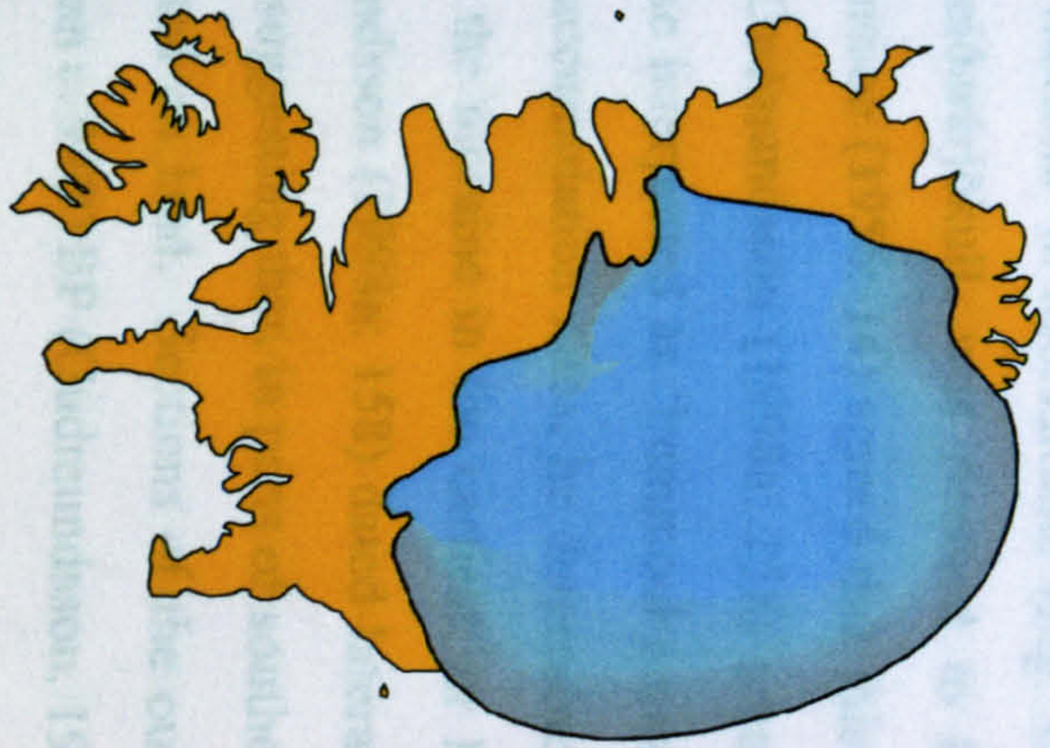
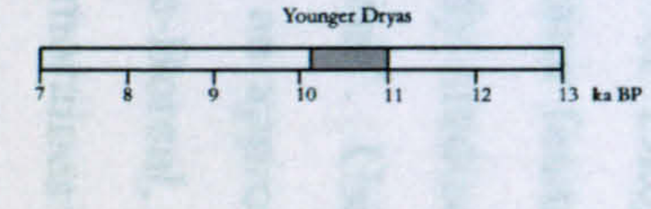


Figure 3.5 Previously identified Lateglacial and Early Holocene moraines in southeast Iceland.

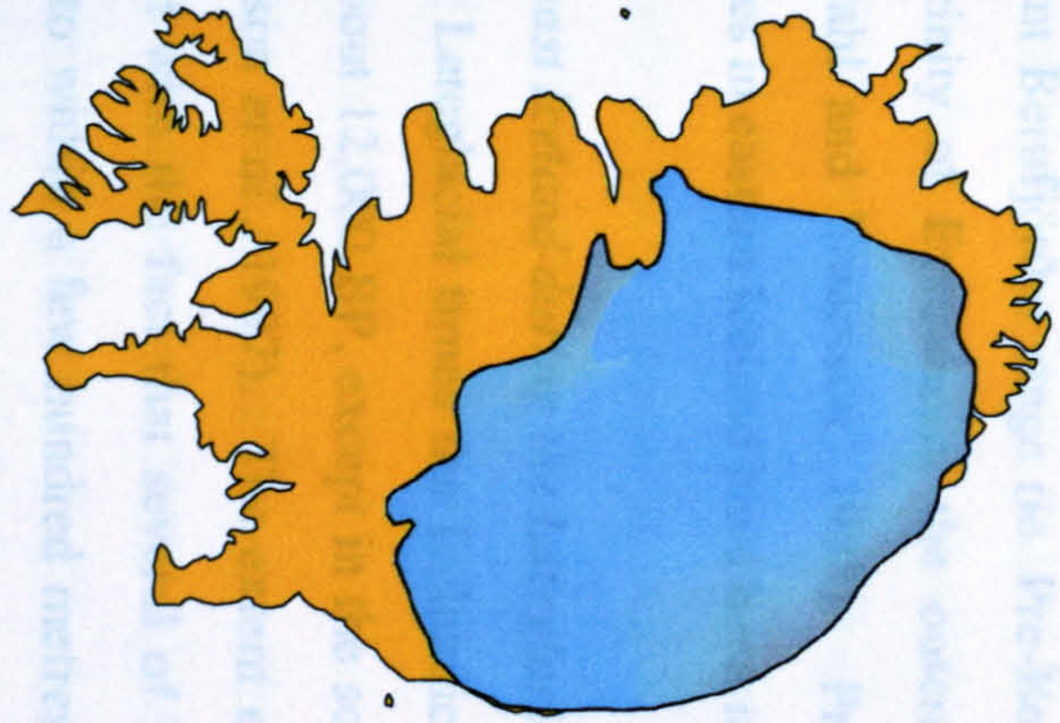
According to Einarsson (1979)



Buði ice-sheet



According to Ingolfsson *et al.* (1997)



Buði Ice sheet

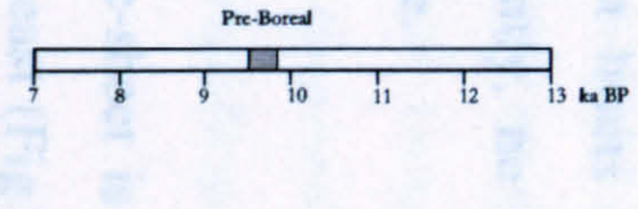
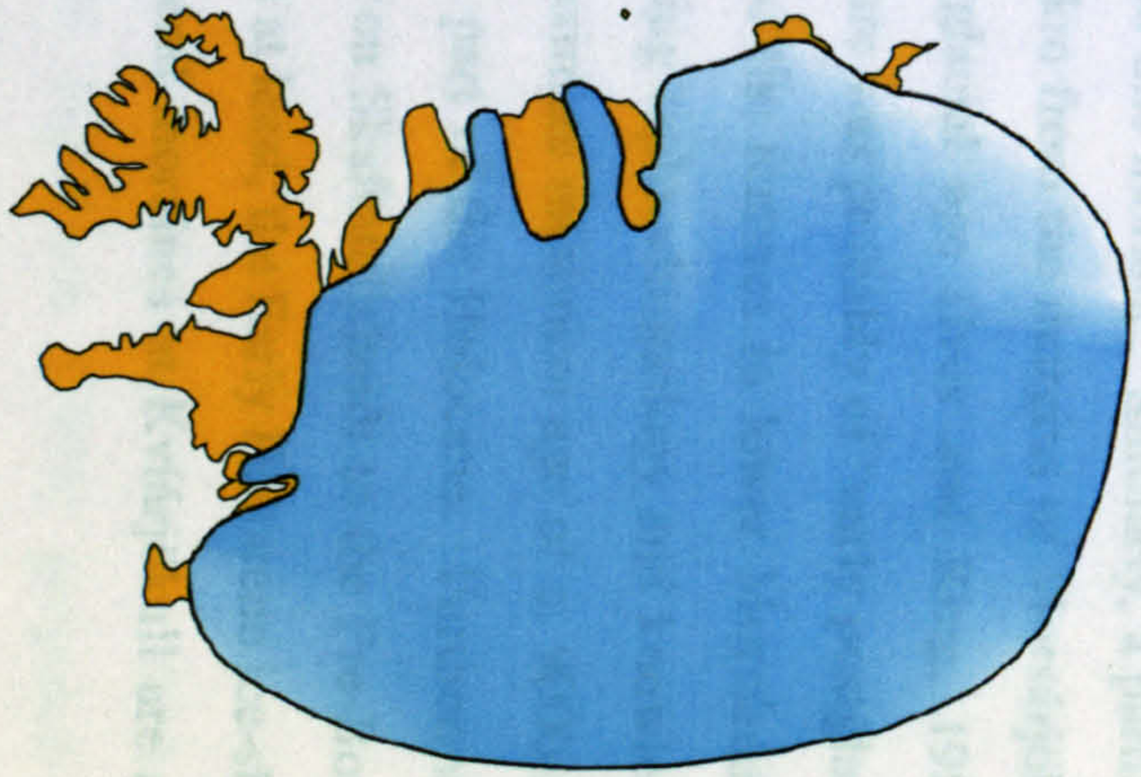
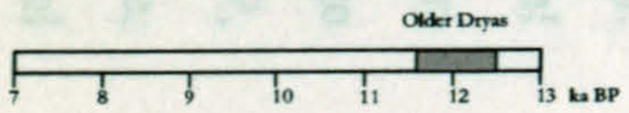


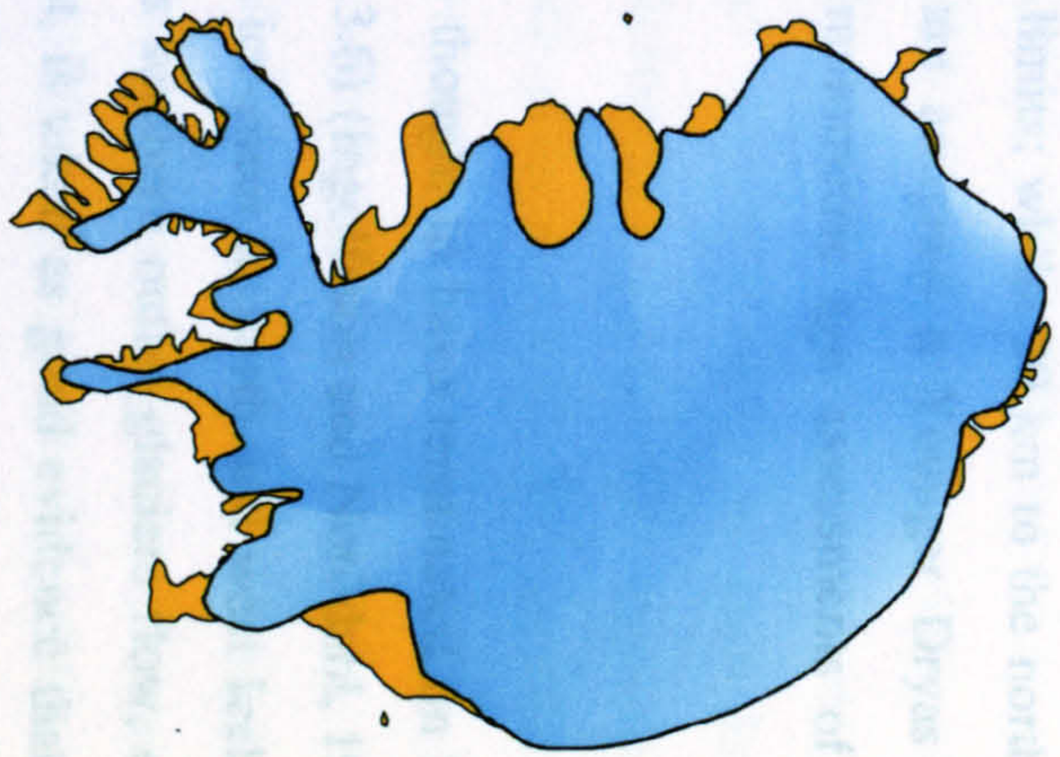
Figure 3.6 Ice cover during the Lateglacial

50 km

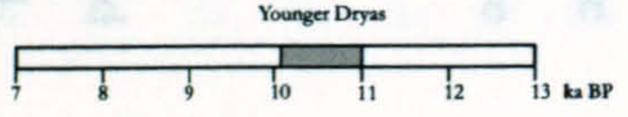


Alftanes ice-sheet





Alftanes ice sheet



and *Berufjörður*, broadly correspond with the Older Dryas, Younger Dryas and Pre-Boreal climatic events (Fig. 3.5). Around Hornafjörður, terminal moraines are thought to represent Berufjörður stage (ie. Pre-Boreal) glacial limits; whilst 20 km to the north, in the vicinity of Eystrahorn, the outermost limits are assigned a Younger Dryas age (Norðdahl and Einarsson, 1988). Presently, no radiometric age assessments of the moraines in eastern Iceland have been made.

Southeast Iceland during the Lateglacial

During Lateglacial times the Icelandic ice-sheet is thought to have terminated on land after about 12,000 BP, except in the southeast (Fig.3.6) (Ingolfsson and Norðdahl, 1994; Ingolfsson *et al.*, 1997). The extent of the inland ice-sheet in north and west Iceland, coupled with the fact that several of Vatnajökull's southern outlet glaciers flow, even today, to within a few hundred metres of the coast, is taken as good evidence that the Lateglacial–Early Holocene ice-sheet terminated offshore in the southeast. One or two small pieces of evidence, however, may need further investigation before this can be proved unequivocally.

Firstly, a terminal moraine in Staðadalur, 5.5 km from the present-day glacier, was thought, by Thorarinsson (1956), to probably pre-date the Postglacial Warm Period (Fig. 3.5). No further studies have yet been conducted on this moraine. Similarly, a partially buried 'moraine' in Morsárdalur (Fig. 3.5), now 5 km from the margins of Morsárjökull and Skeiðarárjökull, was deemed to be of Lateglacial age (Ives and King, 1955). Thorarinsson (1956: 14) agreed that this latter feature was probably of "early postglacial age". Gudmundsson (1998a: 158) recently studied the features in lower Morsárdalur, which he interpreted as 'hummocky moraine'. Using tephrochronology and knowledge of soil accumulation rates, he determined an approximate minimum age of c. 9000 BP, placing the moraine in the earliest, or Pre-Boreal, part of the Holocene. Furthermore, Gudmundsson (1998a: 158) dated lateral moraines on Skaftafellsheiði to the Pre-Boreal period, suggesting that, in parts of southeast Iceland at least, the Early Holocene ice-sheet terminated on land. Sections of the outermost lateral moraines at Kvíárjökull are also older than c. 9400 BP (Gudmundsson, 1998a: 168).

The final features in southeast Iceland deemed problematic by Thorarinsson (1956) are pre-historic, terminal moraine fragments, known as *Rasshólar*, on the sandur plain 2.5 km in front of Virkisjökull (Fig. 3.5). He inferred that these are probably "of the same age as the moraines in Morsárdalur and Staðardalur" (1956: 14). Gudmundsson (1998a: 166) has recently derived a date of at least 5-6000 BP for the advance associated with these moraines. However, Gudmundsson dated lateral moraines which he then linked tentatively to the *Rasshólar* moraines on geomorphological grounds. The exact age of these terminal moraine fragments may be significantly older than 6000 BP, possibly also placing them in the Pre-Boreal period.

Currently, apart from the very small number of cases mentioned above, the southeast coast of Iceland between Mýrdalsjökull and Hornafjörður presently lacks evidence of land-terminating glaciers during the Lateglacial–Early Holocene (13-8ka BP). Therefore, it has been reasonably assumed (Ingolfsson and Norðdahl, 1994; Ingolfsson *et al.*, 1997) that in southeast Iceland during Late-Weichselian deglaciation, the Older Dryas (*Skípanes*) and Younger Dryas (*Alftanes*), and probably Pre-Boreal (*Buði*), ice-sheets covered the entire landscape and extended some distance offshore.

Other evidence and 'ice-free' areas

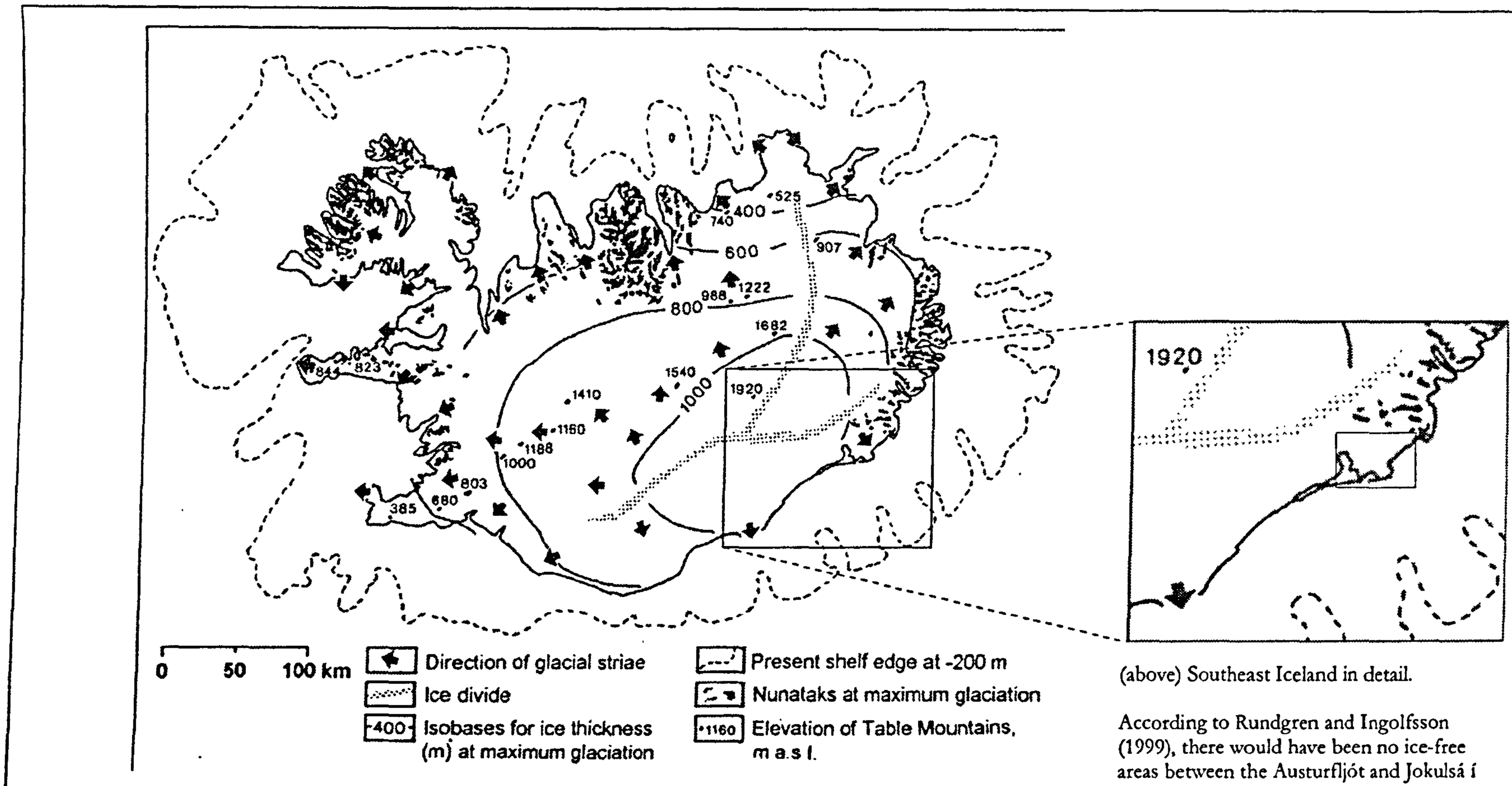
Several workers have proposed the concept of ice-free areas in Iceland even during the most extensive periods of Late Weichselian glaciation (eg. Thorarinsson, 1937; Norðdahl, 1983, 1991; Sigurvinsson, 1983; Ingolfsson, 1988; Rundgren and Ingolfsson, 1999). Geomorphological indicators such as alpine, unscoured, landscapes, deep weathering mantles and apparently ancient periglacial features are often taken as evidence that an area escaped glaciation during the Weichselian period. Furthermore, glacial trimlines on table-mountain volcanoes – which once erupted beneath a considerable thickness of ice – have been used to reconstruct the minimum ice thickness in parts of central Iceland (Einarsson, 1960; Hoppe, 1968). This information, coupled with knowledge of ice-sheet dynamics, allows the probable minimum ice-sheet thickness to be derived in surrounding areas. Rundgren and Ingolfsson (1999) proposed that relatively large mountain tops, or nunataks, protruded through the ice in north, west

and eastern coastal areas during the Last Glacial Maximum in Iceland (Fig. 3.7). There were apparently no nunataks in south coastal areas of Iceland from Thórlakshöfn in the west to Lónsfjörður in the east. However, the extensive, high plateau areas in the north and west such as Tröllaskagi and Vestfirðir show no evidence of glacial erosion or deposition and may have remained largely ice-free (Rundgren and Ingolfsson, 1999: Fig.3).

Hoppe (1982) found evidence of glaciation in the peripheral areas of Iceland. In the eastern fjords, glacial striations and streamlined bedrock were observed on the eastern extremity of Eystrahorn. Previously, Kjartansson (1964) proposed that at least part of this elevated promontory, and a further one between Berufjörður and Hamarfjörður, had probably remained ice-free during Weichselian times. Rundgren and Ingolfsson (1999) also suggest that these coastal mountains, immediately north of Lónsfjörður, would have supported ice-free enclaves during maximum glaciation (Fig. 3.7). Due to a lack of dating constraints, it is presently impossible to determine whether those striations observed by Hoppe (1982) at Eystrahorn relate to the last great ice-sheet or whether they represent a previous, more extensive, Quaternary glaciation.

Summary

In summary, glacial evidence indicates that disintegration of the Weichselian ice-sheet in Iceland was interrupted at various intervals. Re-advances, or stillstands, of the ice-margin left at least three distinct moraine systems on the Icelandic landscape. Examination of the Greenland ice-core (GRIP) oxygen-isotope record, a proxy of air temperature over the last 100,000 years (Dansgaard *et al.*, 1982), shows that these three main Lateglacial events may correspond with brief climatic reversals at around c. 12,000, 10,500 and 9700 BP. However, this multi-stage re-advance model of deglaciation, shown in Figure 3.8, may be more complicated on a regional scale due to local topographic or climatic factors.



(above) Southeast Iceland in detail.

According to Rundgren and Ingolfsson (1999), there would have been no ice-free areas between the Austurfljót and Jokulsá í Lóni rivers (shaded box). Although Eystrahorn (716 m asl) was probably ice-free during the Last Glacial Maximum, Vestrahorn (888 m asl) was not.

Figure 3.7 Ice thickness and ice-free areas at the Last Maximum Glaciation in Iceland (from Rundgren and Ingolfsson, 1999: Fig. 3).

Isobases for ice thickness are primarily based on the height of table-mountains above their surroundings, checked against field data on nunataks in coastal regions. Compilation based on Einarsson (1963), Hoppe (1982), Hjört *et al* (1985), Ingolfsson (1988), and Norðdahl (1991), amongst others.

Ice-free areas would have been those mountain summits and plateaux in coastal areas between ice-streams draining from fjords. The Tröllaskagi Highlands, parts of the Northwest plateau and the mountain tops surrounding the Eastern Fjords are most likely to have remained ice-free throughout the Last Glaciation (Rundgren and Ingolfsson,1999).

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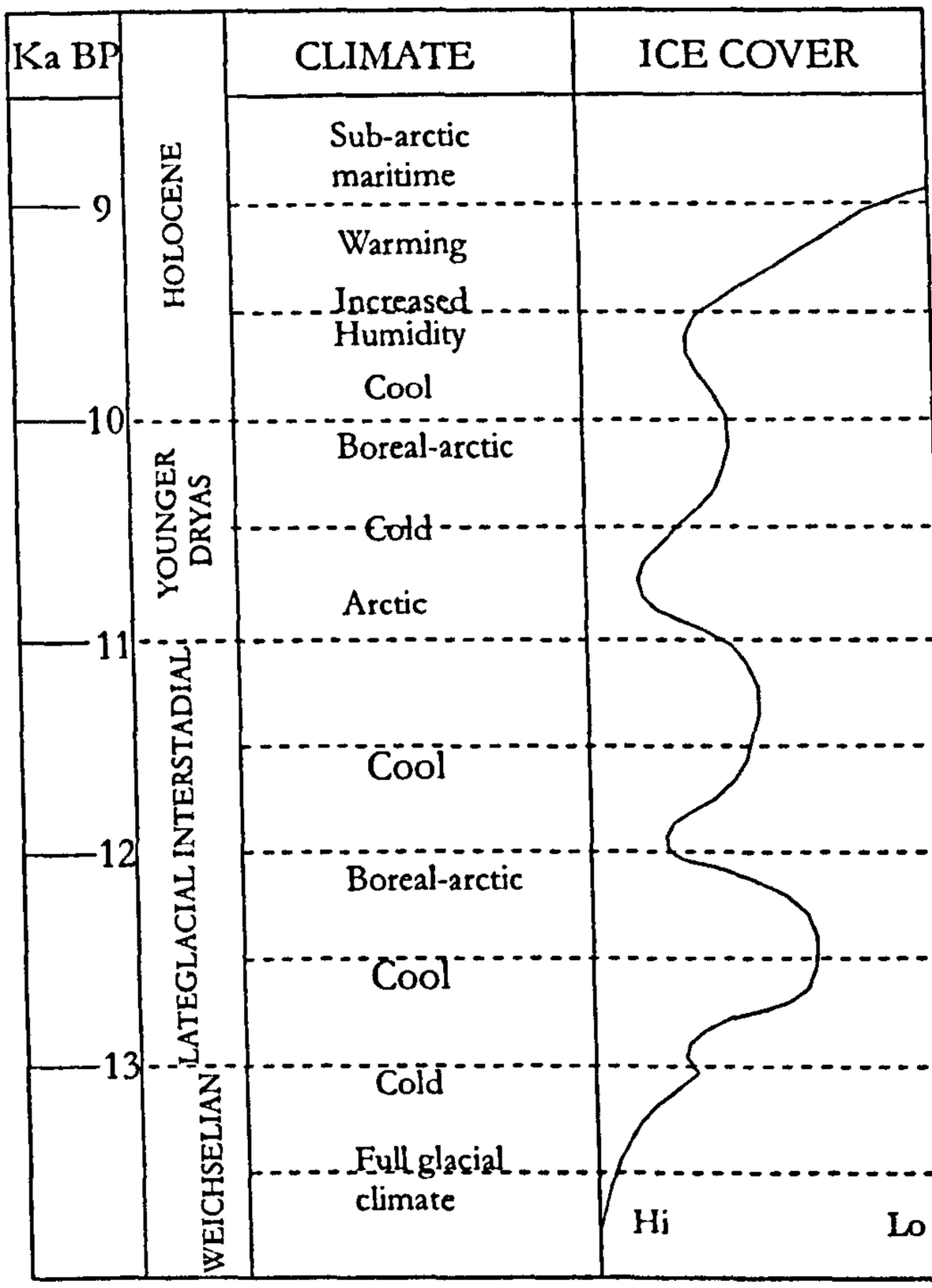

Figure 3.8 A summary of Late Weichselian and early Holocene (9-13 ka BP) climatic change and ice-sheet fluctuations in Iceland (taken from Ingolfsson *et al.*, 1997).

Four main periods of glacial expansion can be identified:
Weichselian ie. Last Glacial Maximum
Lateglacial interstadial ie. Older Dryas or *Skípanes* stage
Younger Dryas ie. *Alftanes* stage
Pre-Boreal ie. *Búði* stage

Summary and implications

The main periods of glacier expansion, so far identified, in Iceland since the onset of deglaciation (~13,000 BP) have been outlined (Fig. 3.3). Although not entirely coincident, the timing of glacier advances in the mid-Holocene does suggest that the peak of 'Neoglaciation' in Iceland at c. 5-7000 BP was an island-wide event, affecting both the north and south alike. Immediately prior to this time, glaciers were probably smaller than at any time since the last Interglacial period some 100,000 years ago. Some researchers have suggested that Iceland may have been completely ice-free for a brief period during the so-called 'Holocene Optimum' (Björnsson, 1979; Kaldal and Vikingsson, 1991; Ingolfsson *et al.*, 1997). Subsequent ice-marginal oscillations, notably at around 3000 and 1200-2000 BP, are observed at various locations. The 'historical' period saw two main phases of glacier growth culminating around 1200 AD (although evidence for this is sparse) and during the 18th and 19th centuries, most notably between AD 1850 and 1900. This second phase of renewed glacier activity corresponded with a marked fall in global temperatures often referred to as the *Little Ice Age* (c. AD 1600-1900).

This review demonstrates that glaciers in Iceland clearly respond to wider climatic changes. However, the exact relationship between glaciers and climate is not straightforward. Certain 'surging' glaciers are not driven by climatic change but by internal ice dynamics (eg. Kamb 1987; Boulton and Clark, 1990; Sigurdsson, 1998). Furthermore, those glaciers overlying active volcanoes exhibit a complex response affected by high geothermal heat fluxes and the subsequent production of large volumes of subglacial meltwater (Ahlmann and Thorarinsson, 1937; Kjartansson, 1964; Björnsson, 1988; Sturm *et al.*, 1991).

Having summarised the evidence regarding glacier fluctuations in Iceland since the termination of the Last Ice Age, several key hypotheses have emerged. These hypotheses resolve into six testable statements:

1. An ice-sheet covered much of Iceland during the Lateglacial period (9-13 ka BP), extending offshore in the south and southeast.

2. Glacier advances during the Holocene (last 9000 yrs) exceeded those of the Historical Period (last 1000 years) in Iceland.

3. Glacier advances during medieval times (c. AD 900-1400) did not exceed those of the *Little Ice Age* (c. AD 1600-1900) in Iceland.

4. The *Little Ice Age* glacial maximum culminated in the late 19th century in Iceland.

5. The fluctuations of non-surging glaciers in Iceland are the result of climatic changes rather than volcanic/geothermal factors.

6. Glacier fluctuations can be used to reconstruct the pattern and magnitude of climatic change on decadal and centennial timescales.

Chapter 4 : The Study Site and Methods

Part I – The Study Site

Selection of Study Area

In order to test the hypotheses outlined at the end of Chapter 3 it was important to select a geographical area that could fulfil several basic criteria: –

1. The area contains a glacier.
2. The glacier is located outside the Neo-volcanic zone and sufficiently far away from high-temperature areas to reduce the effects of geothermal heating on glacier flow.
3. The glacier is not prone to *jökulhlaup*-floods, thus making moraine preservation more likely.
4. The glacier has a high mass turnover, and is thus sensitive to changes in climate.
5. The glacier is non-surging and therefore more likely to reflect climatic change.
6. The glacier possesses a topographically-constrained accumulation area, making problems relating to ice-divide migration less likely.
7. The glacier terminates in a valley, rather than as a broad unconfined lobe, making small changes in mass balance more likely to be expressed in changes in the location of the terminus.
8. The glacier has a good moraine record.
9. The area is accessible.

Only three areas of Iceland fulfil the first three criteria: eastern Vatnajökull, Drangajökull in the northwest and Tröllaskagi in the north. However, the glaciers of the Tröllaskagi highlands are located in the less maritime, sub-polar climate of northern Iceland. Turnover rates here are not ideally suited for the study of high-resolution glacier fluctuations. Although located in a more maritime climate, the glaciers of

Drangajökull are known to surge (Eythorsson, 1935; Sigurdsson, 1998) making them less appropriate for palaeo-climatic study. Only the glaciers draining the eastern flanks of Vatnajökull fulfil the first five criteria. The suitability of each of these will now be examined in more detail.

Table 4.1 Selection criteria for glaciers (see Fig.3.2 for location map):

Eastern Vatnajökull outlet glaciers	Surge type	Valley terminus	Topographically-constrained accumulation area	Moraine record	Accessibility
Breiðamerkurjökull	Y	N	N	good	good
Brokárjökull	N	Y	N	good	fair
Skálafellsjökull	N	N	N	good	good
Heinabergsjökull	N	N	N	good	good
Flaájökull	N	N	N	good	good
Viðbordsjökull	N	Y	N	unknown	difficult
Hoffellsjökull	N	N	N	fair	good
Lambatungnajökull	N	Y	Y	very good	difficult
Austurtungnajökull	N	Y	Y	poor	very difficult
Nordurtungnajökull	N	Y	Y	poor	very difficult
Öxarfellsjökull	N	Y	Y	good	very difficult
Vesturdalsjökull	N	Y	Y	unknown	very difficult
NE margin (un-named)	N	N	N	unknown	very difficult
Eyjabakkajökull	Y	N	N	good	difficult

Of eastern Vatnajökull's 14 separate outlet glaciers (excluding the nine of Öræfajökull – draining an active volcano), 12 fulfilled the fifth criteria (ie. non-surging and therefore more likely to reflect climatic change). Of these 12, only two fulfilled the first eight criteria: – Lambatungnajökull and Öxarfellsjökull. (Austurtungnajökull, Norðurtungnajökull and Vesturdalsjökull exhibit relatively short or uncertain moraine records.) Öxarfellsjökull is almost inaccessible and appears to have a less impressive moraine record, hence Lambatungnajökull was chosen as the ideal focus of this study.

The Study Area

Lambatungnajökull

The glacier at the centre of this investigation is Lambatungnajökull, a little-studied, glacier draining from eastern Vatnajökull (Fig. 4.1). Due to its relative inaccessibility and the nature of the surrounding terrain, Lambatungnajökull is seldom visited. It is a medium-sized, temperate glacier with dimensions similar to those of Skaftafellsjökull and Heinabergsjökull (Fig. 3.2). Lambatungnajökull presently flows in a relatively narrow, U-shaped valley, called Skyndidalur. The floor of this trough is up to 600 m below the surrounding uplands. To the north, Skyndidalur is bounded by the high, alpine, ridge of Múlatindur. To the south, the valley is cliff-lined with the palagonite plateau of Dalsheiði above (550 m asl). Skyndidalur curves through 45 degrees and runs SE for much of its 10 km length. Unlike many other glaciers in Iceland which terminate as a broad lobe on a level outwash plain, Lambatungnajökull is confined by precipitous valley walls to the north and south. The overall surface long-profile of the glacier is steep, falling almost 1000 m vertically in only 10 km distance. The lateral margins of the glacier are also steep and highly crevassed. The accumulation area is not extensive and is well constrained by the underlying topography. The ablation area – defined by the late-summer snowline – represents only 30-40% of the glacier's total area and is in close agreement with the figure derived empirically by Porter (1977) for glaciers in North America.

Although the ice-flow is presently confined to one valley there is geomorphological evidence to suggest that during previous periods of glacier growth the ice breached a col in the valley side and over-flowed into the neighbouring valley of Hoffellsdalur. This valley is now ice-free. Hoffellsdalur runs due south, almost perpendicular to Skyndidalur, and is less than 2km wide and almost 10 km long. Like Skyndidalur, Hoffelldsalur is flanked by mountains of considerable height, Grasgiljatindur (1275 m) and Hoffellsfjall (850 m) to the west and Osdalsheiði (970 m), Reipstindur (1046 m) and Seltindur (998 m) to the east.

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Geological setting

Situated astride the mid-Atlantic ridge, Iceland has formed during the last 60 Million years from outpourings of lava. Figure 4.2 shows the main geological divisions in Iceland. The central region running from southwest to northeast represents the active volcanic zone. Immediately to the east and west of this region are subaerial and subglacial lavas of Plio-Pleistocene age. The eastern and westernmost areas are comprised of stacked flood basalts at least 3.3 million years old. These oldest rocks have been tilted and subsequently faulted by more recent tectonic activity.

The study area lies within the Tertiary Basalt Formation of Eastern Iceland. However, the area also contains the relics of at least four central volcanic systems. Extrusive rocks of acidic composition, principally rhyolite, mark the former extent of central volcanoes at Illikambur in Lónsöræfi, Lambafell in Lón, Laxárdalur in Hornafjörður, and Efstafell adjacent to Hoffellsjökull (Fig. 4.3 and Fig. 4.4). These volcanic systems were probably last active during the Late Tertiary (3Ma BP). Associated with these ancient volcanoes are persistent lateral intrusions of igneous rock, or *dykes*, which cross-cut the lava pile vertically. These dykes, usually composed of dolerite, are orientated predominantly NE-SW in accordance with the tectonic structure of the surrounding country rocks. In the region around Hoffellsdalur and in Seldalur, south of Lambafell, these dykes are highly concentrated into swarms (Fig. 4.4)

The Hornafjörður district also possesses more unusual igneous rocks. Magma of plutonic origin, exhumed at the surface, is evident at several localities in the form of granites and gabbros. At Illikambur and at Svínafell, in front of Hoffellsjökull, plutonic rocks are thought to represent the ancient heart of two large volcanic calderas. The imposing mountains of Vestrahorn and Eystrahorn are chiefly composed of gabbroic intrusions and are also the remnants of extinct volcanic systems (Fig. 4.3). The largest intrusive mass of plutonic rocks in Iceland takes the form of a *batholith* and is found in the mountains immediately north of Vestrahorn. Smaller exposures of granitic rocks occur on Reydarártindur in Lón and in lower Hoffellsdalur (Fig. 4.3).

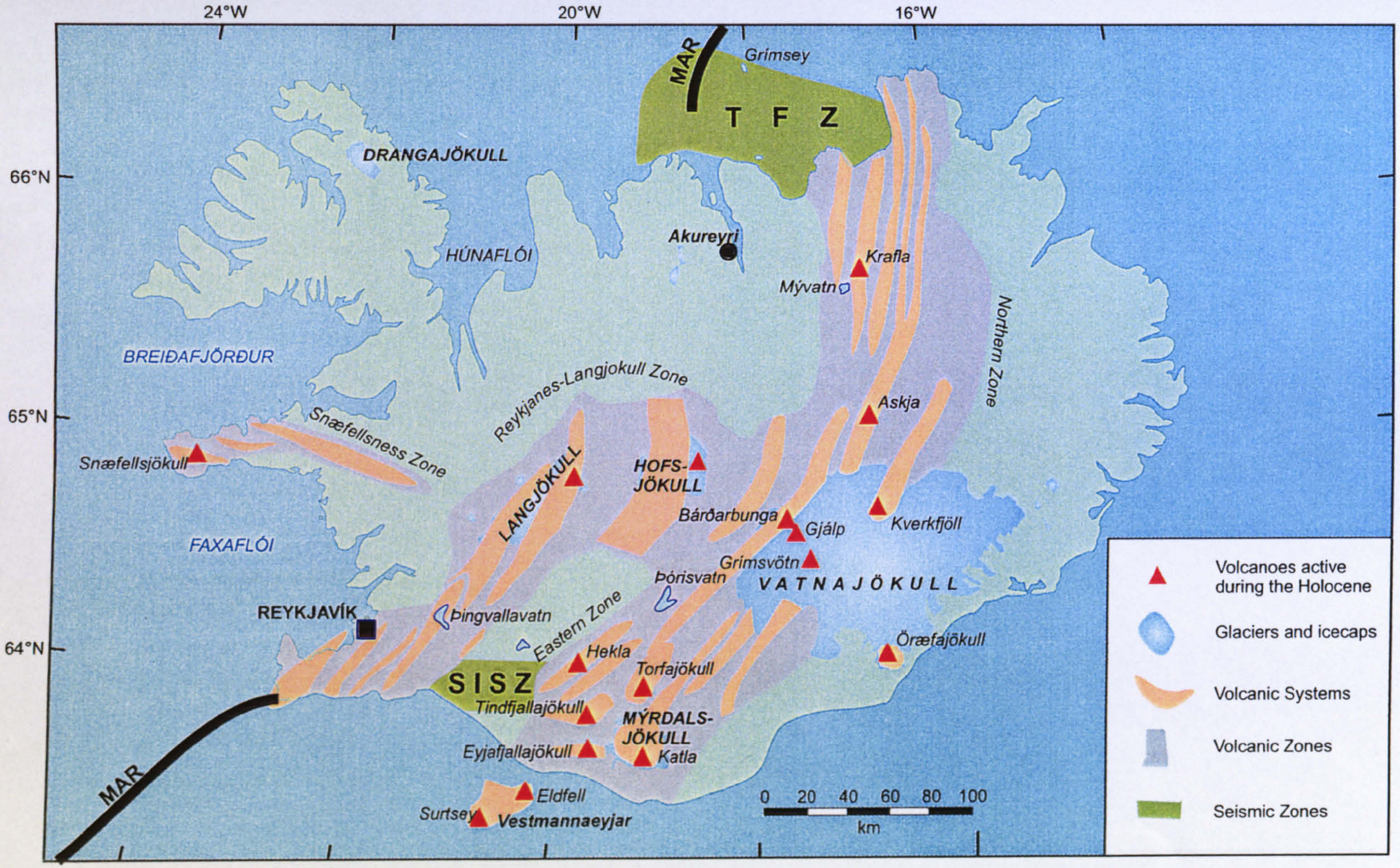


Figure 4.2: Map showing the generalised geology of Iceland. Green area composed of volcanic rocks older than 0.7 million years. Grey area is comprised of rocks younger than 0.7 million years. The active volcanic systems are shown in orange. MAR = Mid-Atlantic Ridge, SISZ = South Iceland Seismic Zone, TFZ = Tjörnes Fracture Zone. Taken from Newton (1999).

Key to geology

- Basic and intermediate extrusive rocks of Tertiary age (>3.3 Ma) - primarily basalts
- Basic and intermediate extrusive rocks (>0.8<3.3 Ma) - including Palagonite
- Acid extrusives (>11ka)
- Acid intrusions, rhyolite, granite, etc.
- Basic and intermediate intrusions, gabbro, dolerite, etc.
- Fluvial sediments
- Glaciers

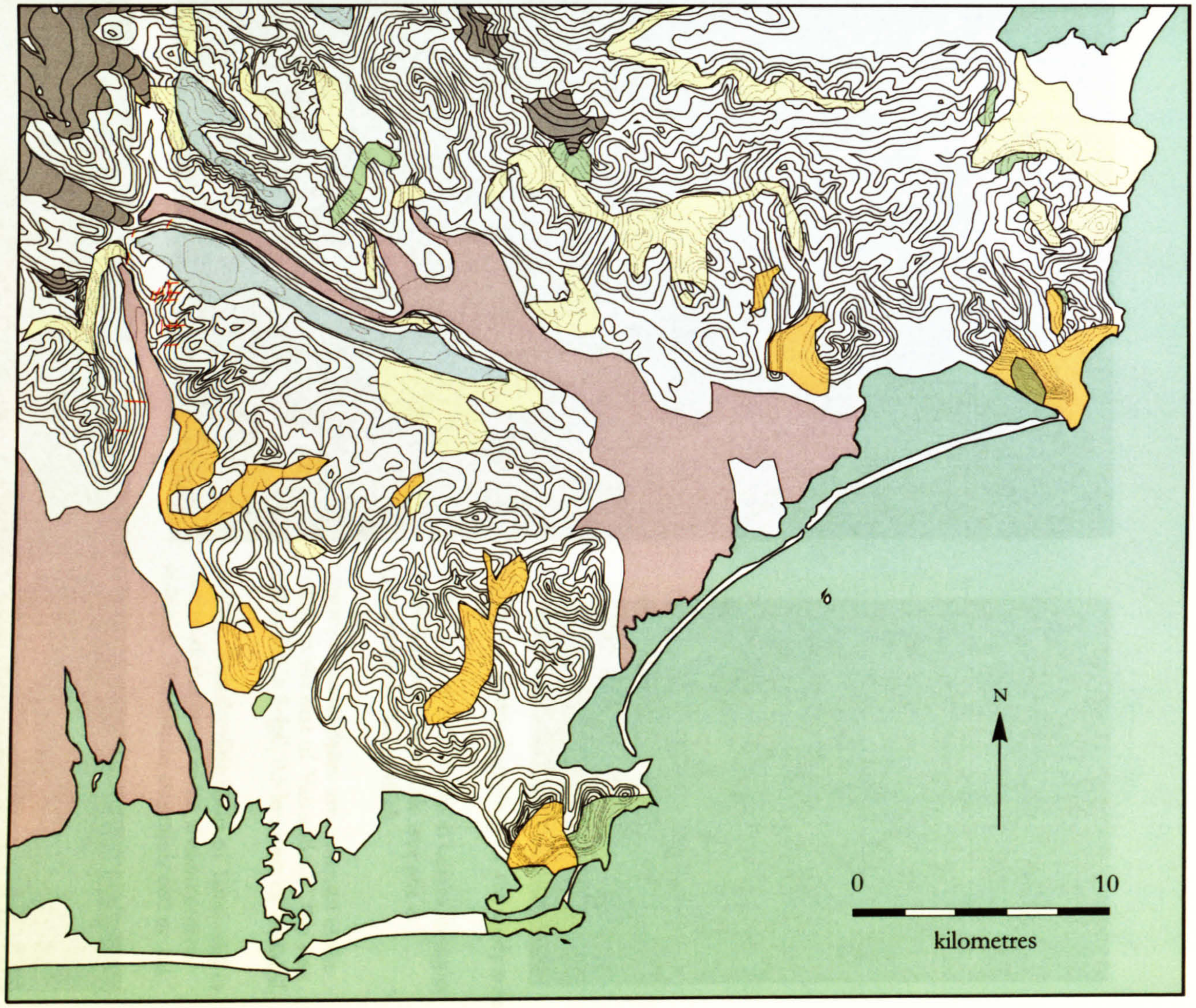
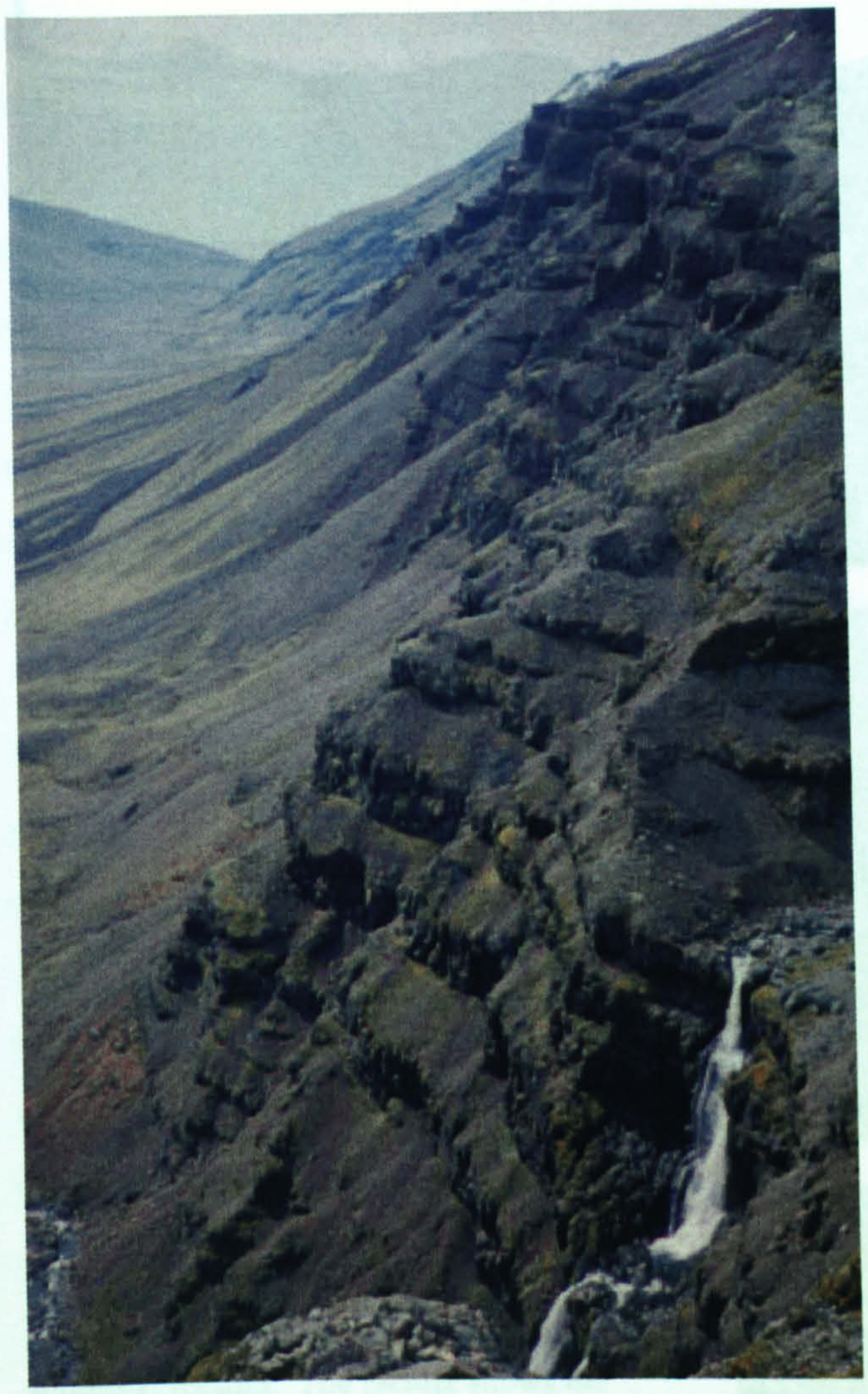


Figure 4.3 The geology of the Hornafjörður and Lón districts.

The map is based on the Geological Map of Iceland 1 : 500,000 sheet – compiled by Johannesson and Saemundsson, 1998.



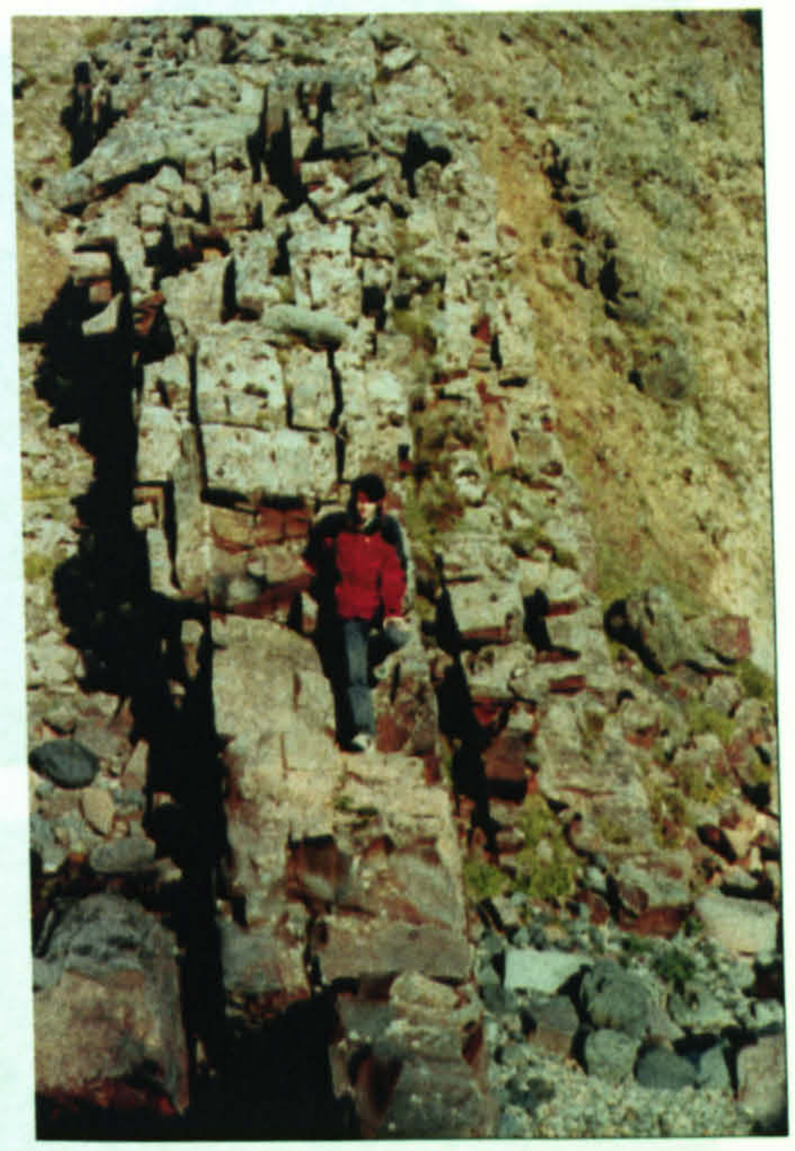


Figure 4.4 Bedrock geology of the study area.

(left) Stacked Tertiary flood basalts in upper Hoffellsdalur. The visible lava pile is approximately 300 m thick.

(above) Dolerite dykes are common in the valleys to the east of Vatnajökull. The one pictured is in Selgil, 0.5 km west of Stafafell.

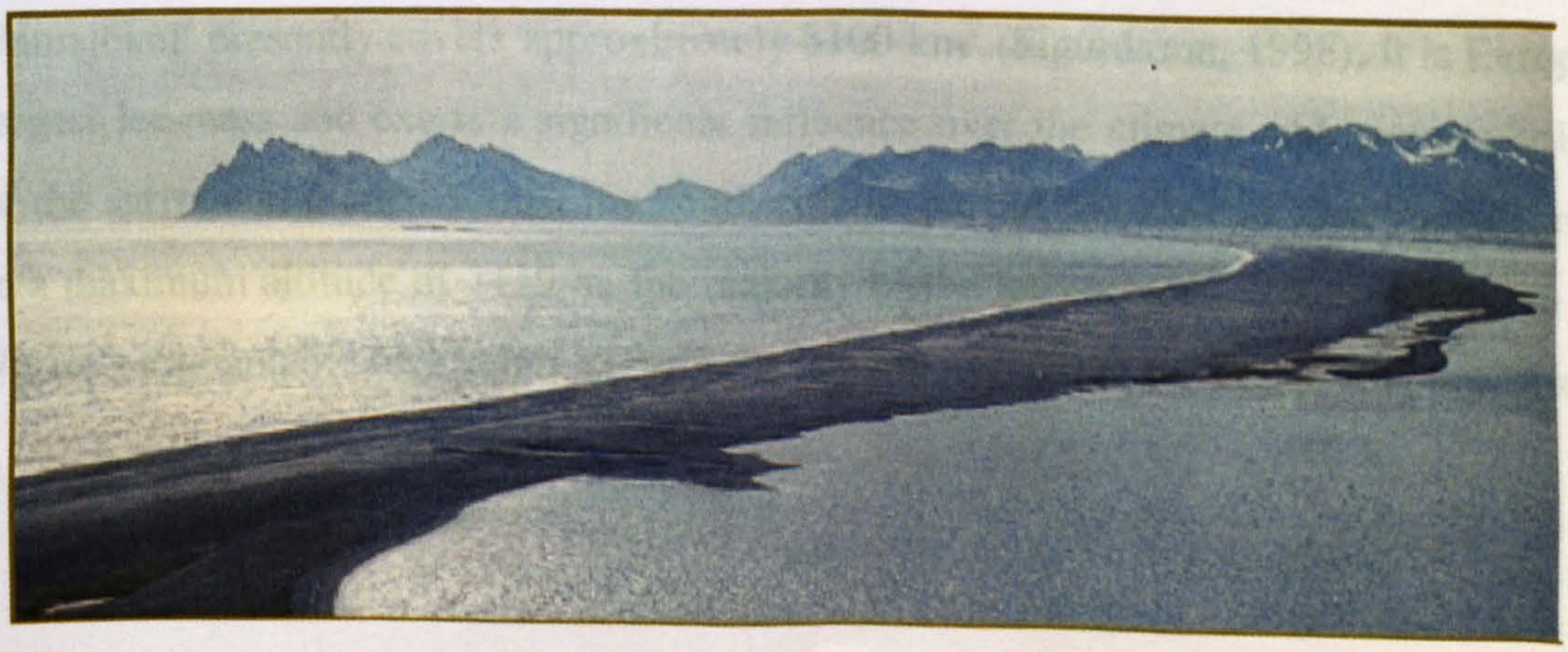
(below) The rhyolitic rocks of Lambafell in Lon mark the heart of an ancient central volcano. The distant summit is 680 m asl.

Figure 4.5 Superficial geology of the study area.

(uppermost) Fluvioglacial sediment or *sandur*, deposited by braided glacial rivers (*Jokulsâ*), covers a large portion of the coastal plain in southeast Iceland. The picture shows the Jokulsá í Lón.

(above and right) The scree-covered flanks of Vestrahorn (888 m asl).
Differing rocktypes – gabbro and granophyre – produce screes of different colour.

(below) Evidence of marine sedimentation – 15km long barrier-spit at Lónsfjorður formed by wave action (Photo - Th. Einarsson, 1994)



Although the bulk of the terrain in the study area is composed of Tertiary flood basalts (>3.3 Ma), isolated exposures of younger rock exist in at least three localities. Volcanic activity beneath glacier ice produces a distinctive suite of igneous rocks. *Palagonite,* a glassy, brecciated, conglomeratic rock, is lithological evidence of subglacial eruptions. Pliocene-Pleistocene age (0.8-2.5 Ma) rocks of this type are found flanking Skyndidalur, most notably on the plateau of Dalsheiði and on the peaks of Múlatindur (Fig. 4.3). A small exposure of Pliocene palagonite also caps Fossdalshnúta, the flat-topped mountain adjacent to Lambatungnajökull. These younger deposits, overlying the Tertiary basalts, have survived erosion during the intervening millennia.

The superficial geology of southeast Iceland is dominated by glacial and fluvial sediments. The outwash of the Jokulsa í Lón and Austurfljót river systems form extensive sandur plains consisting of sand and fluvial gravel (Fig. 4.5). Approximately one-third of the study area is overlain by fluvio-glacial sediment. The exact thickness of these deposits is unknown although, at the mouth of the Jokulsá i Lón, sections reveal at least 10 m of fluviatile sand and gravel. Glacial till is restricted to those hillslopes away from fluvial activity and generally below the 500 m contour. Large amounts of scree cloak the steepest valley sides. High mechanical weathering rates due to frost shatter and freeze-thaw mechanisms constantly add to the supply of rockfall material. The abundance of less resilient, acidic, rock in the Hornafjörður district make scree slopes commonplace. Those of Vestrahorn, Eystrahorn and Hoffellsdalur are particularly impressive (Fig. 4.5). Quarrying of the talus-covered slopes flanking Skálatindur has revealed that the scree is in excess of 20 m thick near the base.

Glaciological setting

Vatnajökull

Vatnajökull presently covers approximately 8100 km^2 (Sigurdsson, 1998). It is Europe's largest ice-mass and exerts a significant influence over the climate and geomorphology of the surrounding area. Whilst the southernmost part of the ice-cap, Öræfajökull, rises to a maximum altitude of 2119 m, the majority of the ice surface lies below 1500 m asl. Various topographic highpoints exist in the ice-cap, reflecting the underlying topography

(Björnsson, 1979). More than 3 x 10^{12} m^3 of fresh water is locked up in the ice-cap. Vatnajökull is classified as a polythermal ice-cap, although has been termed 'temperate' (Björnsson, 1988; Th. Einarsson, 1994). It is difficult to assess whether the entire ice-mass is truly temperate (ie. that the ice is at melting point throughout). It is clear that the outlet glaciers and lobate margins generally behave as temperate glaciers. However, the distribution of temperature within the ice has not been studied fully. Measurements suggest that temperatures at the base of the ice-cap, in the Grimsfjall area, are at the pressure-melting point and basal melting rates are higher than was previously thought (Björnsson, 1988).

Possessing at least 38 discrete outlet glaciers, Vatnajökull's ice-margin measures over 2000 km (Fig. 3.2). Many of these glaciers are surge-type outlets and fluctuate independently of climate, particularly the large lobes in the north and west (eg. Brúárjökull, Dyngjujökull, Tungnaárjökull; Thorarinsson, 1943; Bjornsson, 1979, 1988; Sigurdsson, 1998). Some of the glaciers exhibit both surge-type and climatically-sensitive behaviour (eg. Breiðamerkurjökull, Eyjabakkajökull; Johannesson and Sigurdsson, 1998). The remainder, generally the smaller and steeper glaciers of the south and eastern margin, are not known to surge.

The northeast of Vatnajökull is almost a separate glacial centre. The glaciers are not fed by the central dome of the ice-cap, unlike the larger glaciers to the south. The underlying topography protrudes through the ice quite conspicuously in several places, reaching a height of 1570 m on Goðahnukkur. These nunataks probably represent the remnants of a former volcanic caldera (Saemundsson, *pers. comm.*, 2000). Although no geological investigations have been carried out on these 'ice-islands', aerial photography and satellite imagery shows that they may be the surface expression of an ancient, crater-like, feature. The glaciers of northeast Vatnajökull are constrained by this subglacial topography and a row of nunataks marks the ice-divide between the glaciers of Lambatungnajökull, Öxarfellsjökull and Eyjabakkajökull.

Lambatungnajökull

Lambatungnajökull is the largest and longest of the five glaciers flowing eastward from northeast Vatnajökull (Fig. 4.6). It covers an area of approximately 23 km^2 and measures 14.7 km from the ice-divide to the snout. Some other physical properties of the glacier at the centre of this research are shown in Table 4.2.

Table 4.2 Physical properties of Lambatungnajökull in 1982.

Dimension	*1982 value*
Surface area	22.62 km^2
Length along flow line	14.70 km
Width at 1000 m asl	2550 m
Width at 500 m asl	1000 m
Terminus width	350 m
Terminus altitude	200 m
Maximum altitude	1488 m
Median altitude	900 m
Vertical distance/horizontal distance (ie. gradient)	0.10
Altitude at Point A	340 m
Altitude at Point B	470 m
Altitude at Point C	580 m
Late summer snowline (ie. ELA)	1000 m
ELA (AAR = 0.6)	980 m
Accumulation area	13.6 km^2
Ablation area	9.0 km^2

Detailed maps produced in 1945 and 1982 show that Lambatungnajökull decreased in size by approximately 10% over the intervening ~40 years (Fig. 4.7). This reduction in area is largely due to the effects of climatic change causing a reduction in glacier thickness. The glacier terminus retreated c. 800 m over the same period (Figs. 4.6 and 4.7).

The glacier *equilibrium line* for northeast Vatnajökull is at a relatively low altitude, lying around 1000-1100 m above sea level (Fig. 2.12). According to Ahlmann and Thorarinsson (1937), only the glaciers of Drangajökull (600-750 m), in Northwest Iceland, and of the Snæfells volcano (900 m) have a lower ELA than Lambatungnajökull. Analysis of the topographic survey map from 1982 allows the ELA of Lambatungnajökull to be approximated as 980 m asl. The ELA of any glacier can be estimated from the hypsometry of the ice-flow. The most commonly-practised method of doing this relates to the ratio between the glacier accumulation area and the total

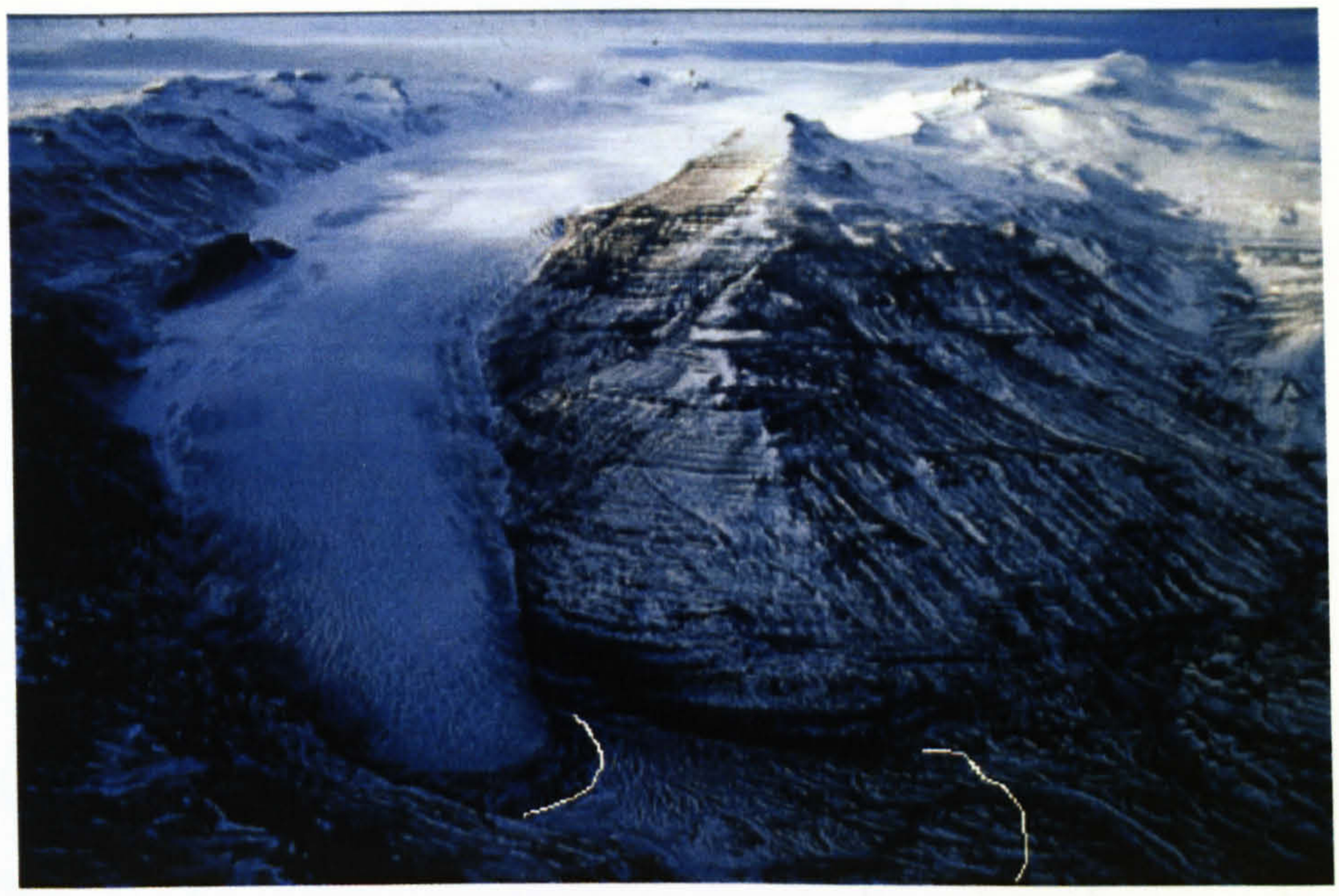
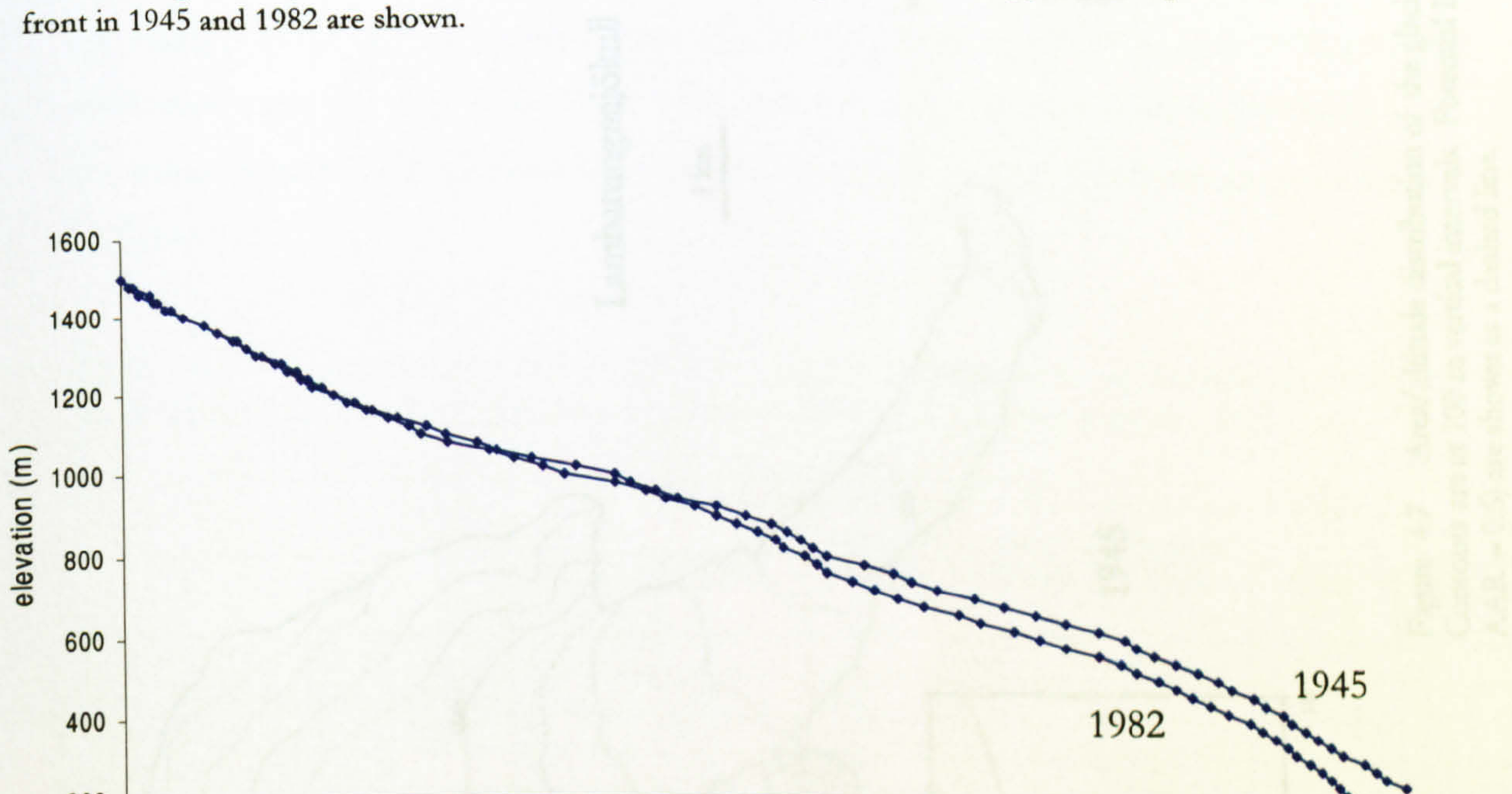


Figure 4.6 Oblique aerial photograph of Lambatungnajökull, taken in March 1990 looking north-west (courtesy of Oddur Sigurdsson). The Vatnajökull ice-cap is in the background. The approximate positions of the ice-front in 1945 and 1982 are shown.



(above) The longitudinal surface profile of Lambatungnajökull in 1945 and 1982.

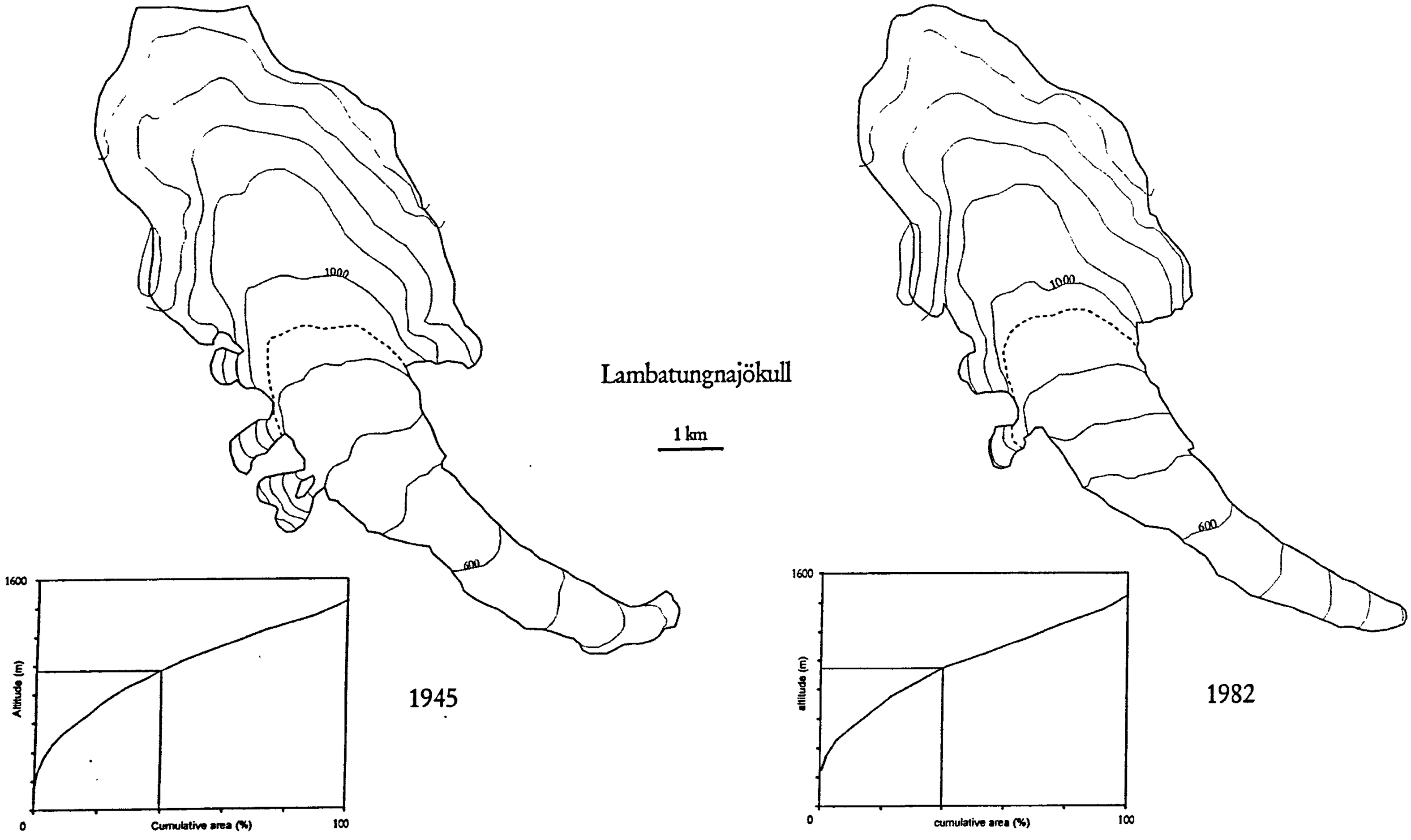


Figure 4.7 Area/altitude distribution of the glacier in 1945 and 1982. Contours are at 100 m vertical intervals. Potential ELAs (calculated using AAR = 0.6) are shown as a dashed line.

glacier surface area (AAR). Porter (1975) found an AAR of 0.6 to be characteristic of valley glaciers under steady-state conditions. Numerous workers have used a value of 0.6 to reconstruct former glacier ELAs (eg. Porter, 1975; Sutherland, 1984; Hawkins, 1985; Ballantyne, 1989). However, departures from this value have been found in glaciers possessing unusual hypsometries (eg. large accumulation area and short narrow tongue) (Meier and Post, 1962; Kuhle, 1988). Other methods have been used to calculate the ELAs of former and modern glaciers, including the *median elevation* (Manley, 1959; Ostrem, 1966), the *toe-to-headwall ratio* (Péwé and Reger, 1972; Murray and Locke, 1989), *Kuhle's formula* (Kuhle, 1988) and the *maximum elevation of lateral moraines* (Andrews, 1975). A critical evaluation conducted by Torsnes *et al.*, (1993) found all these methods to be generally less reliable than the AAR approach.

Changes in the margin of Lambatungnajökull, both laterally and vertically, were observed relative to a fixed point, over the 14-month study period. Photographs of the ice margin taken at two-monthly intervals in the summers of 1998 and 1999 provide additional evidence (Fig. 4.8). These observations could be continued in the future. All the measurements of horizontal ice-margin retreat and vertical lowering are summarised in Table 4.3.

Table. 4.3 Some measurements of ice-marginal fluctuations of Lambatungnajökull, made during the course of fieldwork between July 1998 and September 1999.

	Date of observation and total ice-margin retreat (metres) since the original observation						
	1/7/98	1/8/ 98	22/5/99	29/5/99	10/7/99	1/8/99	20/8/99
Terminus position A	0			1.0	1.3	1.3	
South margin position A (max)	0	0.5	3.0		3.6	4.0	4.6
South margin position A (min)	0	0.2	1.0		1.3	1.5	1.5
South margin position B	0	0.4	2.0		2.3	2.5	2.8
	Date of observation and total ice-margin surface lowering (metres) since the original observation						
	1/7/98	1/8/ 98	22/5/99	29/5/99	10/7/99	1/8/99	20/8/99
Terminus position A	0			1.0	1.0	1.0	
South margin position A (max)	0	0	0.6		0.8	1.0	1.2
South margin position A (min)	0	0	0.5		0.5	0.5	0.6
South margin position B	0	0	1.8		2.0	2.2	2.4

Figure 4.8 The southern margin of Lambatungnajökull in 1998 and 1999.

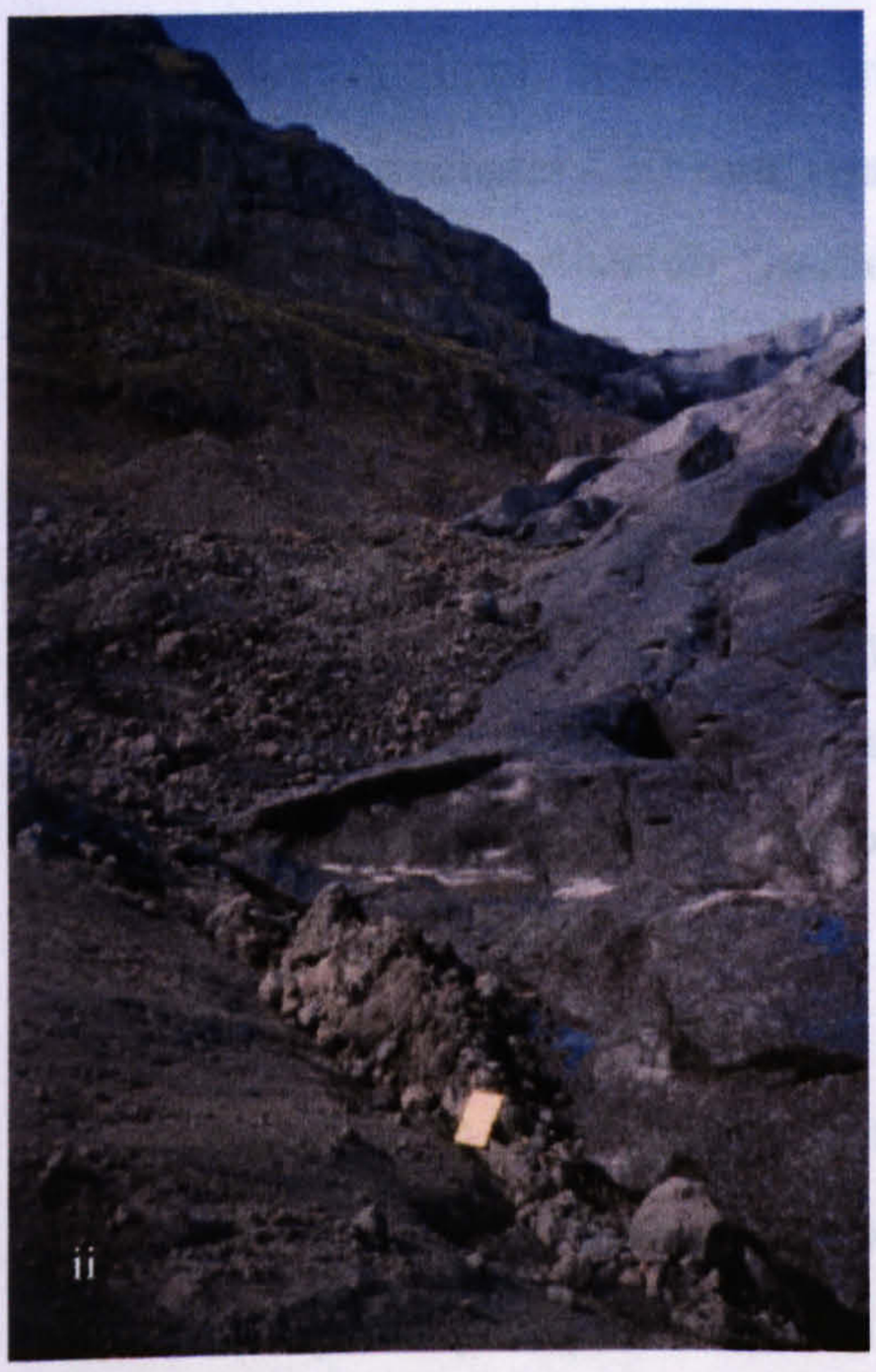
(i) June 1998

(ii) Late July 1998

(iii) Late August 1999

(iv) Late August 1999

Note the amount of recession that occurred over the 15-month period (c. 5 m).



Other features to note are the small moraine ridge (<0.5 m high) in *ii* that has formed by a combination of pressure-squeezing and bulldozing processes; the typical sedimentology of the most recently-deposited glacial debris in *iii*; and, in *iv*, the debris rich basal-ice layer exposed at the margin (person for scale). The boulder in the foreground (iii) was marked to allow the degree of ice-marginal retreat to be determined in subsequent years.

Previous mass balance studies conducted on Vatnajökull

Scientific investigations by a joint Swedish-Icelandic expedition to southeast Iceland in 1936, 1937 and 1938 determined the rate of glacier melting typically experienced across the southern parts of Vatnajökull (Ahlmann and Thorarinsson, 1939). Working on Hoffellsjökull, which shares a common ice-divide with neighbouring Lambatungnajökull, Ahlmann and Thorarinsson (1939) measured ablation at various elevations over a 36-month period (Fig. 4.9). The resulting figures were high as a direct consequence of elevated air temperatures during the unusually warm summers of 1936, 1937 and 1938. In 1936 ablation on the glacier at 200 m above sea level was equal to 800 cm of water, in 1937 this figure was 750 cm and in 1938, 740 cm. On average, 85% of the ablation occurred between May 1st and October 1st, with some melting occurring, albeit minimal, in every month. These measurements translate to daily ablation figures of approximately 30 mm (water) per day in May and September and 55 mm in July (1936-38). Further observations made by Ahlmann and Thorarinsson (1939) regarding ablation on Hoffellsjökull are:

- The decrease in ablation with altitude is not linear – ablation rates decrease with increasing altitude (being twice the rate at the margin (100 m) compared with 500 m, and four times the rate of that experienced at 1000 m).
- The amount of ablation per unit area on different parts of the glacier is correlated with air temperature.
- The relationship between ablation and air temperature is not linear – ablation rates increase with rising temperature.
- Ablation is more dependent on meteorological factors (eg. air temperature, wind speed, precipitation, etc.) than on radiation, this is particularly true in late summer and autumn.

No detailed glaciological investigations had been carried out on any other part of Vatnajökull until recently. Since 1991, mass balance has been regularly monitored on several outlets of the ice-cap (Björnsson *et al.*, 1992, 1998; Oerlemans *et al.*, 1999). The most recent attempt to study the meteorology and mass balance conditions on

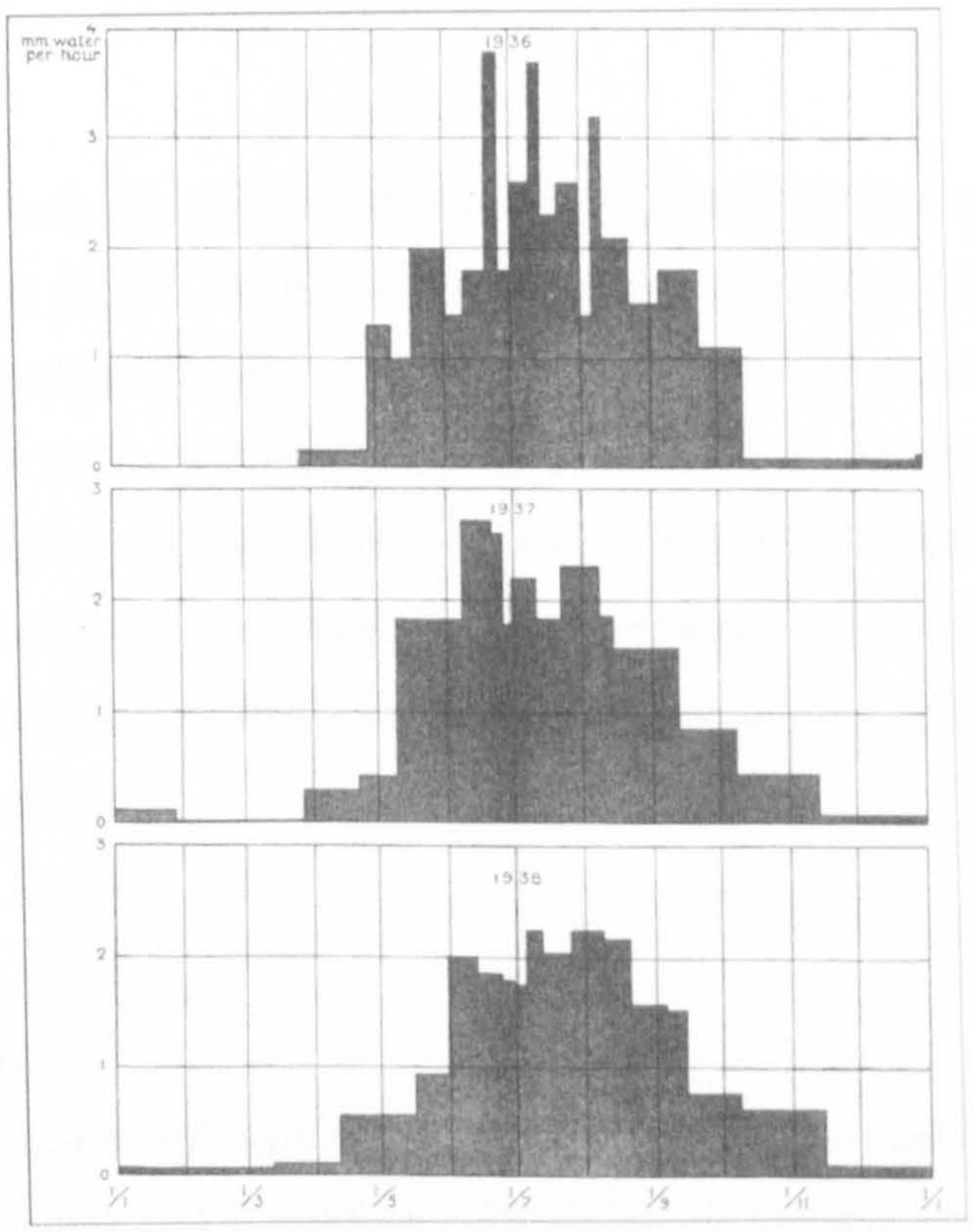
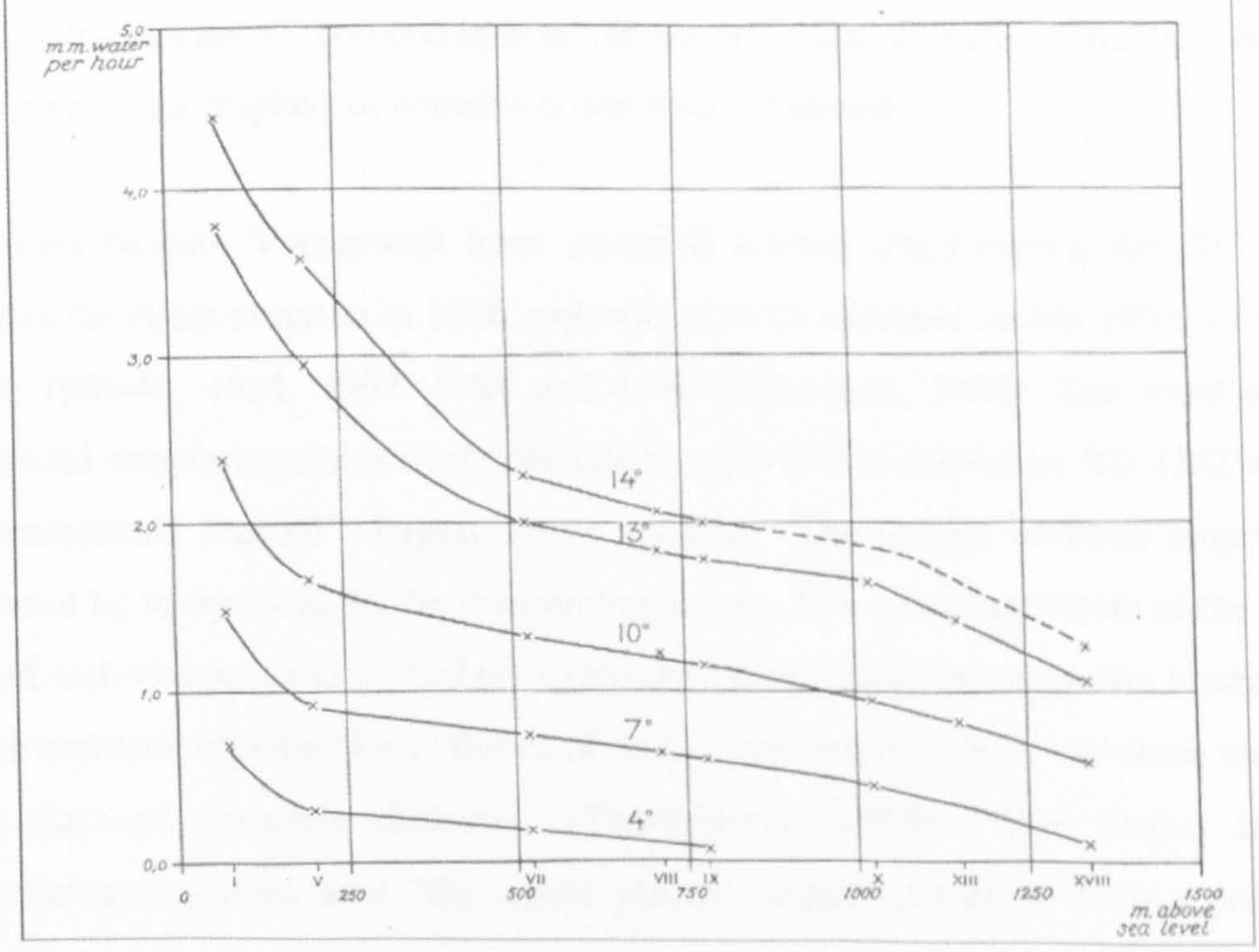


Figure 4.9 Ablation on Hoffellsjökull (from Ahlmann and Thorarinsson, 1942)
(above) The mean ablation at Stake V on Hoffellsjökull, 200 m asl, during 1936, 1937 and 1938.



(above) The relation between ablation, air temperature and altitude. Ablation is measured in mm of water per hour. Temperatures are measured at Hólar.

Vatnajökull has been set up by workers from Iceland, Holland and Austria (Oerlemans, 1998; Oerlemans *et al.*, 1999). Several years of detailed measurements from the southeast and northern outlets of Vatnajökull will provide a framework to study and eventually model the mass balance regime of the ice-cap. Although Oerlemans *et al.* are currently conducting measurements on the broad lobate outlets of Brúárjökull and Breiðamerkurjökull, no mass balance information exists for the steep, valley-terminating glaciers of northeastern Vatnajökull which are likely to be more responsive to short-term climatic perturbations.

Volcanic setting

Presently, three active volcanic centres lie beneath the Vatnajökull ice-cap: Barðabunga, Grimsvötn-Haábunga and Kverkfjöll. These are all located within the Central Volcanic Zone and hence situated under the western half of Vatnajökull. Occasionally, explosive, phreatic eruptions free large amounts of water that flow beneath the ice, before bursting out of the glacier margin. These catastrophic floods, termed *jökulhlaups*, can exceed discharges of 45,000 m^3/s (Russell *et al.*, 1999) and cause great destruction to the surrounding area. The eruption of Barðabunga in September 1996, and the subsequent catastrophic release of 600,000,000 m^3 of water in just 24 hours, remains one of the most spectacular displays of nature's power ever witnessed.

Eruptions beneath Vatnajökull have occurred several times during the 20th century. Besides the *Gjalp* eruption in 1996, volcanic activity occurred in AD 1903, 1922, 1934, 1938, 1945-46, 1954, 1965, 1984 and 1990 (Björnsson, 1988). The most explosive subglacial eruptions in historical time (since ~AD 900) occurred in AD 1362 and 1727 at Öræfajökull, Iceland's largest active volcano. The second of these eruptions was observed by many living in the surrounding farms. Eye-witnesses wrote of the air being "filled with violent shocks", and of "explosions from within the mountain blasting lumps of ice and rock into the sky", "floods of water" covered the plains and there were "three long days of complete darkness" (Thorarinsson, 1958). One further interesting observation describes how "the whole glacier surged forward to such a point that it

obliterated the distant mountain [Lómagnúpur] from view" (Thorarinsson, 1958). This is striking, if somewhat anecdotal, evidence that glaciers can advance catastrophically as a direct result of subglacial volcanic activity.

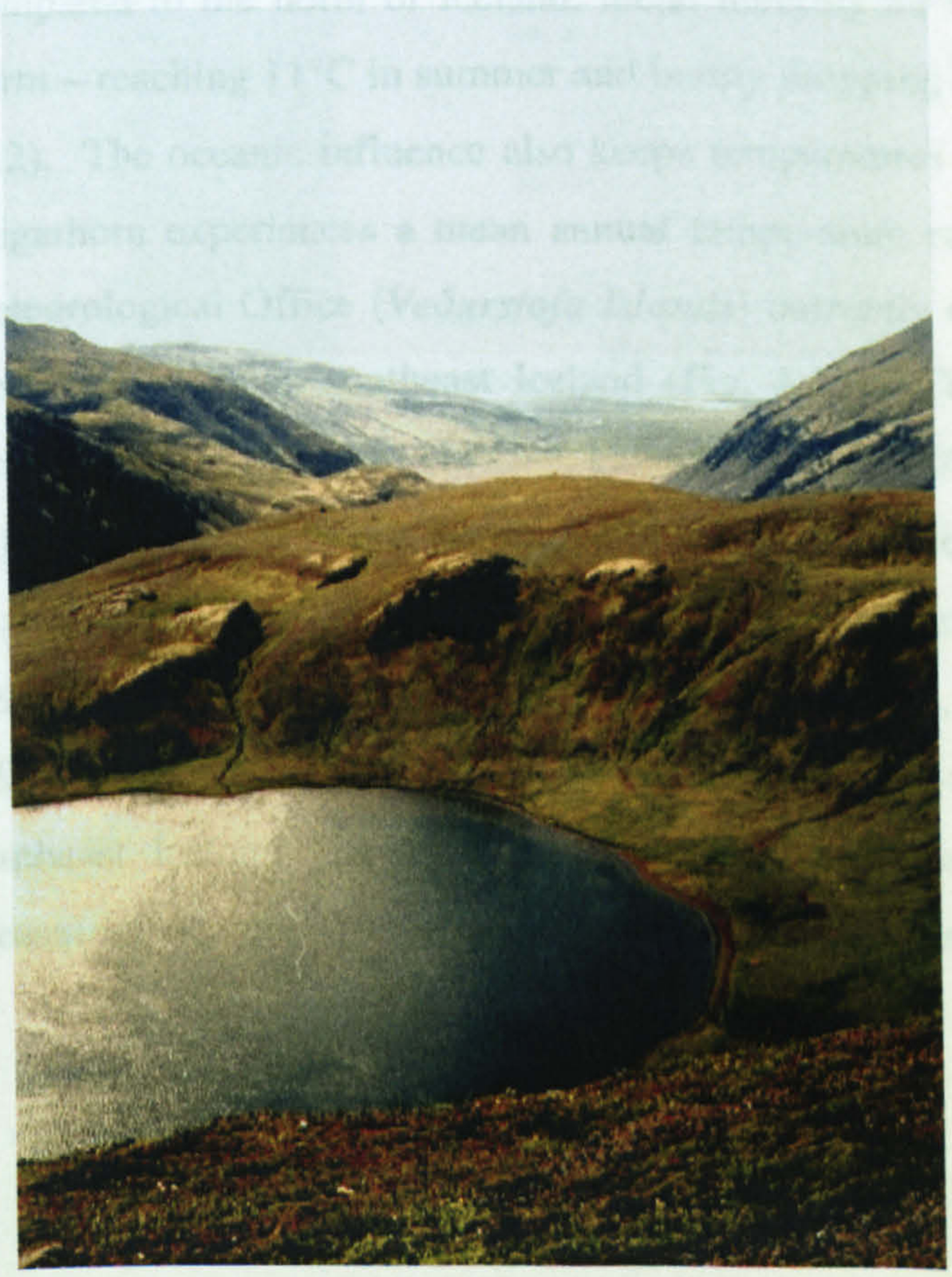
The eastern margin of Vatnajökull is far from Iceland's main geothermal foci and consequently experiences no such volcanic activity. There are no hot springs or high temperature areas within 50 km of Lambatungnajökull, the nearest being Öræfajökull to the southwest, Kverkfjöll to the west and Askjá to the north.

Topographic setting

The topography of the Hornafjörður-Lón area, the location for this research, is more alpine in appearance than much of southeast Iceland. In contrast to the bulk of the southern coast of Iceland, the relief is elevated, steep and often rugged. Apart from the large outwash plains of the Austurfljót and Jokulsá i Lon rivers, the study site consists of uplands lying between 300 and 1000 metres above sea level. The horn-shaped mountain of Grasgiljatindur immediately south of Lambatungnajökull reaches over 1200 m whilst in the east of the region the ice-capped summits around Hofsjökull í Lón attain 1300 m in height. The imposing coastal mountains of Vestrahorn and Eystrahorn, rising to 888 and 606 m respectively, fall steeply into the sea. For the most part, the coastal plain is less than 10 km wide at its broadest point. The mountainous terrain of the Hornafjörður district is dissected by several, deep, U-shaped valleys (Fig. 4.10). Only four of these valleys still host glaciers in their upper reaches, most notably Skyndidalur. Many other valleys are no longer glaciated, such as Hoffellsdalur, Laxárdalur and Endálausidalur in Hornafjörður, Gjaðadalur in Lón and Flugustaðadalur in Altafjörður.

Figure 4.10 Topography of the study area.

(right) The impressive U-shaped valley of Skyndidalur, looking west.
Lambatungnajökull is in the distance, 8 km away.



(left) Hoffellsdalur viewed from lake Hoffellsvatn. The North Atlantic Ocean is in the distance.

Climatic setting

The weather in southeast Iceland is highly variable. The area is situated in the path of North Atlantic cyclones and is constantly threatened by storms moving in off the ocean. The speed and intensity of these cyclones largely determines the weather experienced along the south coast of Iceland, particularly in autumn and winter. Consequently, the climate is maritime, windy and can be very wet. The skies are normally cloudy, the air is rarely still and showers occur, on average, every two days. In the summer sea-fogs are common on the coastal plains. At Teigarhorn the average annual precipitation over the period 1961-1990 was 1229 mm. Records show a small increase (10%) in rainfall from the late 19th to the mid-20th century (Fig. 4.11). At Hólar in Hornafjörður, close to the study site, annual precipitation is around 1500 mm. The wettest months are from September to January; the driest are May, June and July (Fig. 4.11). Snowfall is common in winter but rare between June and October.

Compared to the north of Iceland, mean monthly air temperatures in the southeast are warm – reaching 11°C in summer and barely dropping below freezing in the winter (Fig. 4.12). The oceanic influence also keeps temperatures relatively equable all year round. Teigarhorn experiences a mean annual temperature range of only 9°C. The Icelandic Meteorological Office (*Veðurstofa Islands*) currently records air temperature at several weather stations in southeast Iceland (Fig. 4.12). Those with the longest continuous records are at Teigarhorn (1873–), Fagurholsmyri (1903–) and Holar (1924–). There is a close agreement between the course of mean annual temperature over the last 100 years at all three stations (Fig. 4.13). Presently, the mean annual temperature at Holar is 4.5°C (1961-1990), whilst at Teigarhorn, 70 km to the north, the temperature is typically 3.7°C (1961-1990). Between 1873 and 1960 mean annual air temperature rose by +1.5°C in southeast Iceland. Since the 1960s this trend has been reversed, although recent measurements (post-1990) suggest that temperatures are rising again (Fig. 4.13).

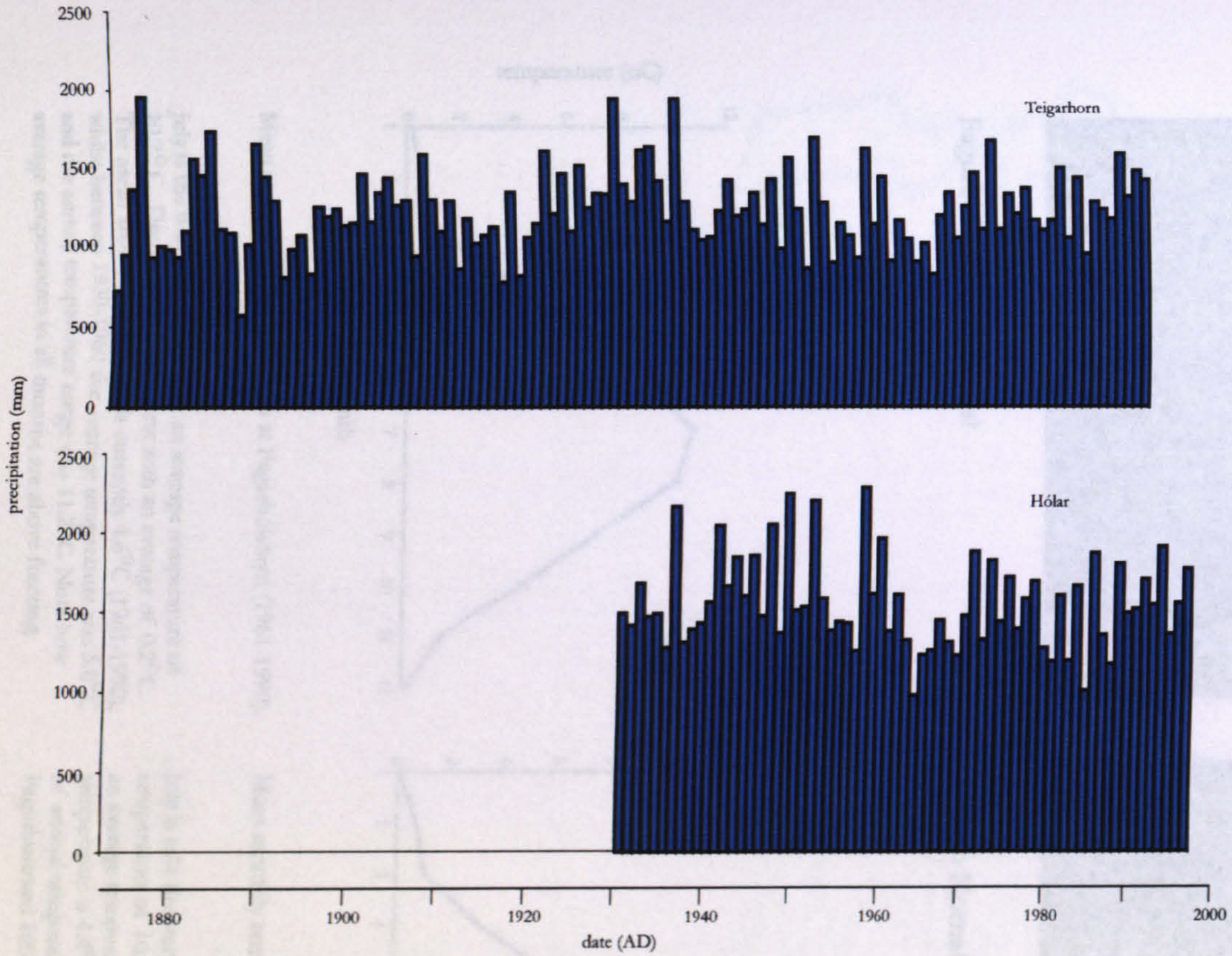


Figure 4.11 Precipitation in southeast Iceland over the last 125 years.

(left) Annual precipitation (in millimetres) at Teigarhorn since 1873.

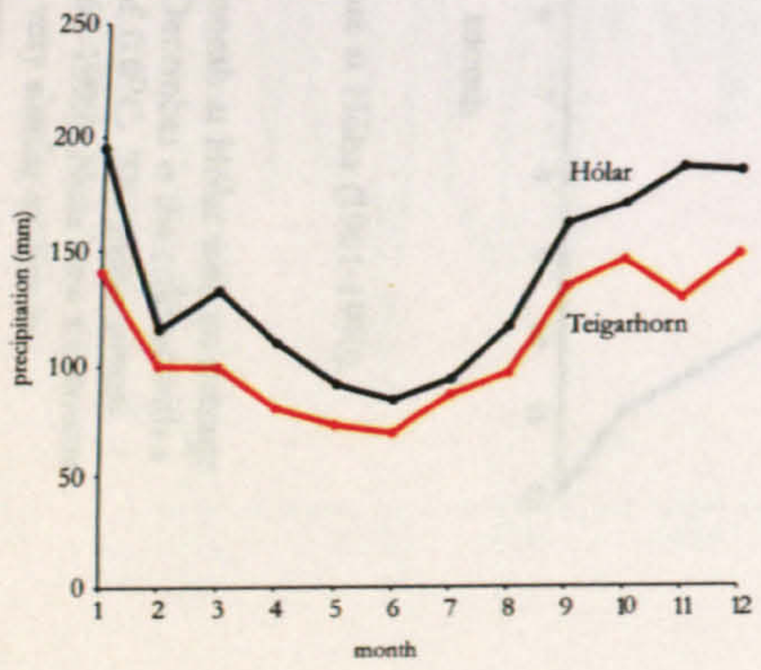
30-year averages were as follows:

1873-1900	1168 mm
1901-1930	1255 mm
1931-1960	1292 mm
1961-1990	1229 mm

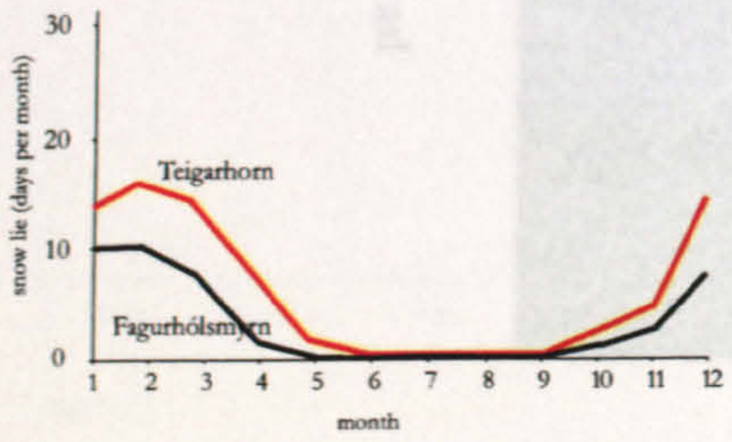
(left) Annual precipitation (mm) at Hólar i Hofn since 1931.

30-year averages were as follows:

1931-1960	1636 mm
1961-1990	1453 mm



(far left) Mean annual distribution of precipitation at two coastal stations in southeast Iceland (1931-1960).



(left) Average duration of snow lie in the vicinity of Fagurholsmyri (45 m asl) and Teigarhorn (22 m asl) between 1931-1960.

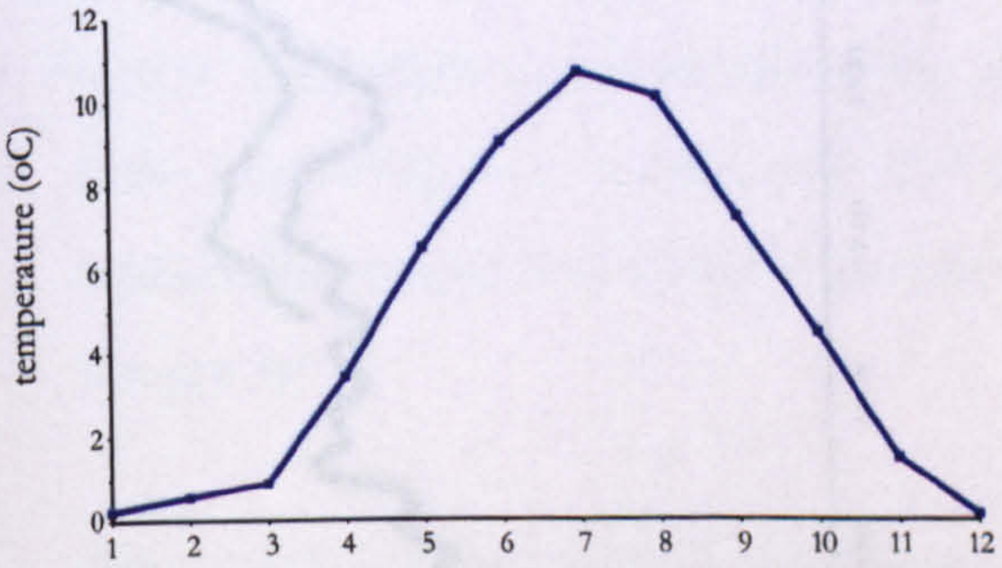
All data supplied by *Vedurstofa Islands*.

Figure 4.12 Air temperatures in southeast Iceland

The map shows the six weather stations in southeast Iceland and the year in which the first temperature measurements were made. All are located less than 50 m above sea level and within 25 km of the coast. Data supplied by *Veðurstofa Islands.*

Teigarhorn 1873-
Fagurhólsmyri 1903-
Hólar 1924-
Kirkjubæjarklaustur 1930-
Kambanes 1940-
Akurnes 1970-

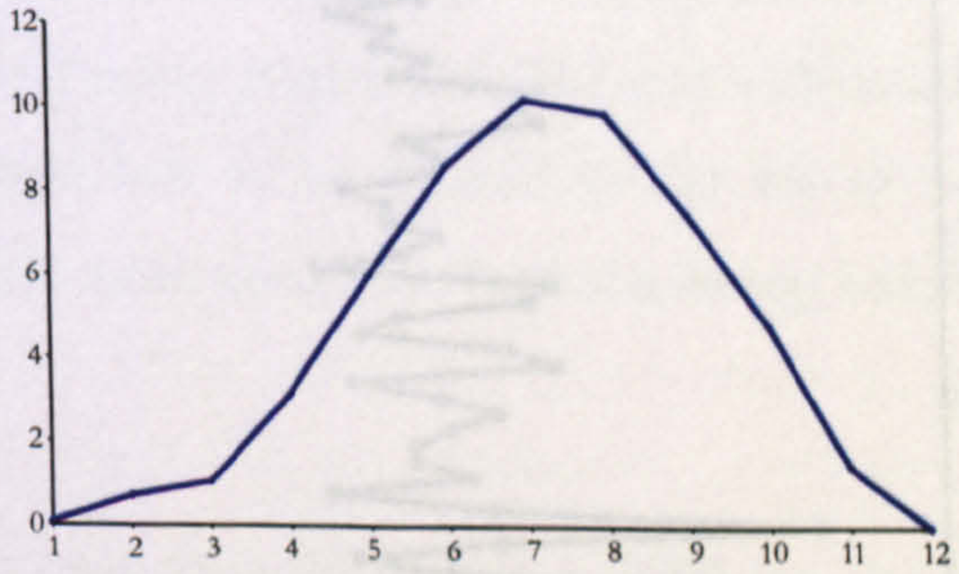
Fagurhólsmyri 45 m asl



Mean monthly air temperatures at Fagurhólsmyri (1961-1990).

July is the warmest month with an average temperature of 10.7° C. December is the coldest with an average of 0.2°C. The mean annual temperature is currently 4.6°C (1961-1990), whilst between 1930-1960 the average temperature was 5.0°C and the annual temperature range was 11.2°C. Note how average temperatures in all months are above freezing.

Hólar in Hornafjörður 22 m asl



Mean monthly temperature at Hólar (1961-1990).

July is aslo the warmest month at Hólar with an average temperature of 10.2°C. December is the coldest with a an average temperature of 0.0°C. The mean annual temperature is 4.4°C (1961-1990). Note how the course of annual temperature is very similar to that in Fagurholsmyri 100 km away.

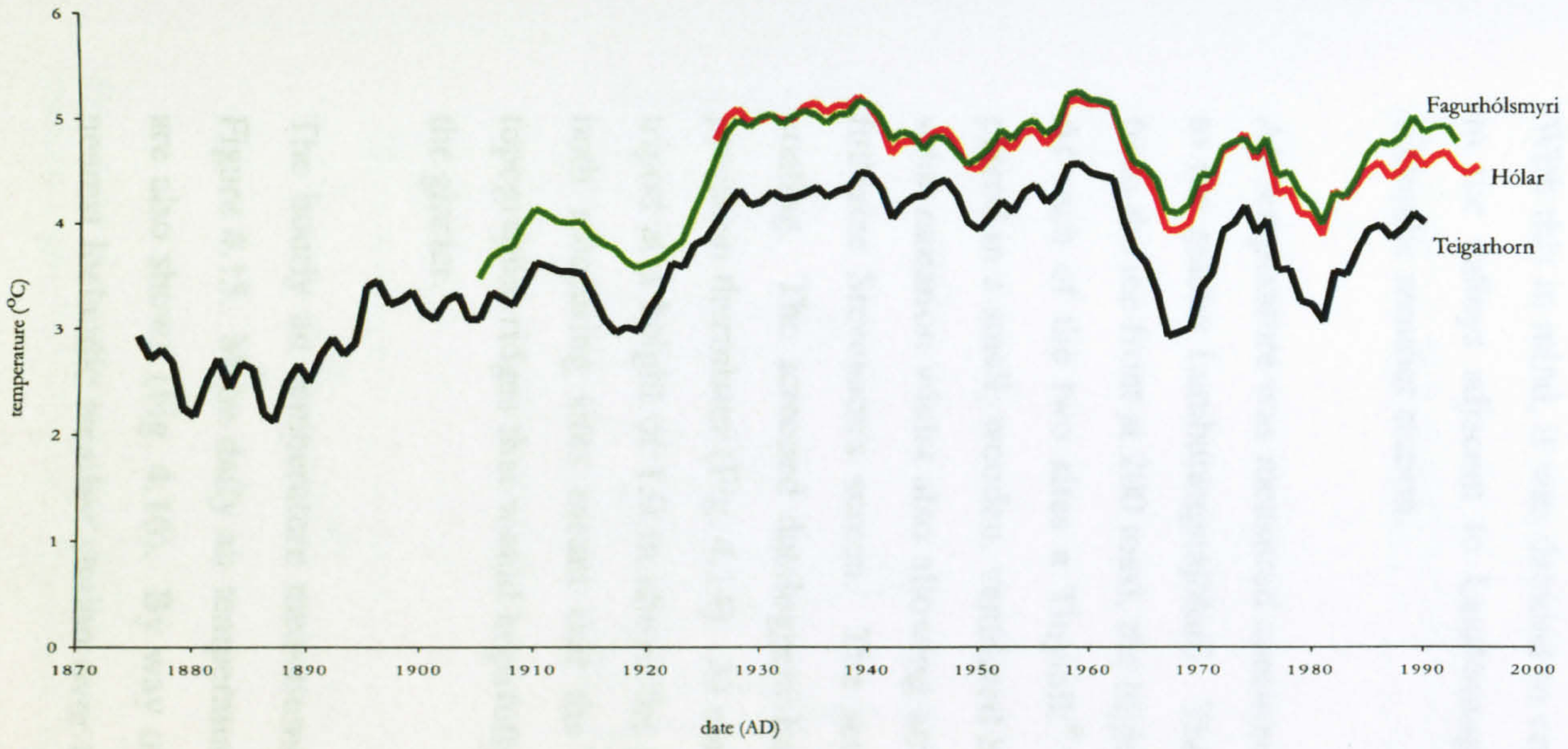


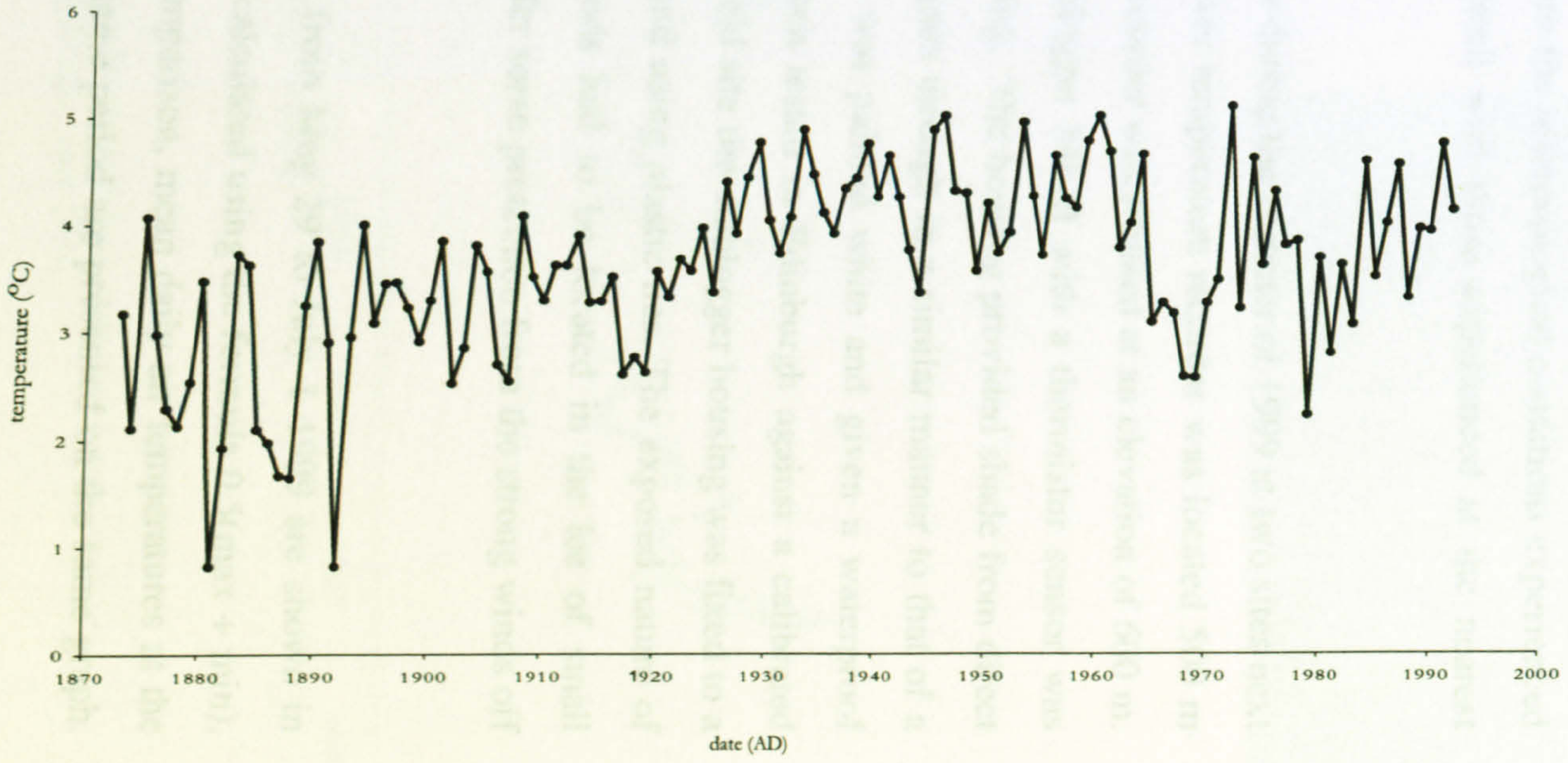
Figure 4.13 Air temperatures in southeast Iceland over the last 125 years.

(left) Five-year running means of annual temperature at three coastal stations in southeast Iceland:

Teigarhorn, since 1873
Fagurhólsmyri, since 1903
Hólar i Hofn, since 1924

Data supplied by *Vedurstofa Islands.*

Note the close agreement between all three records. Mean annual temperatures at Teigarhorn are generally 0.6 - 0.9 °C lower than those at Hólar.



(left) Mean annual temperature at Teigarhorn, 40 km NE of the study area, since 1873.

This is the second longest temperature series from Iceland and, although over 400 km to the east, correlates well with data from Stykkisholmur.

30-year averages from Teigarhorn are:

1873-1900	2.77 °C
1901-1930	3.51 °C
1931-1960	4.30 °C
1961-1990	3.70 °C

The topography of the area is mountainous which makes the weather conditions locally variable. The Vatnajökull ice-cap exerts a considerable influence over the area and can affect meteorological conditions markedly. In the summer, cold winds sometimes blow from the glacier reducing temperatures in the surrounding valleys and producing a *föhn effect*. Whilst the farm of Stafafell, at sea level, 20 km from the margin of Vatnajökull, enjoys high temperatures and unbroken sunshine, the valleys on the southern side of the Hornafjörður massif can be enveloped in low cloud and exposed to strong gusts of glacier-chilled air. Any general statements about the climate of the Hornafjörður and Lón regions, therefore, have to be quantified before extrapolation to the study area. With this in mind, it was decided to compare the meteorological conditions experienced in the valleys adjacent to Lambatungnajökull with those experienced at the nearest Icelandic weather station.

Air temperature was measured continuously during the summer of 1999 at two sites next to the glacier Lambatungnajökull. The lower temperature recorder was located 500 m from the ice-front at 200 masl, the higher recorder was situated at an elevation of 600 m. At each of the two sites a Tinytalk® datalogger Mk II with a thermistor sensor was placed in a small, wooden, ventilated housing. The housing provided shade from direct solar radiation whilst also allowing air to pass through in a similar manner to that of a full size Stevenson's screen. The screen was painted white and given a waterproof coating. The screened dataloggers had been tested in Edinburgh against a calibrated precision thermistor (Fig. 4.14). At each field site the datalogger housing was fixed to a tripod at a height of 1.0 m above the ground using plastic ties. The exposed nature of both measuring sites meant that the tripods had to be located in the lee of small topographic ridges that would hopefully offer some protection from the strong winds off the glacier.

The hourly air temperature measurements from May 29 to July 1 1999 are shown in Figure 4.15. Mean daily air temperatures, calculated using the formula 0.5(max + min), are also shown (Fig. 4.16). By way of comparison, mean daily air temperatures at the nearest Icelandic weather station over the same period are presented on the same graph.

(left) Wooden housing designed to shield datalogger from direct solar radiation. Ruler is 35 cm in length.

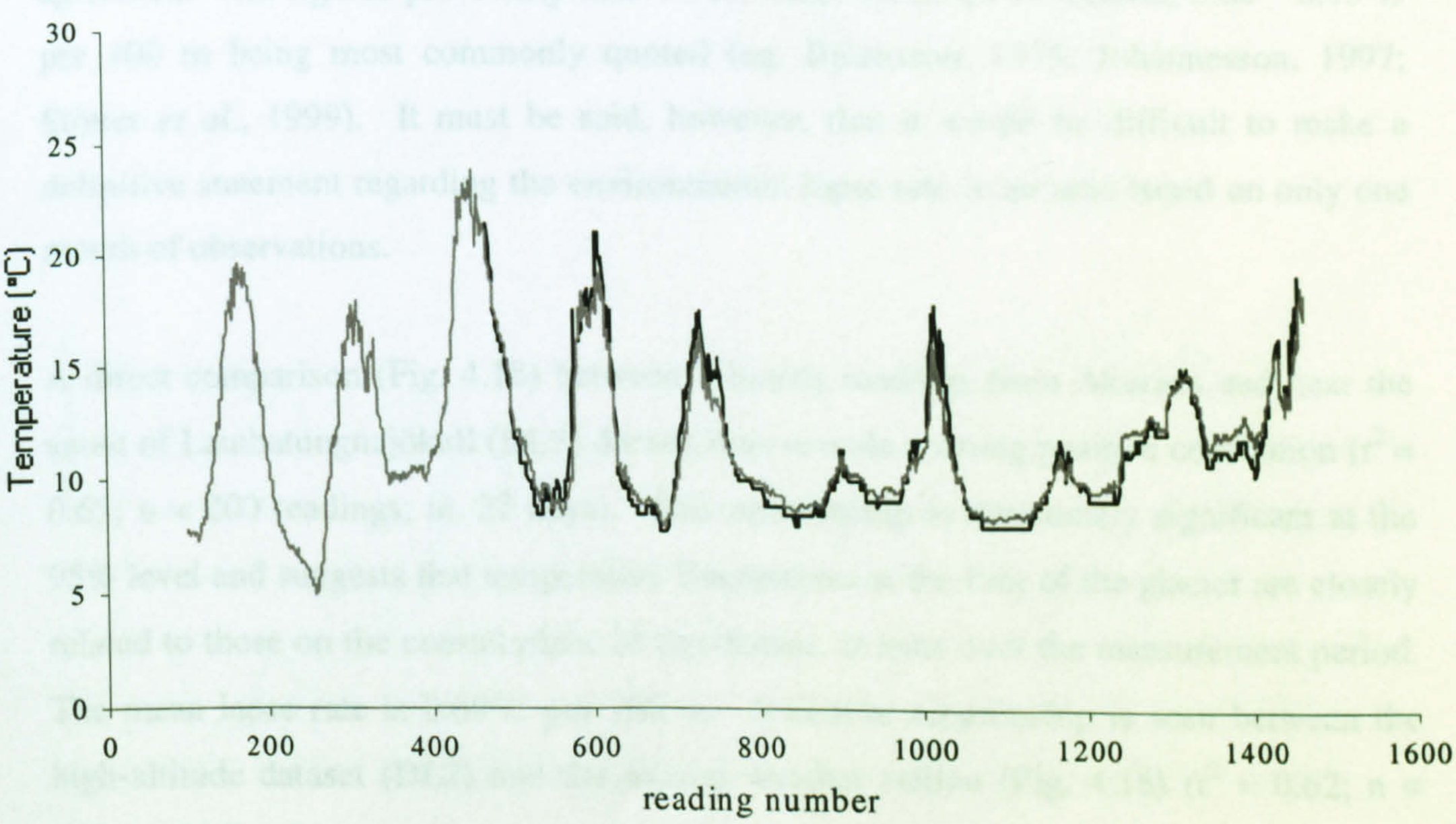


Figure 4.14 Comparison of Tinytalk datalogger and heat shield (black line) against screened, precision thermistor (grey line), Geography Department roof, Edinburgh, April 1999.

The main characteristics of air temperature variations measured at Lambatungnajökull are outlined in Table 4.4, below.

Table 4.4 Summary of air temperature data collected from Lambatungnajökull in 1999.

	DL2	*DL3*
Altitude (m)	200	600
Measurement period (days)	29	32
Maximum temperature (°C)	20.6	18.8
Minimum temperature (°C)	-1.4	-4.0
Extreme diurnal range (°C)	13.9 (21/6)	17.1 (29/6)
Mean daily temperature (°C)	7.6	5.1
Environmental lapse rate (°C/m)	0.0063	

There is a good correlation (r^2 = 0.78; n = 700 readings) between air temperature variations near the glacier snout (200 m asl) and at 600 m altitude (Fig. 4.17). The average temperature difference between the two sites, from May 29 to July 1 1999, was 2.50°C. This equates to a mean lapse rate of 0.63°C per 100 m altitude. This is in agreement with figures previously derived for other localities in Iceland; 0.60 - 0.75°C per 100 m being most commonly quoted (eg. Björnsson, 1975; Johannesson, 1997; Stötter *et al.*, 1999). It must be said, however, that it would be difficult to make a definitive statement regarding the environmental lapse rate in an area based on only one month of observations.

A direct comparison (Fig. 4.18) between 3-hourly readings from Akurnes and near the snout of Lambatungnajökull (DL3) during June reveals a strong positive correlation (r^2 = 0.65; n = 200 readings; ie. 27 days). This relationship is statistically significant at the 95% level and suggests that temperature fluctuations at the foot of the glacier are closely related to those on the coastal plain 20 km distant, at least over the measurement period. The mean lapse rate is 0.69°C per 100 m. A similar relationship is seen between the high-altitude dataset (DL2) and the coastal weather station (Fig. 4.18) (r^2 = 0.62; n = 240; ie. 30 days).

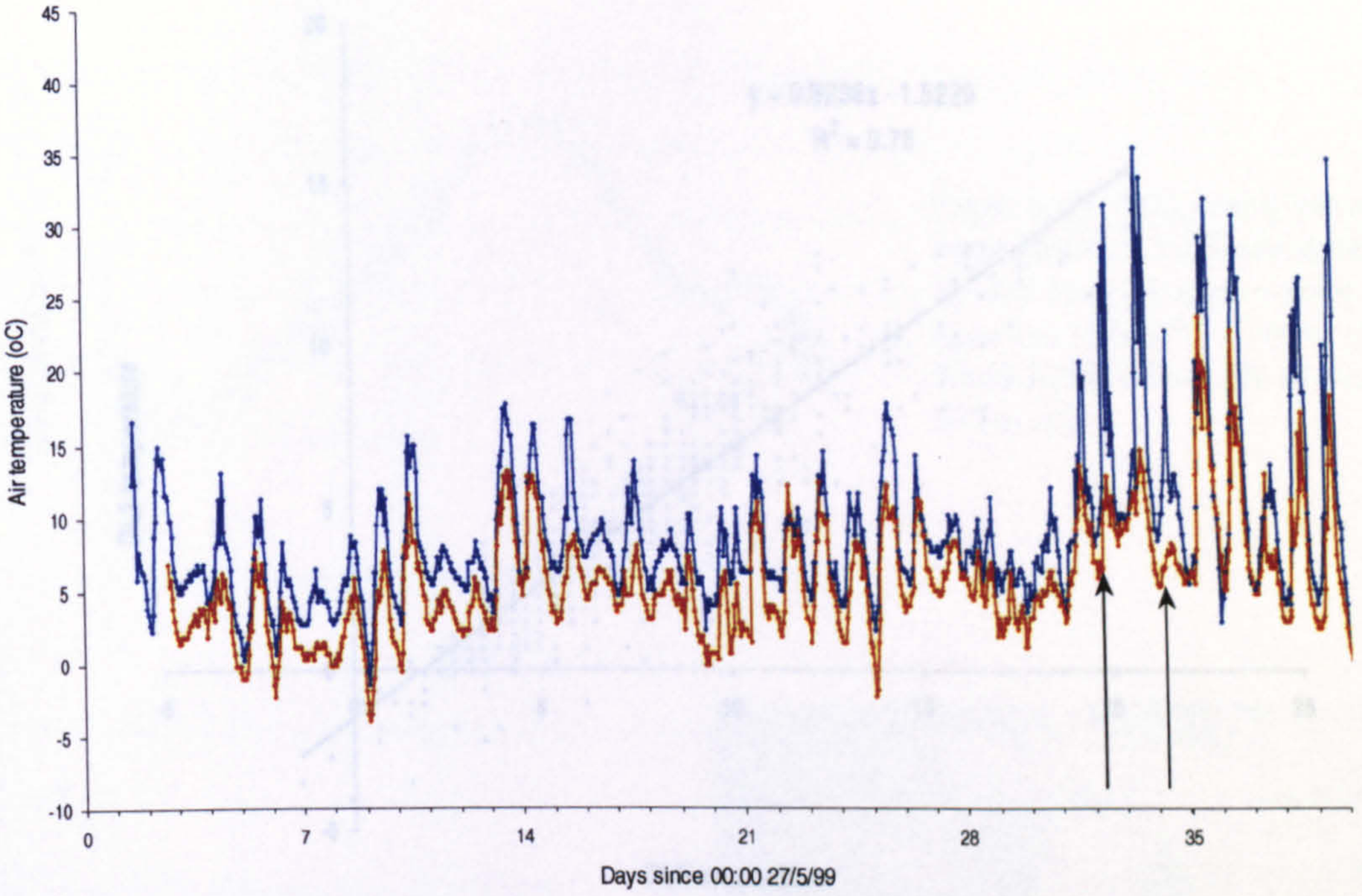


Figure 4.15 Hourly temperature data from DL2 (red) and DL3 (blue) spanning 40 days. The arrows indicate the time at which dataloggers were disturbed, probably by grazing sheep or deer.

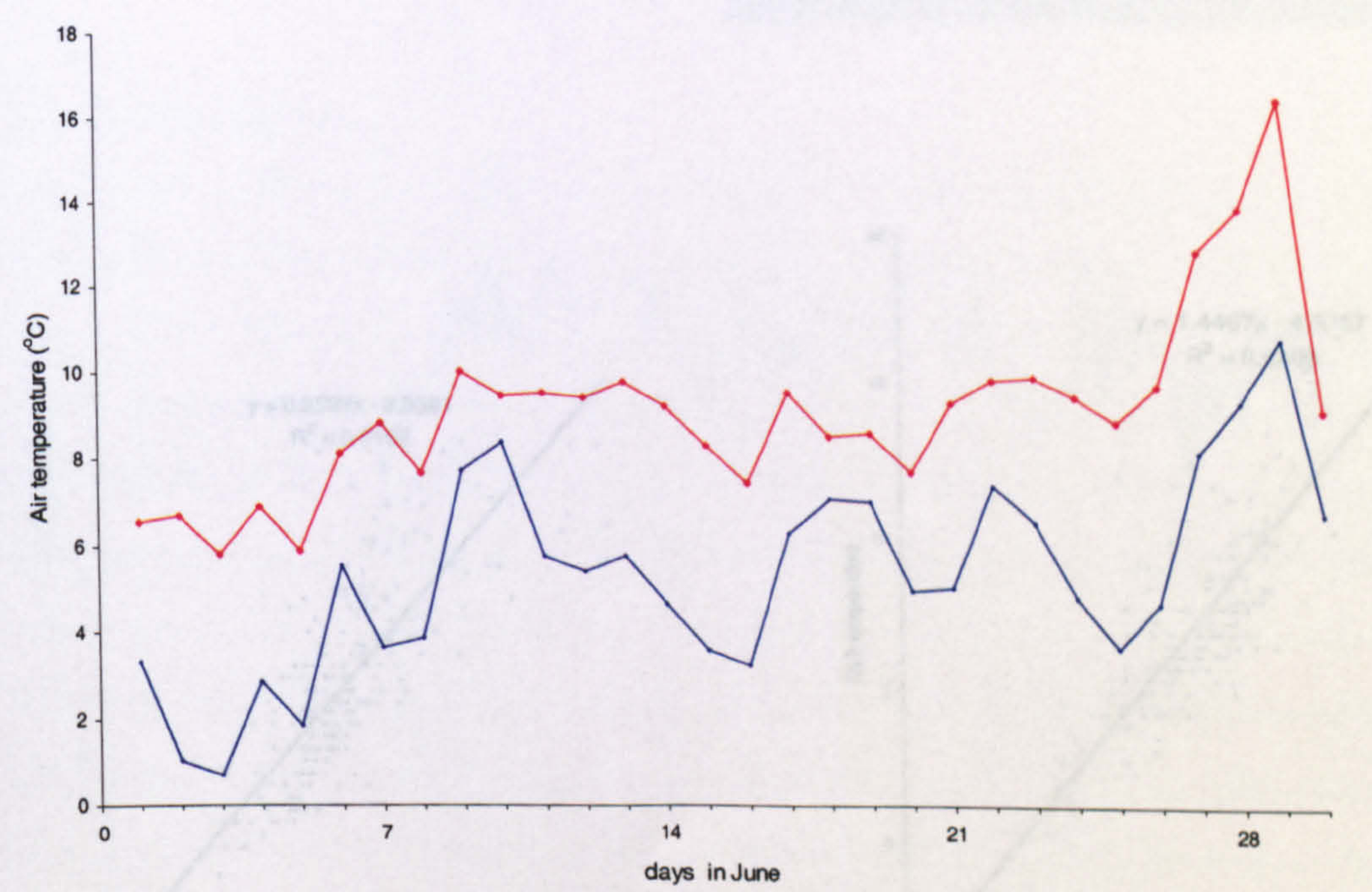


Figure 4.16 Mean daily air temperatures at DL2 (blue) and Akurnes (red), 01.06 - 31.06.99 (calculated using the formula: 0.5(max+min)). Data for Akurnes – the nearest Icelandic weather station – supplied by T. Einarsson (*Veðurstofa Islands*).

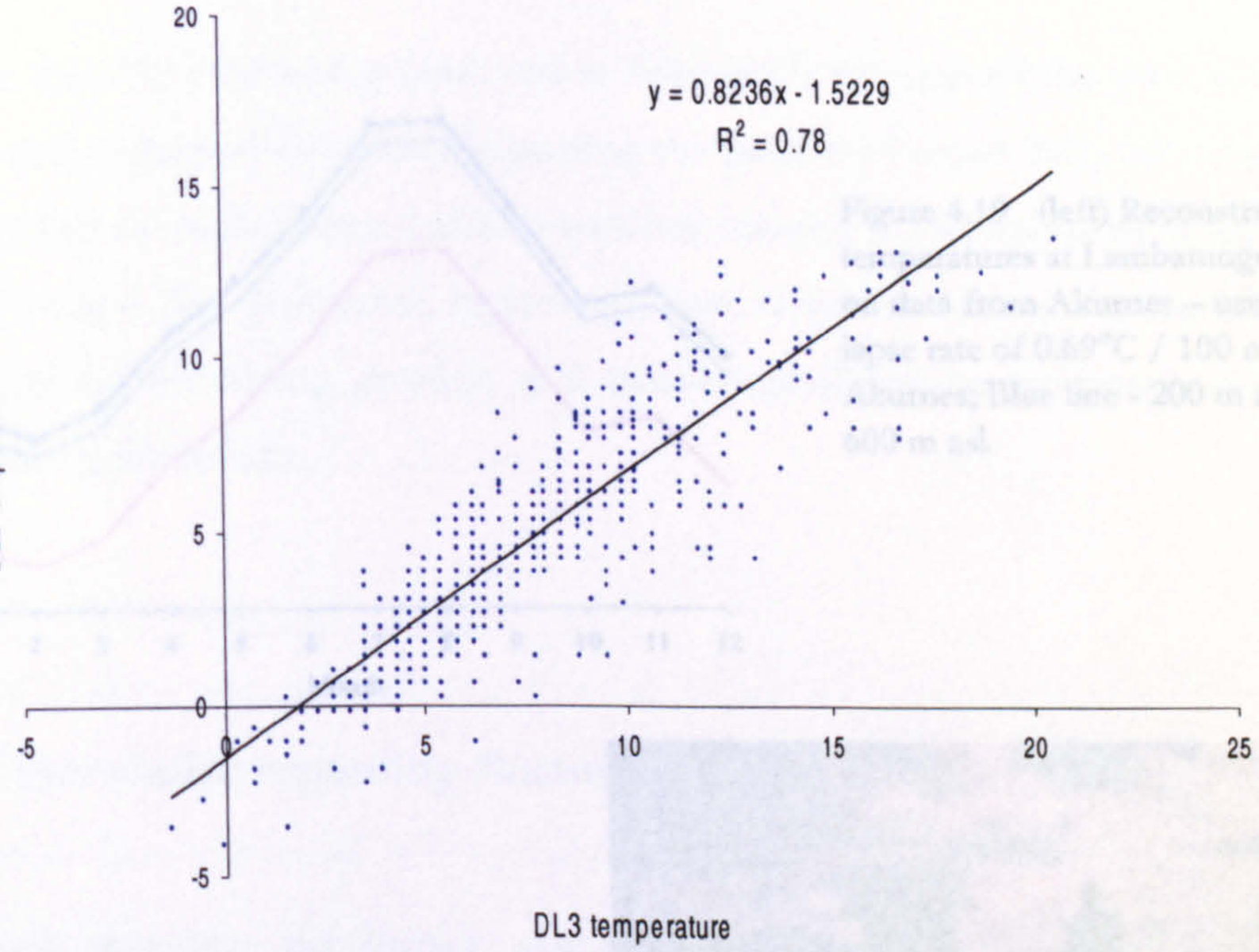


Figure 4.17 Correlation of hourly air temperatures at DL2 (located at 600 m asl) and DL3 (200 m) between 29.05.99 - 29.06.99.
Note the linear relationship and high r-squared value (0.78).

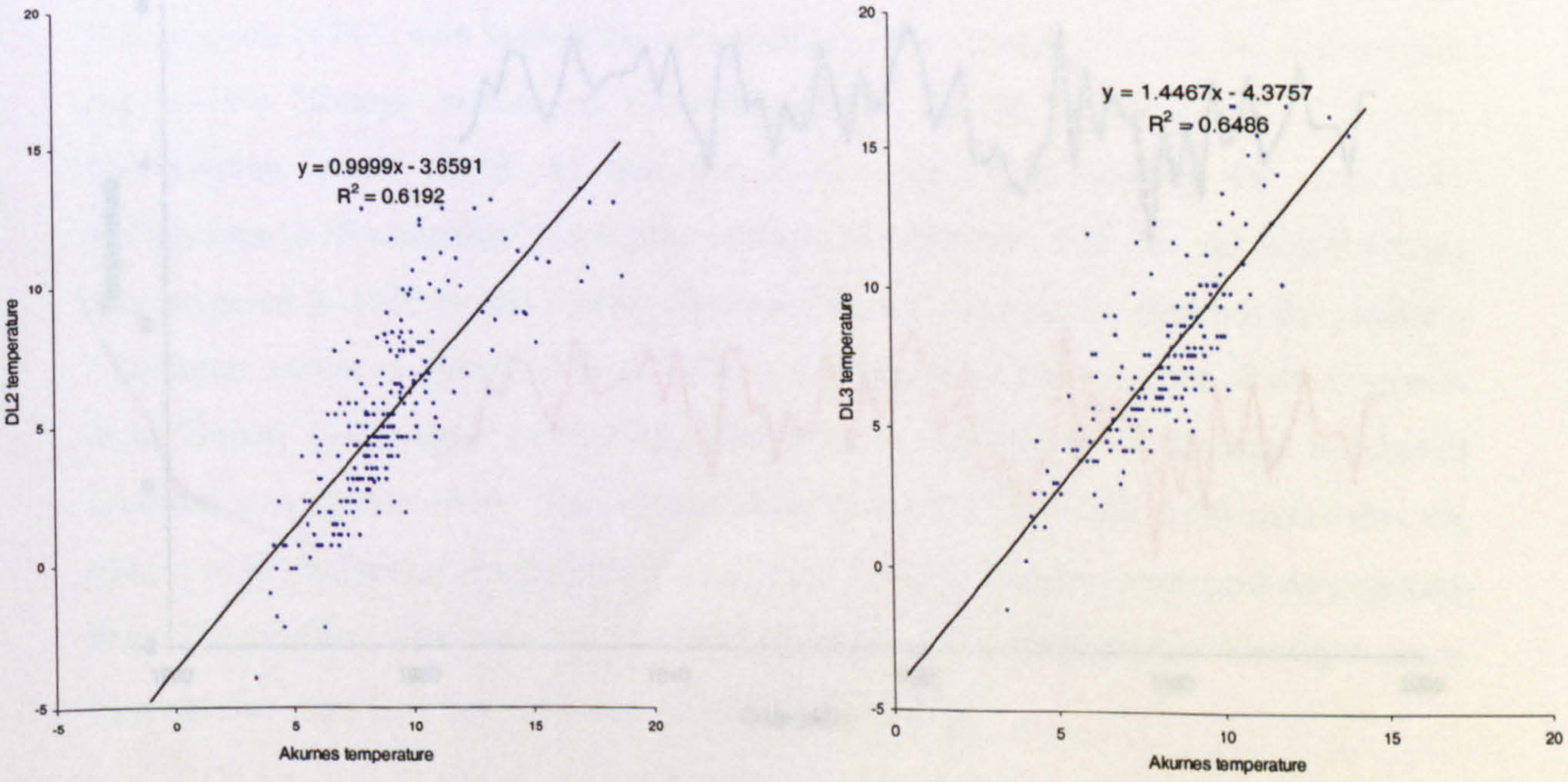


Figure 4.18 Correlation of three-hourly air temperatures at DL2 (left) and DL3 (right) with Akurnes, 20 km away (29.05.99 - 01.07.99).

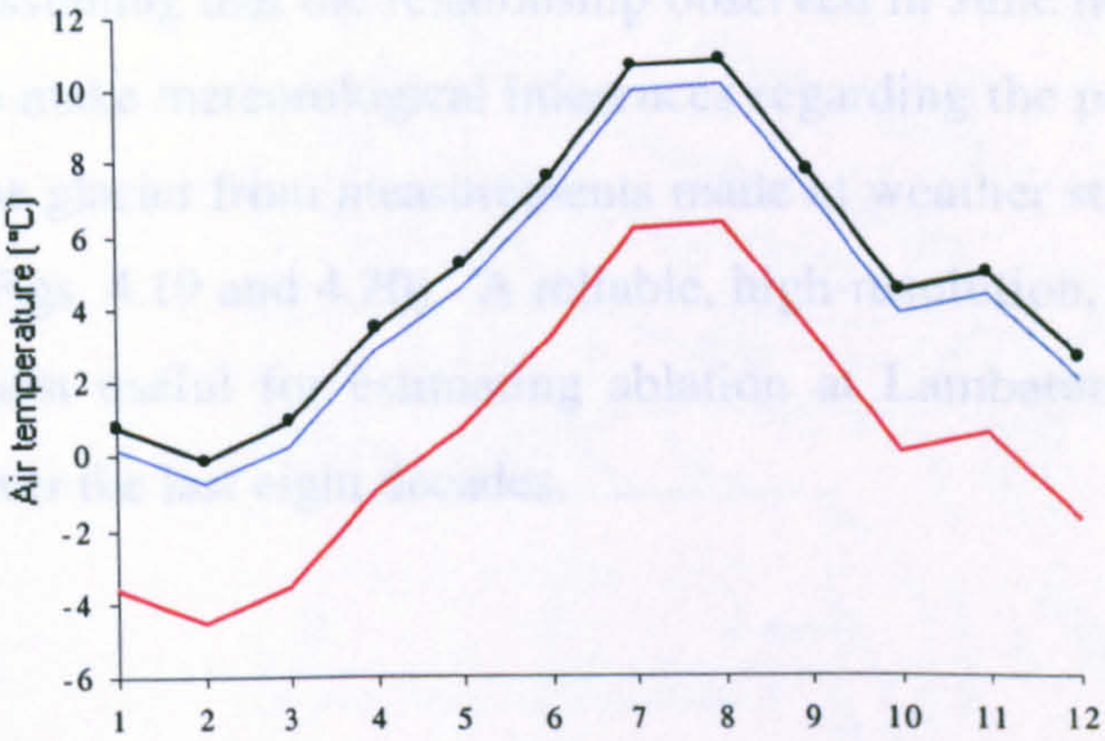


Figure 4.19 (left) Reconstructed 1997 air temperatures at Lambatungnajökull, based on data from Akurnes – using a constant lapse rate of 0.69°C / 100 m.; Black line - Akurnes; Blue line - 200 m asl; Red line - 600 m asl.

(right) Hólar weather station in Hornafjörður, 2 km E of Akurnes, has been recording air temperatures since 1924.

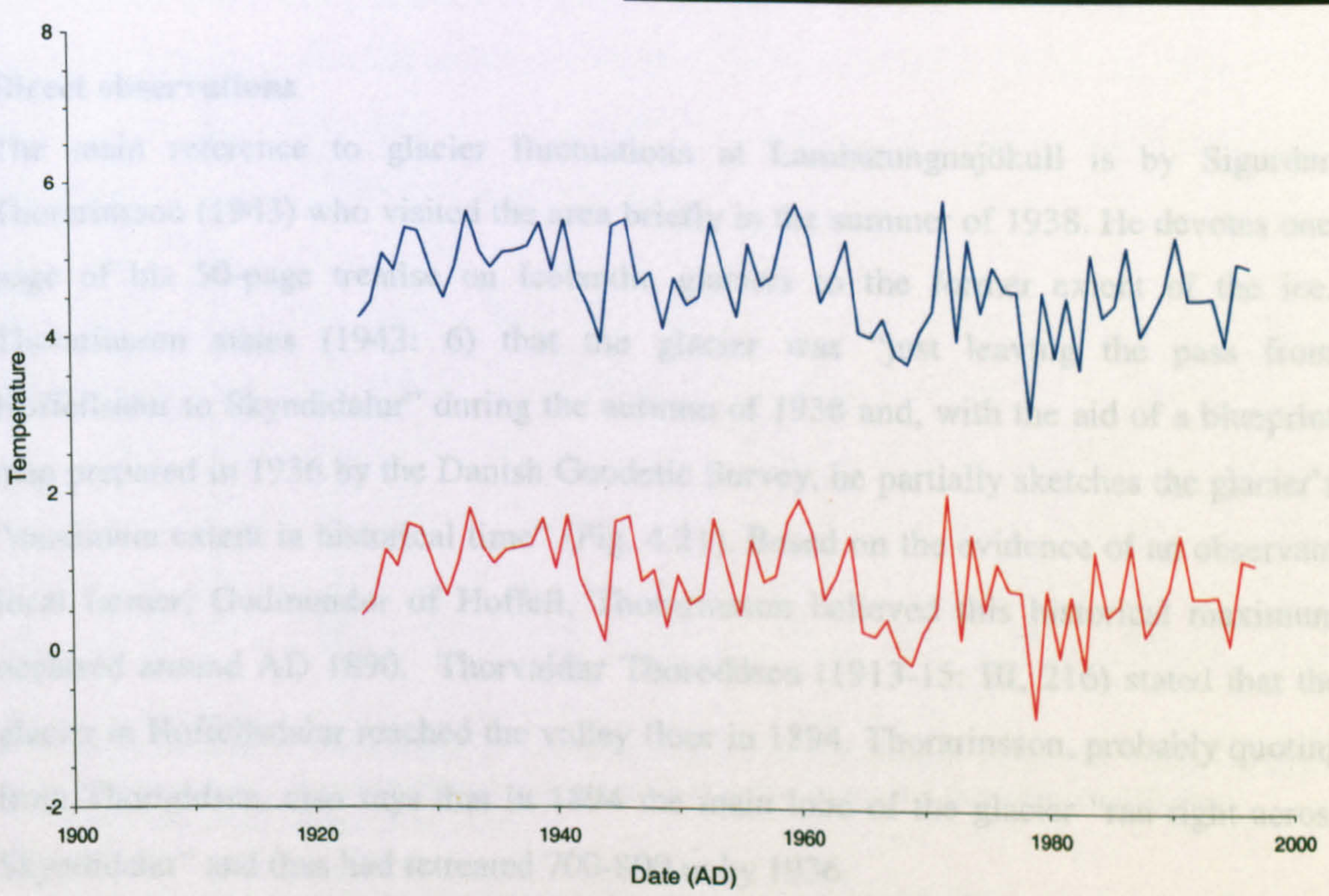


Figure 4.20 Mean annual air temperatures at Lambatungnajökull (600 m asl) since 1924 (red line). Reconstructed from Hólar temperature data (blue line). (Meteorological data supplied by *Veðurstofa Islands*.)

Assuming that the relationship observed in June holds throughout the year, it is possible to make meteorological inferences regarding the pattern of mean daily air temperature at the glacier from measurements made at weather stations in Hornafjörður since the 1920s (Figs. 4.19 and 4.20). A reliable, high-resolution, reconstruction of this nature would be most useful for estimating ablation at Lambatungnajökull by the degree-day method, over the last eight decades.

Present knowledge regarding fluctuations of Lambatungnajökull

Information regarding the former extent of the glacier Lambatungnajökull is very limited. No detailed published studies of the valleys of Hoffellsdalur or Skyndidalur or of the glaciers in this area exist in the geographical or geological literature. However, there are fragmentary pieces of documentary, cartographic and photographic evidence relating to ice-marginal fluctuations of Lambatungnajökull.

Direct observations

The main reference to glacier fluctuations at Lambatungnajökull is by Sigurdur Thorarinsson (1943) who visited the area briefly in the summer of 1938. He devotes one page of his 50-page treatise on Icelandic glaciers to the former extent of the ice. Thorarinsson states (1943: 6) that the glacier was "just leaving the pass from Hoffellsalur to Skyndidalur" during the autumn of 1938 and, with the aid of a blueprint map prepared in 1936 by the Danish Geodetic Survey, he partially sketches the glacier's "maximum extent in historical time" (Fig. 4.21). Based on the evidence of an observant local farmer, Gudmundur of Hoffell, Thorarinsson believed this historical maximum occurred around AD 1890. Thorvaldur Thoroddsen (1913-15: III, 216) stated that the glacier in Hoffellsdalur reached the valley floor in 1894. Thorarinsson, probably quoting from Thoroddsen, also says that in 1894 the main lobe of the glacier "ran right across Skyndidalur" and thus had retreated 700-800 m by 1936.

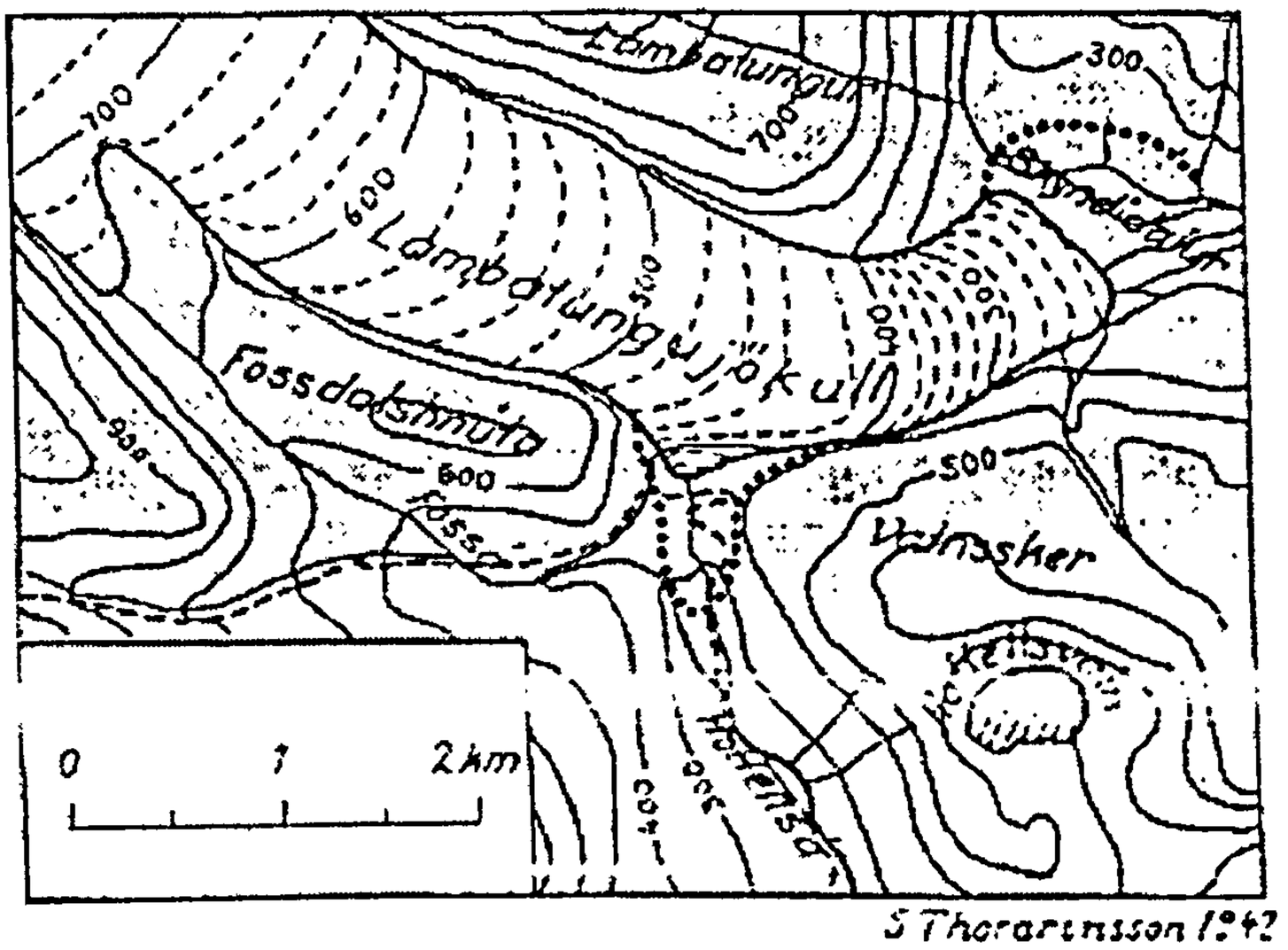
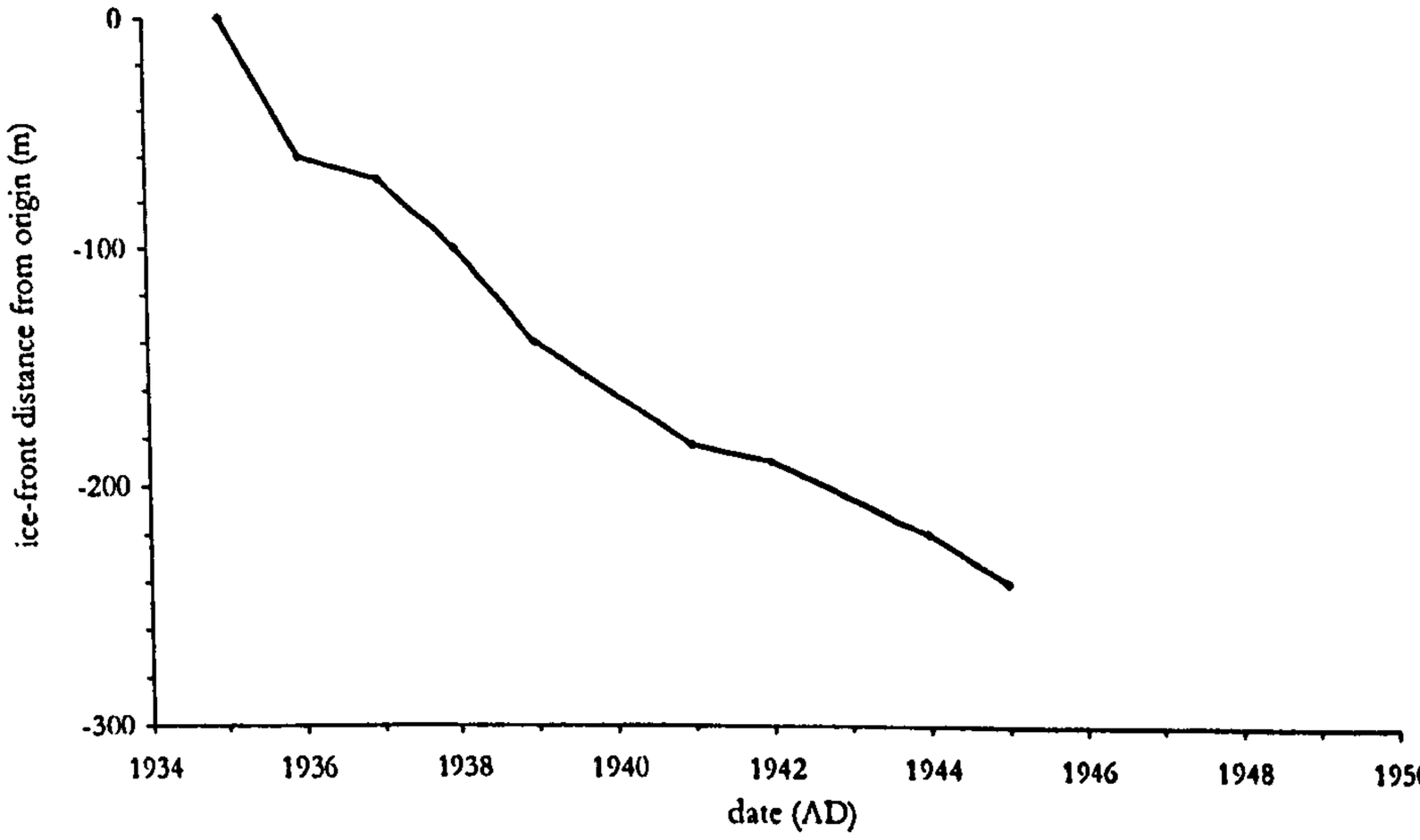


Figure 4.21 The former extent of Lambatungnajökull – direct observations.

Lambatungnajökull was first surveyed in 1936 by the Icelandic Geodetic Survey. The resulting map is reproduced above as it appears in Thorarinsson (1943). Onto this map Thorarinsson added the extent of the glacier in c. 1890 – according to Gudmundur of Hoffell (dotted line). This was supported by Thorodssen's observations in 1894 . Thorarinsson (1943: 6) thought these limits represented the glacier's "maximum in historical times" (ie. in the last ~1000 years) (see text for a more detailed discussion).



(above) Measured ice-front retreat of Lambatungnajökull in Hoffellsdalur 1935-1945. Observed by a local farmer over a 10-year period, the glacier is listed in the Icelandic Glaciological database as 'Hoffellsdalsjökull' (Sigurdsson, 1998).

During this time, glacier recession was most rapid in the late 1930s (30-50 m/yr) and slowest around 1941-42 (10 m/yr). By 1945 the glacier no longer extended into Hoffellsdalur (see Fig. 4.22) and monitoring ceased.

More recently, the retreat of the southern margin of Lambatungnajökull, which breached the pass at the head of Hoffellsdalur and was referred to as *Hoffellsdalsjökull*, was monitored for a few years between 1935 and 1945 (Fig. 4.21) (Sigurdsson, 1998).

Cartographic evidence

Maps of Iceland date back to the early 15th century. However, the scale and accuracy of these earliest maps makes them of little use for quantitative studies of glacier fluctuations (Ahlmann and Thorarinsson, 1937). The first detailed trigonometrical survey of Iceland was carried out between 1831-43 by Bjorn Gunnlaugsson. The resulting map is remarkably detailed and accurate. All five major ice-caps are shown to the correct scale, as are, interestingly, glaciers on Glamu, Hekla and Ok. The glaciers on the mountains of Ok and Hekla had all but disappeared by 1930 (Thorarinsson, 1943), whilst the plateau ice-cap 'Glamu-jökull' had melted by 1890, leaving only a few perennial snow patches surviving into the 20th century (Thorodssen, 1916-17; Thorarinsson, 1943). Vatnajökull, or *Klofajökull* as it was sometimes referred to, has several named outlet glaciers on Gunnlaugsson's map. Although the margin of Vatnajökull was not accurately surveyed in the 1830s, there is a clear suggestion that a glacier tongue existed in the upper part of Skyndidalur (64°25'N, 14°50'W). Referred to generally as 'Heinabergs Jökull' this could be the earliest cartographic representation of the glacier now known as *Lambatungnajökull.*

The first comprehensive survey of Iceland was carried out by the Danish General Staff and was conducted between 1902 and 1904. Maps were published at 1:50,000 scale of most of the coastal regions but unfortunately not of the more inaccessible regions such as Lambatungnajökull. The resulting large-scale map of Iceland, published in 1904, shows in some detail the size of various southern Vatnajökull outlets, from Skeiðarárjökull in the west to Hoffellsjökull in the east, but gives no indication of the size of Lambatungnajökull, Austurtungnajökull or Öxarfellsjökull.

Two further maps of Vatnajökull produced in 1919 (Wadell) and 1932 (Roberts) supply no additional information as to the exact location or size of Lambatungnajökull. This part of eastern Vatnajökull remained unsurveyed until 1936 when Steinthor Sigurdsson mapped the region on behalf of the Danish Geodetic Institute. The resulting topographic maps were published in 1938 at 1:100,000 scale [sheet 106, 107].

Subsequently, Lambatungnajökull has appeared on various maps, most notably the 1:50,000 US Army Series C762 [sheets 55 (Hoffell) and 56 (Lambatungujökull [sic])] compiled from aerial photographs taken in 1945-46. These maps were revised in 1988-90 using high-resolution space-borne images. Furthermore, *Landmaelingar Íslands* has produced revisions of the 1936 1:100,000 topographic maps [106, 107] in 1962, 1983 and 1998 (Fig. 4.1). Presently, no maps at a scale larger than 1:50,000 exist of Lambatungnajökull or the immediately surrounding area.

Photographic evidence

Aerial photographs of Iceland, taken at regular intervals since 1945, provide a valuable source of information regarding ice-front fluctuations during the last 55 years. Immediately after the Second World War, the US Army carried out a complete survey of Iceland. Photographs from these reconnaissance flights in the summer of 1945 and 1946 are kept at *Landmaelingar Íslands*. Subsequent aerial photography of Vatnajökull was carried out by the Icelandic Geodetic Survey in 1957, 1960-61, 1967, 1982, 1989-90 and 1998. Five aerial photographs of the terminus of Lambatungnajökull were taken between 1945 and 1999, three of which are shown in Figure 4.22.

Documentary / anecdotal evidence

Although the extent of Lambatungnajökull prior to the advent of aerial photography is uncertain, Thorarinsson (1943: 5-7) makes some general statements regarding the former size of the glacier based on a limited amount of indirect evidence. He believes that Lambatungnajökull was smaller in c. AD 1640 than in the early 1930s, enabling farmers and fishermen to make the journey from Hornafjörður to Fljótsdalur across the Vatnajökull ice-cap. Arni Magnusson's *Chorographica* dating from c. AD 1705 tells of

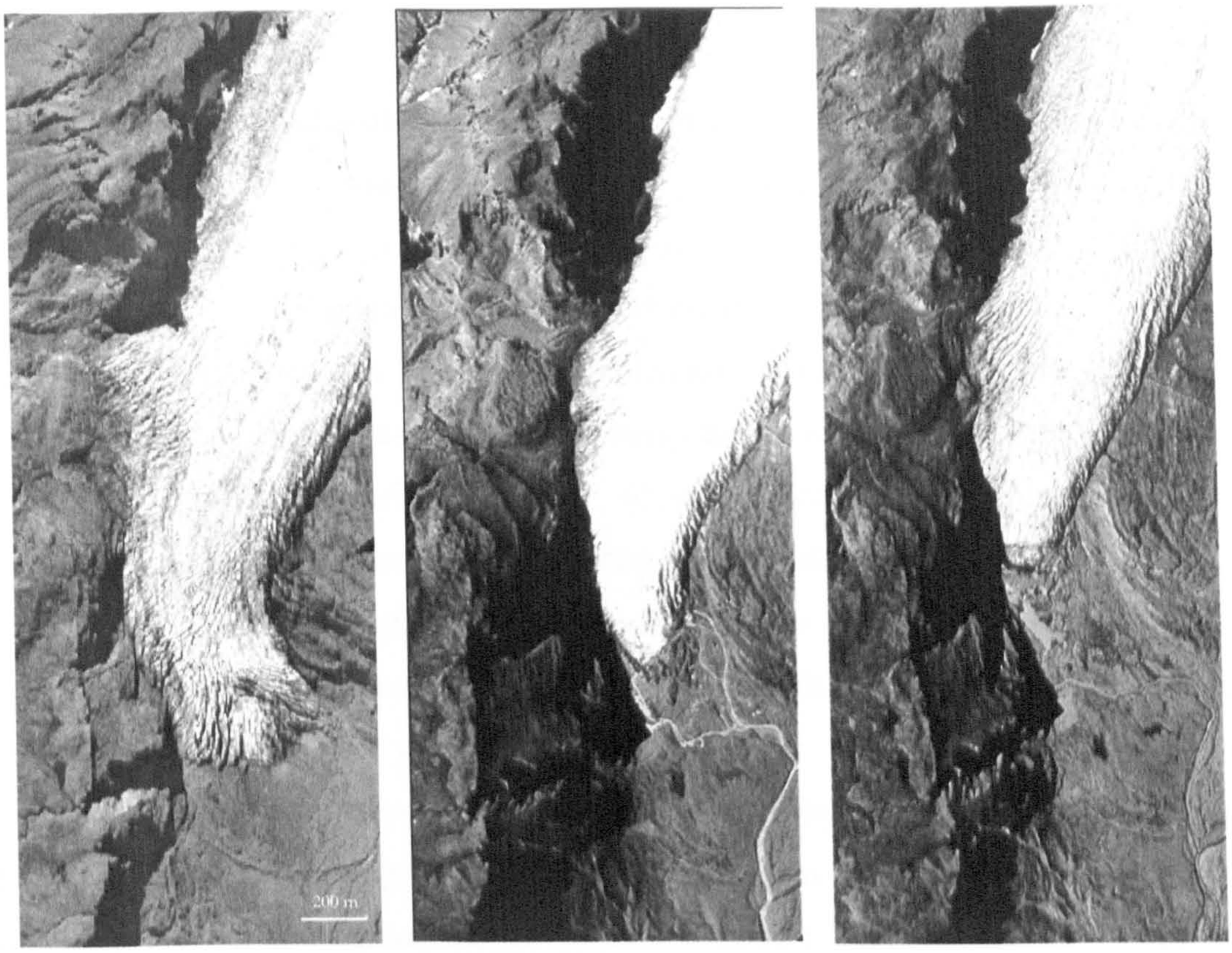


Figure 4.22 The former extent of Lambatungnajökull – photographic evidence.
Vertical aerial photographs taken in 1945, 1967 and 1982 reproduced at the same scale (above). North is to the right. Note the retreat of the southern margin (known as 'Hoffellsdalsjökull') into the main valley – Skyndidalur – between 1945 and 1967. Note also the highly crevassed ice-front in 1945 compared with 1967 and 1982 [Original photographs are held at *Landmaelingar Islands* .]

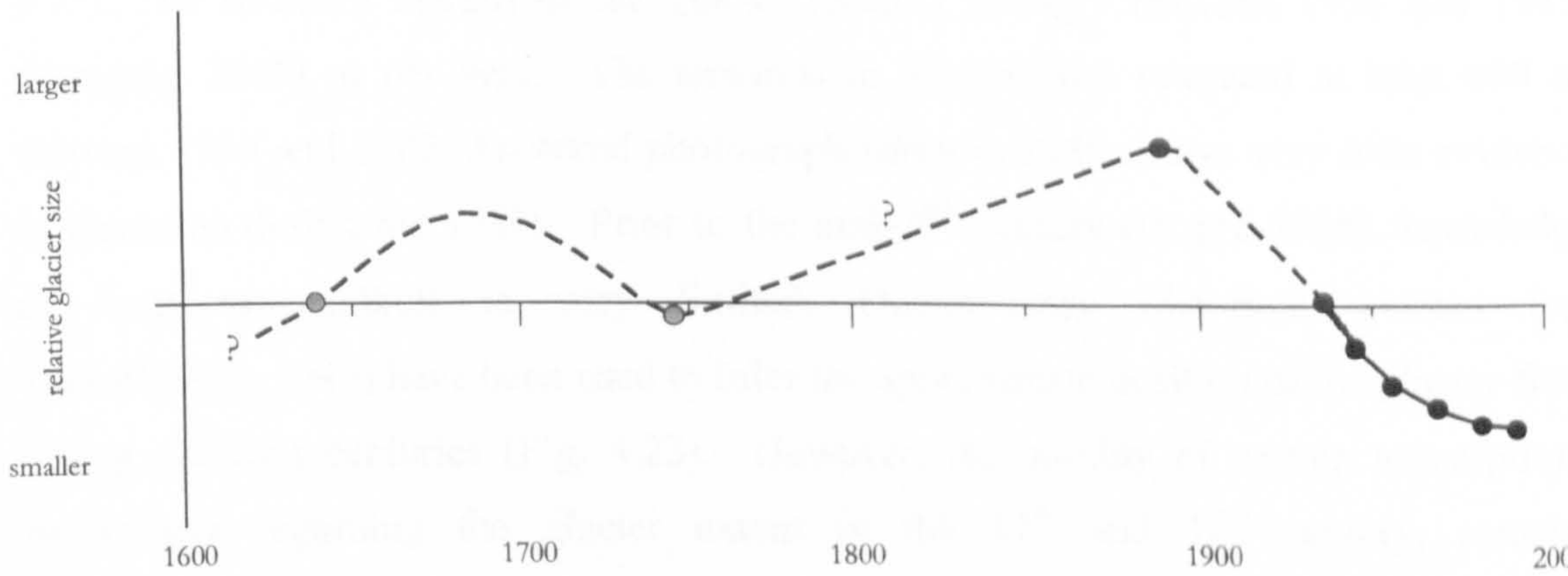


Figure 4.23 The size of Lambatungnajökull since c. AD 1600 based on the all the evidence prior to this research. Glacier size is expressed relative to its extent in AD 1936. The size of Lambatungnajökull in most of the 17th, 18th and 19th centuries is uncertain (dashed line). See text for a more detailed discussion.

Blue points - based on documentary sources (Thorarinsson, 1943)
Red points - based on direct observations (Thorodssen, 1913-15; Sigurdsson, 1998)
Black points - based on aerial photographs

a path that was only accessible when the margin of Lambatungnajökull did not occupy the col in Hoffellsdalur. Soon after c. 1650 the path was impassable "owing to glaciers" (Thorarinsson, 1943: 7). In 1936 this path became accessible once again, allowing farmers access into the valley of Fossdalur and the interior of Vatnajökull.

A further written account by Sigurdur Stefansson of Holt farm in 1746 states that the river flowing in Hoffellsdalur was a 'bergvatn' not a glacial meltwater stream ('jokulsa'). This led Thorarinsson (1943: 6) to propose that Lambatungnajökull was "less extensive in 1746 than in 1936", when the Hoffellsá river still transported some glacial meltwater. The validity of this statement can be questioned particularly as even during periods of more extensive glaciation the Hoffellsá river drains a predominantly non-glacial catchment. Any glacial meltwater streams emanating from Lambatungnajökull may have been sufficiently diluted by the influx of water from non-glacial sources to have appeared as a 'bergvatn' on arrival at Holt some 10 km downstream. The assumption that a "mountain stream with clear water" could not be associated with the presence of a glacier in Hoffellsdalur should be interpreted cautiously.

Summary

The former extent of Lambatungnajökull in various years over the last six decades has been deduced from existing observations, aerial photographs and published maps (Fig. 4.23). The southern margin of the glacier receded strongly between 1936 and 1945, averaging 20-30 m per year. The terminus in Skyndidalur retreated at least 600 m between 1945 and 1982. An aerial photograph taken in 1989 shows very little evidence of recession during the 1980s. Prior to the mid-20^{h} century (ie. pre-1936), knowledge of Lambatungnajökull is very limited. Documentary historical sources (eg. Thorarinsson, 1943) have been used to infer the approximate position of the glacier-front during previous centuries (Fig. 4.23). However, the validity of certain assumptions, particularly regarding the glacier extent in the 17th and 18th century, remains questionable.

Part II – Methods

The approach employed in this research involves a combination of geomorphological mapping and chronological techniques. The aim is to constrain firmly the spatial and temporal extent of glacier fluctuations in the study area.

Geomorphological mapping and geomorphological investigations

Geomophological mapping involves two stages, one based in the laboratory and one based in the field. Prior to any field investigations, large-scale geomorphological maps were compiled from the most recent stereographic aerial photographs of the study area. These maps were then taken into the field where they were supplemented, annotated and checked. 'Ground-truthing' allowed the interpretation of geomorphological features to be verified, whilst also allowing the further identification of ambiguous landforms and those features too small to be recognised from aerial photographs. The results of the geomorphological mapping investigations are presented in Chapter 6.

Chronology – Tephrochronology

The word *tephra* was first used by Sigurdur Thorarinsson (1944) as a collective term for all pyroclastic materials that travel through the air. Both ash-fall and ash-flow deposits are composed of tephra which accumulate as *tephra layers*. *Tephrochronology* is the technique of dating surfaces, landforms and structures using tephra layers as chronostratigraphic markers.

Tephrochronology is now widely used in Quaternary dating studies (Self and Sparks, 1981; Westgate *et al.*, 1985; Lowe, 1988; Osborn and Luckman, 1988; Haflidason *et al.*, 2000). Nowhere is the technique more established than in Iceland (eg. Thorarinsson, 1944, 1956, 1975, 1981). Numerous studies have used tephra layers to constrain various events chronologically, including glacier fluctuations and flow rates, palaeo-ecological studies (using pollen or plant macrofossils), farm abandonment, soil erosion and soil accumulation rates. There have been over 100 documented volcanic eruptions in Iceland

since the time of Nordic settlement. Typically there are two eruptions per decade. The high concentration of active volcanoes with distinct geo-chemical signatures allows the origin of late Holocene tephra layers to be traced with a high degree of certainty. Detailed documentary evidence, that records the history of Icelandic eruptions since c. AD 900, allows the precise calendar-age of many ash layers to be derived. Annals dating from the 13th and 14th centuries, for instance, describe eruptions of Hekla in 1104, Katla in 1245 and Öræfajökull in 1362. These texts have provided an invaluable source of information for scientists seeking to date features of 'historical age' – ie. the last ~1100 years. (eg. Thorarinsson, 1944, 1981; Larsen, 1978; Dugmore, 1989). A further link can be made between the volcanic ash in Iceland and those traces found in high-resolution Greenland ice-cores. Counting annual layers within the ice has revealed the timing of certain volcanic eruptions not recorded in the Icelandic annals (Palais *et al.*, 1991).

Once the tephra stratigraphy of an area has been determined, examination of the soil stratigraphy within the study area allows a relative chronology to be derived. Glacial deposits, such as moraines, can be age-bracketed by the tephra layers in the surrounding soil. Light-coloured tephras are most easily detectable in dark-coloured, humic soil sections. At least 12 of these acidic (rhyolitic) tephra layers are found in Holocene soils across Iceland. The presence or absence of a particular tephra layer of known age allows a maximum or minimum age estimate to be made.

In this research pits were excavated in the soil adjacent to the ice-proximal and ice-distal slopes of moraines. Soils were logged at sub-centimetre scale in the style used by Thorarinsson (1944, 1967), and also adopted by Dugmore (1987, 1989) and Gudmundsson (1998a). Each discrete tephra layer was described on the basis of depth from the surface, thickness, colour, grain size and contact definition. Other notes were made regarding the degree of weathering and/or induration, the lithic content of the tephra and any localised variations in texture or thickness. Samples were taken of key tephra layers for geochemical analysis. The results of the tephrastratigraphic investigations are shown in Chapter 7.

Chronology – Soil accumulation rates

In much of Iceland, soil accumulates as a result of aeolian sedimentation or due to mechanical weathering of the rock materials at the surface or in the subsoil. Consequently, soil formation is a slow process. The rate of soil accumulation varies from place to place and through time. Studies have shown that soil accumulation rates in Iceland between 9ka and 1ka BP varied little (Thorarinsson, 1960; Dugmore, 1987, 1989; Dugmore and Buckland, 1991). Prior to human settlement and the establishment of farming in c. AD 900, soils accumulated much more slowly and relatively constantly. Since c. AD 900 rates have varied greatly, particularly in the inhabited lowlands of southern Iceland. This contrast in soil accumulation rates is believed to relate to the introduction of grazing animals to the island by the Norse (Thorarinsson, 1981; Dugmore, 1987; Dugmore and Buckland, 1991). With knowledge of mean soil accumulation rates within a specific area it is possible to estimate the age of a soil-covered surface by extrapolation. This method has been widely used as a supplementary relative-dating method in Iceland. However, caution is advised where little is known regarding the environmental and anthropological history of the area in question.

Chronology – Lichenometric dating

Lichenometric dating – calculating the age of a surface from the size of lichens growing on it – has been utilised as a dating method in many aspects of earth science since pioneered by Roland Beschel in the 1950s. The combination of extremely slow growth in certain crustose species and the relative abundance of specimens in areas devoid of other vegetation has been exploited in polar and alpine areas. Traditionally, knowledge of species growth rates have be used to determine the age of the largest, or oldest, lichen growing on a surface and, hence, the age of the substrate itself. Perhaps the best example of such 'applied' lichenometry is the dating of moraine sequences in recently deglaciated terrain (eg. Beschel, 1950, 1961; Andrews and Webber, 1964; Denton and Karlén, 1973; Benedict, 1985). More recently, lichens have been used to date landslides, earthquake-induced rockfalls, periglacial features, raised beaches, archaeological remains and soil erosion rates all with varying degrees of success.

During the course of this research it became critically important to assess the strengths and limitations of various different lichenometric techniques. Hence, the whole of the next chapter is dedicated to the use of lichens in surface dating studies, whilst Chapter 8 presents the results of the lichenometric studies and Chapter 9 deals with the wider implications of these findings.

Chapter 5 : Lichen Growth and Lichen Dating – A Review

Introduction

Lichenometry has been utilised as a method of surface dating in many aspects of earth science over the last fifty years. The combination of extremely slow growth in certain lichen species and relative abundance in areas devoid of other vegetation has been exploited in dating surfaces populated by lichens. Knowledge of a species' growth rate can be used to determine the age of the lichens growing on a surface and, hence, the age of the substrate itself, assuming that lichen colonisation was almost instantaneous. Lichenometric-dating studies, such as those on moraine sequences in previously-glaciated areas, have been performed by various workers since the technique was pioneered by Beschel in 1950.

The wide range of applications of lichenometric dating is summarised in Table 5.1. Most studies have been conducted in high latitude or high altitude environments, the vast majority using yellow-green *Rhizocarpon* lichens. Since the technique was first performed, 50 years ago, numerous exciting and ingenious lichenometric applications have been examined:– dating man-made ancient monuments (Follmann, 1961; Winchester, 1984), calculating the rate of crustal uplift (Broadbent and Bergqvist, 1986), and determining the frequency and distribution of prehistoric earthquakes (Bull *et al.*, 1994), are just some examples. However, the absence of suitable calibration surfaces of known age in many uninhabited and remote parts of the world currently restricts the use of lichenometry both spatially and temporally.

Although studies utilising the growth rate of lichens are numerous (Table 5.1), few attempts have been made to analyse the actual growth mechanisms or study the growth history of the lichen thallus. Such a poor understanding has led to much controversy regarding the use of lichenometry, particularly from biologists. A detailed review of lichen physiology, with particular emphasis on *Rhizocarpon* species, will help in the analysis and interpretation of the results obtained in this research.

Table 5.1 Some worldwide applications of lichenometric dating.

Type of surface	*Location*	*Author(s)*
River channels (abandoned)	England	Milne (1982); Macklin (1986); Macklin *et al.* (1993)
River terraces	Iceland	Maizels and Dugmore (1985); Thompson and Jones (1986)
	Norway	Maizels and Petch (1985)
Lake shorelines	Baffin Island	Andrews and Barnett (1979)
	Norway	Matthews *et al.* (1986)
Raised beaches	Finnmark	Donner *et al.* (1977)
	Antarctica	Birkenmajer (1981)
	Spitsbergen	André (1985, 1986)
	Gulf of Bothnia	Broadbent and Bergqvist (1986)
Rock glaciers	Colorado	Birkeland (1973)
	Swiss Alps	Haeberli *et al.* (1979)
	Norway	Vere and Matthews (1985)
	Iceland	Martin *et al.* (1991)
Protalus ramparts	Norway	Shakesby *et al.* (1987)
Patterned ground	Scotland	King (1971)
	Norway	Ballantyne and Matthews (1982); Cook-Talbot (1991)
Talus	Spitsbergen	André (1985, 1986)
Debris flows	Swedish Lapland	Rapp and Nyberg (1981); Nyberg (1985)
	Poland	Jonasson (1991); Jonasson *et al.* (1991)
Landslides	Italian Alps	Porter and Orombelli (1981)
	Norway	Dawson *et al.* (1986)
Polynesian statues	Easter Island	Follmann (1961)
Stone walls	England	Laundon (1980)
	Colorado	Benedict (1985)
Stone circles	England	Winchester (1984, 1988)
Earthquake-induced faults/disruptions	Tadjikistan	Nikonov and Shebalina (1979); Smirnova and Nikonov (1990); Bull (1996)
Earthquake-induced Rockfalls	New Zealand	Bull *et al.* (1994); Bull and Brandon (1998);
Snow-free areas	Greenland	Pitman (1973); Koerner (1980); Benedict (1993)
Moraines	Southern Iceland	Jaksch (1970, 1975); Gordon and Sharp (1983); Thompson (1988); Bradwell (1998); Gudmundsson (1998a, 1998b); Evans *et al.* (1999); Kirkbride and Dugmore (2001b)
	Northern Iceland	Caseldine (1985, 1990, 1991); Hjört et al (1985) Haberle (1991); Kugelmann (1991); Stotter (1990)
	Southern Norway	Bornefeldt and Osterborg (1958), Andersen and Sollid (1971); Matthews (1974, 1975, 1977, 1983); Matthews *et al.* (1979); Haines-Young (1983); Erikstad and Sollid (1986); Innes (1986); Bickerton and Matthews (1992)
	Northern Norway	Denton and Karlén (1973); Ballantyne (1990)
	Jan Mayen	Anda *et al.* (1985)
	Spitsbergen	Werner (1993)
	East Greenland	Geirsdottir *et al.* (2000)
	West Greenland	Beschel (1958, 1961); Gribbon (1970) Weidick and Beschel (1973);
	Baffin Island	Andrews and Webber (1964);
	Alaska	Denton and Karlén (1973); Reger and Péwé (1969)
	Canada	Luckmann (1977); Smith *et al.* (1995)
	Chile	Winchester and Harrison (1984, 1994)
	Kenya	Spence and Mahaney (1988)
	New Zealand	Burrows and Lucas (1967); Orwin (1970); Burrows (1975); Winkler (2000)

Lichen biology

The lichen thallus, a symbiotic entity, consists of a small proportion of algal cells set within a greater number of fungal cells (Fig. 5.1). Lichens can be crustose, foliose or fruticose. This section of the review deals mainly with epilithic (surface), saxicolous (rock-dwelling), crustose and foliose varieties and the studies conducted on ascertaining their growth rates.

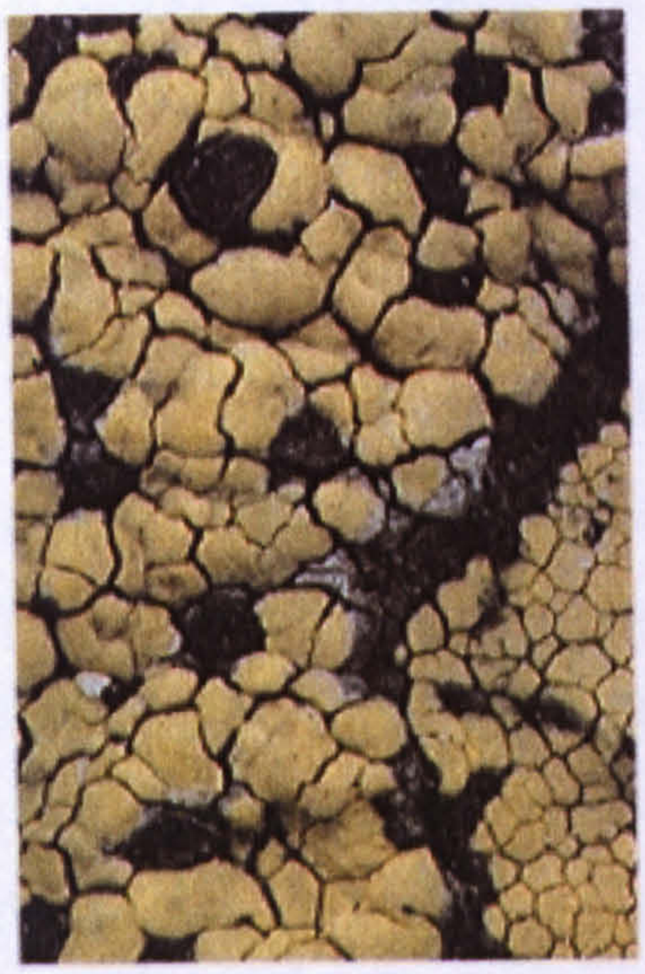
Crustose lichens are renowned for their slow growth yet surprisingly little is known about the mechanisms and physiological processes responsible for growth. Consisting of a symbiosis between algae and fungus, the lichen plant or *thallus* is thought to grow outwards once it has become established. In *Rhizocarpon* species, most often used in lichenometric dating studies, the yellow-green alga is *Trebouxia* and the dark brown-black disk on which the algae grow is made up of cemented fungal *hyphae*. The thallus margin is free of algal cells. This distinctive, black *hypothallus* is a key characteristic of *Rhizocarpon* lichens (Fig. 5.1).

Most investigations into lichen growth have been carried out on relatively rapid-growing foliose species (eg. Hale 1959, 1967; Phillips, 1963; Armstrong, 1974; Proctor, 1977; Aplin and Hill, 1979). Lichen growth is normally a marginal process resulting in enlargement of the plant by expansion of the peripheral zone. Hale (1967: 77) states that:

> "Lichens have a basic centrifugal pattern of growth, and unless the substrate or other plants interfere, the thallus will grow outward uniformly at the margins and form an orbicular colony... Beyond a certain point lateral transport of nutrients between the margins and the centre of the thallus is hampered and the older parts thicken somewhat."

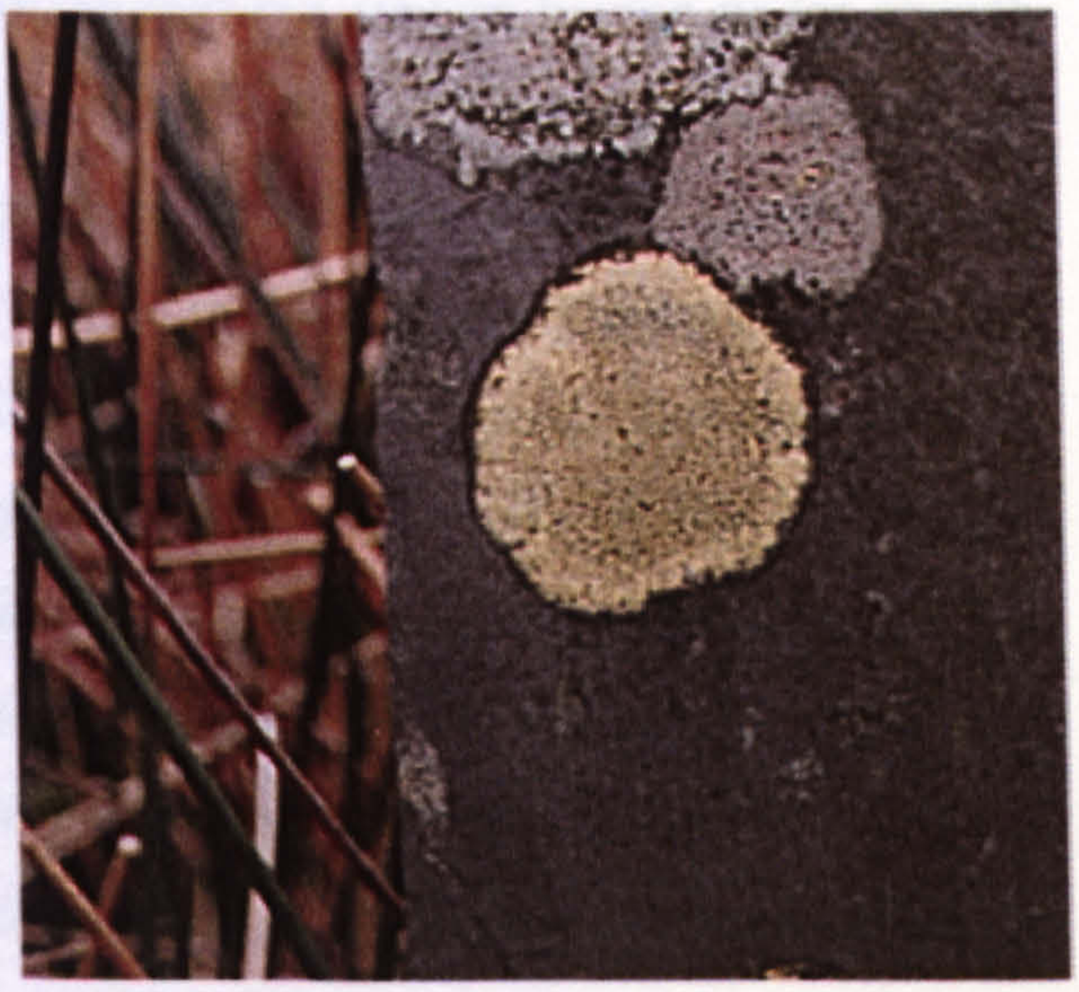
Several workers have described the centrifugal growth of epilithic saxicolous lichens, although very few systematic measurements have been carried out on crustose species. Trinci (1971) first demonstrated the existence of a peripheral growth zone in fungal colonies. After removing the central portion of tissue, he found that the outer ring continued to grow, and that the width of this marginal zone was a function

Fig. 5.1 The crustose lichen – *Rhizocarpon* Ram. Em. Th. Fr. Subgenus *Rhizocarpon*

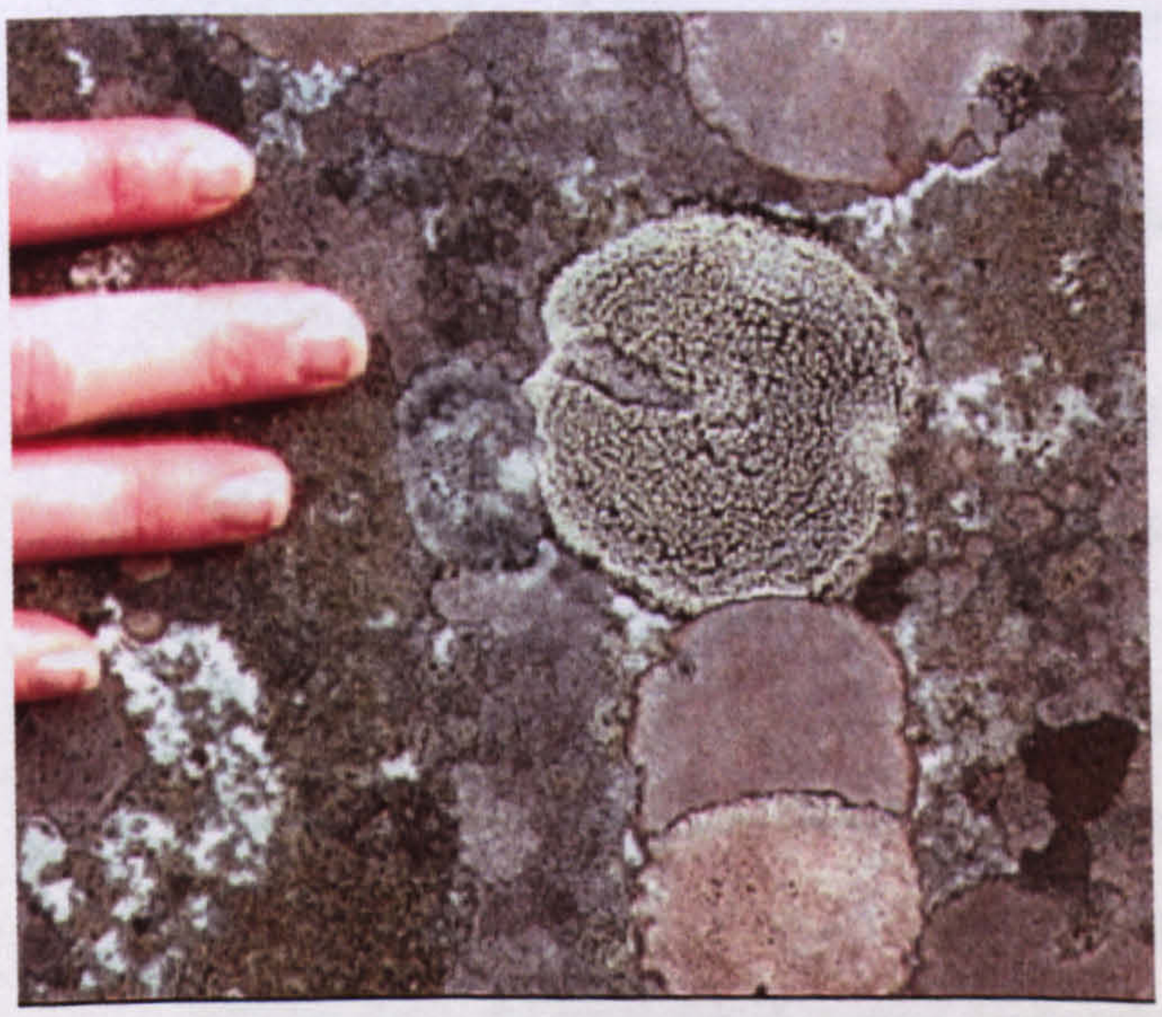


(i) *Rhizocarpon geographicum (L.) DC.*

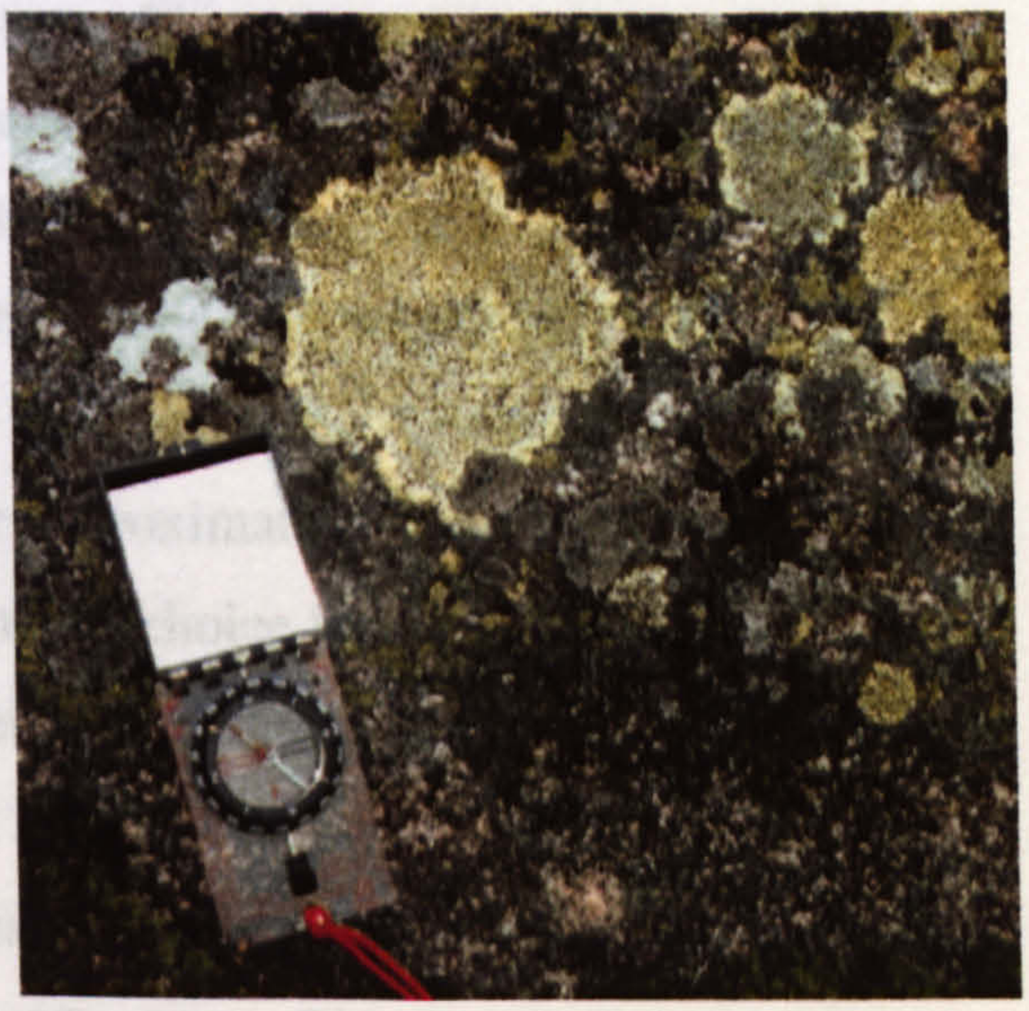
Black areolae set within yellow-green *trebouxia* algae. [5x magnification]



(ii) 42 mm yellow-green *Rhizocarpon* thallus growing on slate fencepost, Nant Ffrancon, Gwynedd, Wales. Note the wide, black, marginal hypothallus.



(iii) *Rhizocarpon* Section *Rhizocarpon*.
53 mm diameter thallus –
St.Columba's Isle, Skye, Scotland.



(iv) *Rhizocarpon* Section *Alpicola*
182 mm diameter thallus – Coire an Lochain, Cairngorms, Scotland.

of nutrient concentrations within the colony. The peripheral growth zone also varied in width from species to species.

Benedict (1985) asserts that lichen growth occurs at the periphery with only a narrow band or *annulus* contributing to thallus diameter increase. He suggests that the growth rate of crustose species, such as *Rhizocarpon geographicum,* is related to the width of this marginal zone, with slower-growing species having wider growth zones. His observations are supported, at least in part, by various other workers (Topham, 1977; Proctor, 1977, 1983; Armstrong, 1983; Innes, 1985; Armstrong and Smith, 1997) who find little evidence to doubt the marginal growth model for crustose lichens, including those within the *Rhizocarpon* subgenus. A recent study by Armstrong and Bradwell (*in press*) has found that the width of the hypothallus margin in *Rhizocarpon* may be an indication of thallus growth rate, although the physiological reasons for this link are still unclear. A few workers, such as Hill (1981), still dispute the presence of a marginal growth zone, finding little structural or anatomical evidence. Nevertheless, experiments have consistently shown that the central portion of larger lichens is not essential for continued growth (Armstrong, 1979; Armstrong and Smith, 1996).

A few, relatively fast-growing lichen species, such as *Pertusaria* and *Lecanora,* exhibit visible growth bands or marginal zonation. This concentric pattern results from growth over periods of several years and permits various inferences to be made. The growth contrast between summer and winter is discernible in the structural difference between the new cells generated at the margin. The pattern of marginal zonation also allows the age of the thallus to be calculated. Counting rings in a similar fashion to dendrochronology permits the approximate age of the thallus to be determined. This species would serve as an obvious choice for lichenometric dating studies, except for the fact that it is uncommon at high altitudes and high latitudes where most lichenometric work is carried out. Unusual though *Pertusaria* may be, other relatively fast-growing lichens almost certainly exhibit similar annual zonation, which can only be seen under powerful magnification. This has not been fully explored in lichenometry but could facilitate the construction of absolute growth curves for certain species.

Although growth refers to an increase in mass, most workers resort to measuring changes in surface area or marginal extension of the lichen thallus. Approximately 10% of the plant body by weight is attributable to algal cells, therefore it is the fungal component of the lichen (90%) that is responsible for most of the mass increase during growth.

Aplin and Hill (1979) proposed that lichen growth is governed by two basic biological processes: firstly, production of carbohydrate by photosynthesis and secondly, consumption of carbohydrate across the whole thallus at a rate proportional to the carbohydrate concentration. In the central areas this carbohydrate consumption causes thickening; at the margins it results in an increase in thallus radius. Childress and Keller (1980) used these basic assumptions to derive a mathematical model of lichen growth. Their model fits the empirical data collected by Proctor (1977), at least over the initial growth phases. Childress and Keller's model also tends to support the view that carbohydrate diffusion radiates outwards from the centre. Furthermore, they predicted that the transition from an initial exponential, or '*prelinear*', growth phase into a period of relatively constant growth rate would take place at a certain radius, determined by the rate of this diffusion. The mechanisms responsible for radial growth in larger (ie. older) lichens are not well understood. Theoretical modelling has not yet addressed this problem satisfactorily.

In *Rhizocarpon*, as in many lichens, carbohydrate is released from the algae as ribitol and converted into a more manageable form by the fungal component (Hill and Smith, 1972; Farrar, 1976). However, in *Rhizocarpon*, unlike many other lichens, it is thought that the margin obtains materials for growth from two sources – internally, from the areolae or 'pioneer' algal cells in the marginal hypothallus and externally from the decomposition of the rock substrate or by leaching nutrients from the surrounding soil and neighbouring lichens (Farrar and Smith, 1976; Armstrong, 1993a). Armstrong and Smith (1996) have shown that, although the hypothallus may utilise exogenous materials, radial growth in *Rhizocarpon geographicum* is driven primarily by the transfer of carbohydrates from the areolae through the fungal cortex.

A good deal is known about the biological processes responsible for growth in foliose and fruticose lichen varieties. The main findings have been comprehensively summarised in works by Hale (1967), Seaward (1977), Galun (1988a, 1988b, 1988c) and Nash (1996). The mechanisms governing growth in *Rhizocarpon* and similar – *Rhizocarpon*-like – crustose species, however, are still poorly understood. Ongoing research is currently addressing some of these problems (M. Jahns, *pers.comm.*, May 2000). However, it is probable that certain crustose lichen species possessing an algal-free marginal hypothallus, such as *Rhizocarpo*n, do not behave as a uniform plant or undergo simple radial growth throughout their entire life cycle.

Lichen growth measurement

Lichens are renowned for their slow growth. Yet surprisingly few detailed studies on the long-term growth rates of lichens have been carried out. Although progress has been made over the last fifty years or so, studies have not confirmed the longevity of lichens or directly measured growth rates over time periods of more than a decade, with two notable exceptions (Frey, 1959; Lange, 1990).

Casual observations on lichen growth have been found to exist from the 17th century (Tournefort, 1694). Acharius (1798) was probably the first to comment on the rates and mode of lichen growth, classifying them for the first time as a distinct order of *cryptogamia*. Observations increased during the 19th century due to the heightened scientific interest in all living things. In 1825 Meyer first published data relating to the growth rate of lichens. However the first detailed treatise on the subject was not published until half a century later. Schneider's (1897) volume offered a comprehensive summary of the state of knowledge at the close of the nineteenth century. Currently, at the beginning of the 21st century, the number of scientific studies on lichen growth numbers into the hundreds.

The direct approach

Traditionally, lichenologists measure lichen growth directly. This entails measuring a specific dimension of the lichen thallus, normally a function of the diameter, at regular time intervals. The time interval required to identify observable growth is

related to the growth rate of the particular species in question. Crustose species, the slowest growing of the epilithic lichens with typical radial growth rates of less than 1 mm per year, require long-term research commitments to obtain meaningful results.

Direct measurements of lichen growth from year to year provide valuable information on the rates of algal and fungal development which, in turn, can be related to external variables, such as climate, air quality and nutrient supply. It must be stressed, however, that directly-measured growth **cannot** be used to infer growth rates outside the measurement period. Extrapolation of such growth curves over longer periods is inappropriate as this assumes constant, optimum growth under uniform conditions. In lichenometric studies this will almost never be the case.

Various techniques have been utilised to directly measure lichen growth, each with its own level of accuracy. A less precise technique involves measuring the diameter of a specially selected thallus, using a micrometer, calipers or ruler, and subsequently re-measuring the same lichen the following month or year (eg. Vallot, 1896; Hausmann, 1948; Frey, 1959; Hale, 1954). The radial growth rate can be found by dividing the measured growth increment by two. The accuracy of such results ultimately depends on the device used for measurement. Micrometer-measured errors of +/-0.1 mm are typical, whereas ruler-measured errors are likely to be +/-0.5 mm. Generally, this technique is satisfactory for obtaining *average* growth rates over long periods (ie. two or more years).

Measurement of marginal growth in crustose species and lobe elongation in foliose species can also be expressed in relation to a fixed point on the substrate near the thallus in question. Linkola (1918) successfully measured growth in several saxicolous specimens from a chisel mark in the rock. The technique of setting up permanent base points from which to measure radial growth has been adopted by many workers (eg. Hakulinen, 1966; Armstrong, 1983, 1993b; Benedict, 1990a). Hale (1970) used a point on the surface of the thallus itself as a base point from which to measure lobe elongation in relatively rapid-growing foliose species. Marginal growth measurements of this nature may yield a more accurate assessment of growth than simple diameter measurements, although the precision of the results is again dependent on the nature of the measuring equipment.

Tracing the outline of the thallus onto acetate or clear plastic allows growth in all directions to be monitored, rather than merely one. Comparing the original tracing

with the same specimen some years later permits a spatial expression of growth to be calculated. Past proponents of the tracing method include Hale (1954, 1959), Rydzak (1956, 1958, 1961, 1966) and Brodo (1965). Rydzak worked with crustose and foliose species growing in forested areas of southern Poland and derived growth rates in square millimetres per year. Areal growth rates need to be interpreted with caution because, as Brodo (1965) points out, *areal* growth rates are largely a function of the size of the thalli. Radial growth rates can, however, be effectively calculated from areal measurements ($r = (A/\pi)^{1/2}$), as shown by Hale (1959). The accuracy of the tracing technique depends primarily on the width of the writing implement (typically, +/- 0.2 mm) and the precision of the scribe (realistically, +/-0.2mm). Radial growth increment in millimetres is now the preferred measurement parameter.

The most widely advocated technique for measuring lichen growth is photography. Being less cumbersome and more accurate than the methods described so far, photographs provide a faithful, permanent record of the specimen in question. High-quality, close-up photography combined with image enlargement facilities allow growth to be detected at the micron scale (0.001 mm) (Hale, 1970). Sub-millimetre growth measurements allow variations over periods of several months, or even weeks in some species, to be monitored.

Recommended by Beschel (1958) as the preferred technique, photography has been used by various lichenologists to determine growth over relatively short time periods (eg. Phillips, 1963, 1969; Hakulinen, 1966; Hale, 1970, 1974; Ten Brink, 1973; Hooker and Brown, 1977, Proctor, 1977, 1983; Hooker, 1980; Benedict, 1990a). Advocacy of the photographic technique will allow more permanent measuring stations to be set up and revisited at convenient times in the future. Assuming that a lichen thallus does not sustain damage, become atrophied or die it would be possible, patience permitting, to monitor the life history of a crustose lichen. A photographic record of such a thallus developing completely over several decades would be an invaluable asset to the study of lichenometry.

The indirect approach

Since the introduction of lichenometry 50 years ago (Beschel, 1950), lichen growth rates have also become the interest of earth scientists. Lichen colonies on surfaces of known age offer a time-saving way of studying the growth rates of lichens over

periods of many decades. Rather than measure growth rates *directly*, lichenometrists utilise this *indirect* method to ascertain the growth rate of slow-growing crustose species and hence the age of the surface in which they are interested.

It must be remembered that growth rates derived using the indirect method are less accurate than those determined by direct monitoring. Various assumptions must be made relating to ecesis (establishment), colonisation lag time, and growth regularity from place to place and through time. Ultimately, the indirect approach provides only an approximate, minimum figure rather than an absolute lichen growth rate.

A new approach, currently being developed by Clark *et al.* (2000) involves determining the growth rate of lichens by directly dating the living tissue. Using mass spectroscopy ^{14}C they determined the absolute age of the tissue at the centre and the margin of a single crustose *Caloplaca trachyphylla* thallus. Their results suggest that the thallus was 20 years old and that marginal growth had been relatively constant over that period. Initial tests on *Rhizocarpon* thalli have been less promising (B. Clark, *pers. comm.*, May 2000). Further studies of this nature are needed before the method can be applied more generally. Currently the technique is limited to the last 50 years. However, this new radiometric technique could provide an independent alternative for indirectly deriving the growth rates of lichens.

Lichenometric studies often involve determining the maximum lichen growth rate within a certain area. Many workers have sought to ascertain this by measuring the largest, and presumably fastest-growing, thallus on a dated surface. Much debate surrounds the optimum number of thalli to be measured (1, 5, 10, etc.) and the thallus parameter that should be measured. However, the methodological issues regarding the various merits of the 'largest-lichen' and the 'mean of the five largest-lichens' have been addressed in detail by Innes (1982, 1984, 1986).

The indirect approach, unlike the direct approach, does not require re-measurement of thalli at a later date; therefore the precision of measurement is rarely high. Many studies have used a ruler to determine lichen dimensions to the nearest millimetre (eg. Andrews and Webber, 1964; Benedict, 1967, Gellatly, 1982; Caseldine, 1990; Smirnova and Nikonov, 1990; Bradwell, *in press*). This level of accuracy is realistic in view of the errors associated with surface roughness, thallus margin definition and

operator subjectivity. Recent studies (Bull *et al.*, 1995; Bull and Brandon, 1998; McCarthy, 1999) have advocated the use of dial or digital calipers capable of accuracies of +/-0.01 mm. A critical, comparative study of the two approaches has not been carried out. However, the need for micron accuracy in lichenometry, bearing in mind that between-operator variance can be significant (95% level; Innes, 1982), may be somewhat unnecessary.

The three most often used measurement parameters are longest axis, shortest axis and the largest inscribed circle. Many indirectly-derived lichen growth rates are difficult to compare due to the different measurement parameter adopted. Long axes measurements have been deemed most representative of optimum growth by various workers (eg. Innes, 1985; Matthews, 1994; Evans *et al.*, 1999), as thallus irregularity or elongation is likely to result from growth being suppressed in one particular direction. Other parameters such as largest inscribed circle (= shortest diameter) (Locke *et al.*, 1979), average diameter (ie. [long + short]/2) (Pitman, 1973), and largest theoretical diameter (Erikstad and Sollid, 1986) all employ some degree of subjectivity and are generally less replicable.

Growth rates

Published studies of directly-measured lichen growth presently number in the hundreds. A summary of some of these allows a good impression to be gained of the typical growth rate that might be expected for a certain lichen species. Due to the nature of this review, only foliose and crustose species have been included in Table 5.2 below. Special emphasis has been placed on the growth rates of *Rhizocarpon* lichens. It can be seen from Table 5.2 that typical growth rates differ widely depending on the lichen species in question. However, comparison of growth rates between species is complicated by two other factors:– thallus age and climate.

Table 5.2 Directly-measured average growth rates of foliose and crustose lichens.

species	*radial growth rate mm/yr*	*author (date)*	*location of study*
Foliose			
Alectoria miniscula	0.50	Haworth *et al.* (1986)	Alaska
Cetraria pinastri	1.15	Hakulinen (1966)	Lapland
Hypogymnia encausta	1.00	Frey (1959)	Switzerland
Lobaria pulmonaria	4.82	Phillips (1969)	Tennessee
Menegazzia terebrata	2.54	Phillips (1969)	Tennessee
Parmelia centrifuga	2.50	Linkola (1918)	Germany
Parmelia conspersa	1.60	Hale (1959)	Connecticut
	5.30	Phillips (1963)	Tennessee
	2.50	Armstrong (1974)	Wales
Parmelia glabratula	2.25	Armstrong (1974)	Wales
Parmelia sulcata	1.60	Linkola (1918)	Germany
	2.22	Degelius (1964)	Sweden
Parmeliopsi ambigua	0.70	Linkola (1918)	Germany
	0.90	Hakulinen (1966)	Finland
Physcia aipolia	1.30	Hakulinen (1966)	Finland
Umbilicaria deusta	2.30	Hakulinen (1966)	Finland
Xanthoria parietina	2.15	Hakulinen (1966)	Finland
	2.50	Degelius (1964)	Sweden
Crustose			
Buellia canescens	1.6	Proctor (1977)	England
Buellia aethalea	1.6	Armstrong (*unpub.*)	Wales
Diploshcistes scruposus	0.44	Hale (1959)	Connecticut
Rhizocarpon geographicum	0.03 – 0.23*	Beschel (1961)	Greenland
	<0.05	Hooker (1980)	Antarctica
	0.4 – 0.9	Armstrong (1983)	Wales
	0.48	Proctor (1983)	Swiss Alps
	0.33*	Matthews (1994)	Norway
Rhizocarpon obscuratum	1.8	Armstrong (*unpub.*)	Wales
Lecanora tumida	1.0	Armstrong (*unpub.*)	Wales

Note:
* values originally expressed as diameter increase have been converted to radial growth rates (ie. diameter increase per year / 2).

The growth of a lichen is rarely uniform over its entire life span (Hale, 1974). It is now believed that thalli of different age grow at different rates. Several studies have pointed towards slower growth rates in the smallest thalli (<10 mm) (Phillips, 1963; Brodo, 1965; Armstrong, 1983; Proctor, 1983), although problems surround the measurement of very small lichens (Nienburg, 1919; Beschel, 1958; Rydzak, 1961). Direct measurements have also demonstrated that the highest growth rates are normally attained once the thallus has become established and is going through a period of maturation (Frey, 1959; Beschel, 1958; Armstrong; 1974, 1983; Proctor, 1983). The largest thalli may grow at a reduced rate. Comparison of growth rates should therefore only be made between specimens of similar size, assuming that they are in the same phase of growth.

Phases of growth

Biological growth in plants normally describes a sigmoidal curve. Sachs (1882) coined the term "Grand (or Great) Period" to refer to the cumulative period of growth that includes an initial slow phase followed by rapid growth which gradually declines and ultimately ceases in old age. This is the definition of 'Great Period' most commonly used by plant physiologists (Whaley, 1961).

Within the lichenological literature there is some confusion surrounding the use of the term "Great Period". Beschel originally proposed that: "the period of relatively rapid growth be termed the 'great period'" (Beschel, 1973: 303; (translated from 1950: 152)). He suggested, based on his own studies, that it should end "within a few decades". This statement may have influenced future lichenological and lichenometric studies (eg. Andrews and Webber, 1964; Mottershead and White, 1972; Proctor, 1983; Kugelmann, 1991; Bull *et al.*, 1995; Smith *et al.*, 1995). In a comprehensive review on lichenometry, Innes (1985) equates the *linear* phase of growth with the 'great period' described by Beschel (1950). Armstrong (*pers.comm.*, 2000) prefers to assign the 'great period' to the complete phase of accelerating and linear growth, which ends when radial growth starts to decline. This, however, is an unfortunate use of terminology as the *linear* phase (ie. constant radial growth rate) is

now encompassed within the 'great period, with the subsequent declining growth rate phase often being termed "*linear*" (Armstrong, 1974, 1975). Furthermore, in his review on lichen growth, Hale (1974) chooses to distinguish between the initial *juvenile* period and "The 'Great' Period" (*sensu* Beschel, 1950). The grounds for doing so are still unclear, especially as Beschel (1950) may not have supported such a distinction. To avoid unnecessary complication, the term '*great period*' will not be referred to further within this review. It is suggested, instead, that the various growth phases in the life of a lichen thallus be referred to as: establishment, juvenile, maturation, maturity and senescence.

a. Establishment / Colonisation

Prior to any physical expression of growth a lichen propagule must become established on a substrate. Very little work has been done on the nature and relative speed of propagule colonisation although it is known to depend on four key factors:- supply of spores, substrate inclination, substrate texture and micro-climate. Bailey's (1975) review of lichen propagule establishment treats the subject of dispersal, deposition and attachment of spores very thoroughly. Most workers agree that some time must elapse, which may be "a considerable period" (Beschel, 1950: 152, *translated* - 1973), before the lichen thallus becomes visible (ie. macroscopic) on the rock surface. This microscopic phase of growth is often termed the '*colonization lag-time*' by many lichenometrists. There is little knowledge regarding growth during the microscopic phase. It is probably controlled chiefly by substrate texture. This initial, microscopic development of the thallus should be referred to as the ***establishment period.***

b. Juvenile

Armstrong (1974, 1976) classifies the growth history of a lichen thallus into three distinct phases: (i) pre-linear, (ii) linear and (iii) post-linear. The *pre-linear* phase is supposed to directly follow establishment of an algal-fungal propagule on the substrate, although no accurate definition exists. Armstrong (1976) equates his pre-linear phase with the 'juvenile' period proposed by Hale (1974). Hale (1967, 1974) considers crustose and foliose thalli less than 10 mm in diameter to be in the juvenile phase. Direct measurements of the radial growth of small thalli (Hale, 1967 - *Parmelia conspersa;* Armstrong, 1973, 1974 - *Parmelia glabratula;* Proctor, 1977 -

Buellia canescens; Armstrong, 1983 - *Rhizocarpon geographicum*) would tend to support this statement. Growth rates during this ***juvenile*** phase are variable. However most studies have shown that radial growth is logarithmic (Armstrong, 1973, 1974, 1976; Proctor, 1977, 1983), being slow initially and then accelerating through time until a certain diameter is reached – typically between 10-15 mm. Debate still surrounds the exact shape of the growth curve in thalli less than 10-15 mm in diameter. A recent pilot study by Clark *et al.*, (2000), using AMS ^{14}C to date lichen tissue directly, points towards a linear relationship between thallus size and age over the first 20 years of the lichen's life (*Caloplaca trachphylla*; 60 mm in diameter). Further studies are needed to verify these initial findings.

c. Maturation

There is no consensus on the duration or even the form of the second phase of lichen growth after propagule establishment. Armstrong (1976) refers to a *linear* phase of growth where the radial growth rate is constant over many years.

Direct measurements suggest that lichens, once of a certain diameter, will continue to increase in size at a relatively constant rate. The juvenile phase can be said to have ended, therefore, when radial lichen growth ceases accelerating and a constant growth rate is established. The linear growth phase will vary in duration from species to species and from place to place, but normally covers the entire period of ***maturation*** and may last for several (3-10) decades. In *Rhizocarpon geographicum*, for example, Armstrong (1983) found that a generally constant growth rate existed in thalli over 15 mm and up to 45-50 mm in diameter. Where radial growth rates are low (<0.5 mm/yr), due to less favourable environmental conditions, this *linear* phase may last in excess of 100 years. Lichenometrists working in high latitudes have amassed evidence suggesting that certain species (particularly *Rhizocarpon*) can grow at uniform rates for many centuries (Beschel, 1961; Denton and Karlén, 1973; Lindsay, 1977; Hooker, 1980).

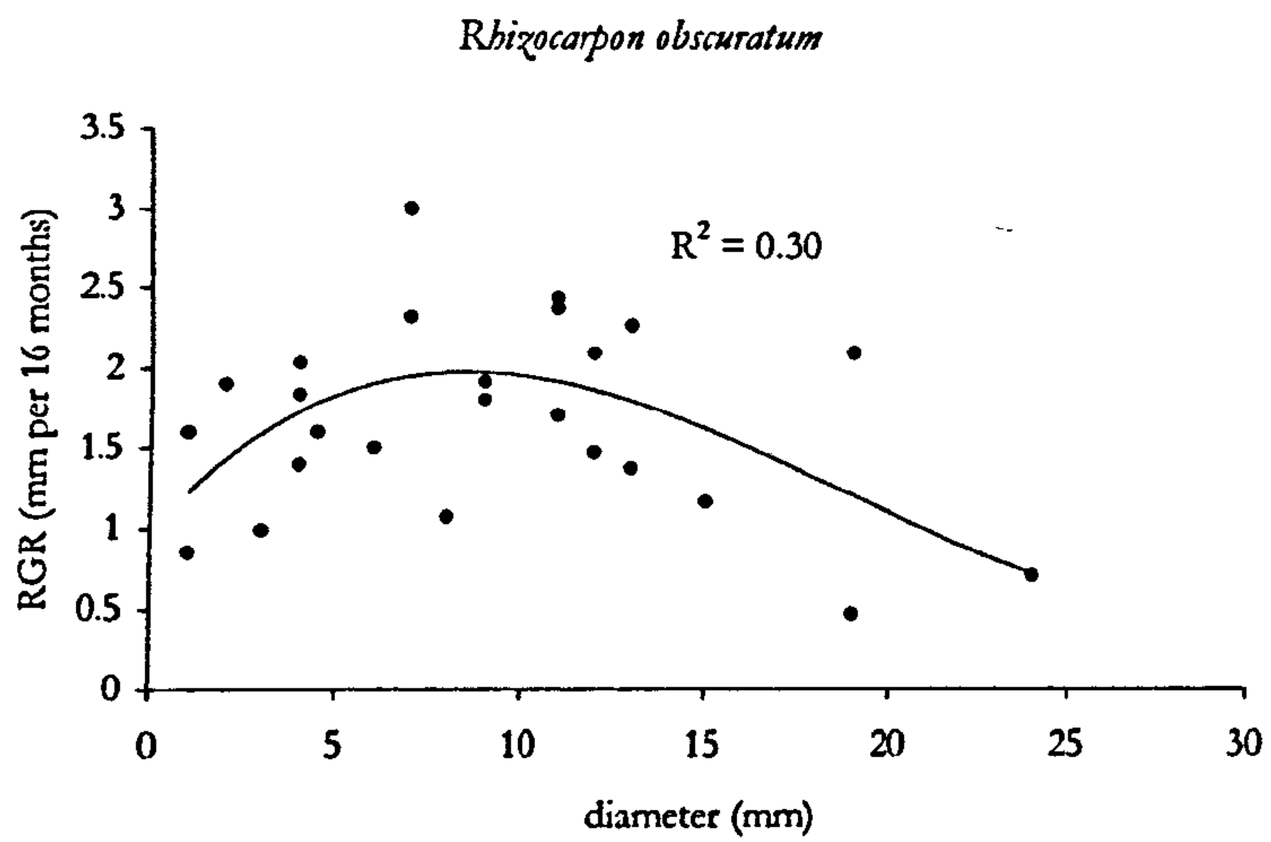
d. Maturity

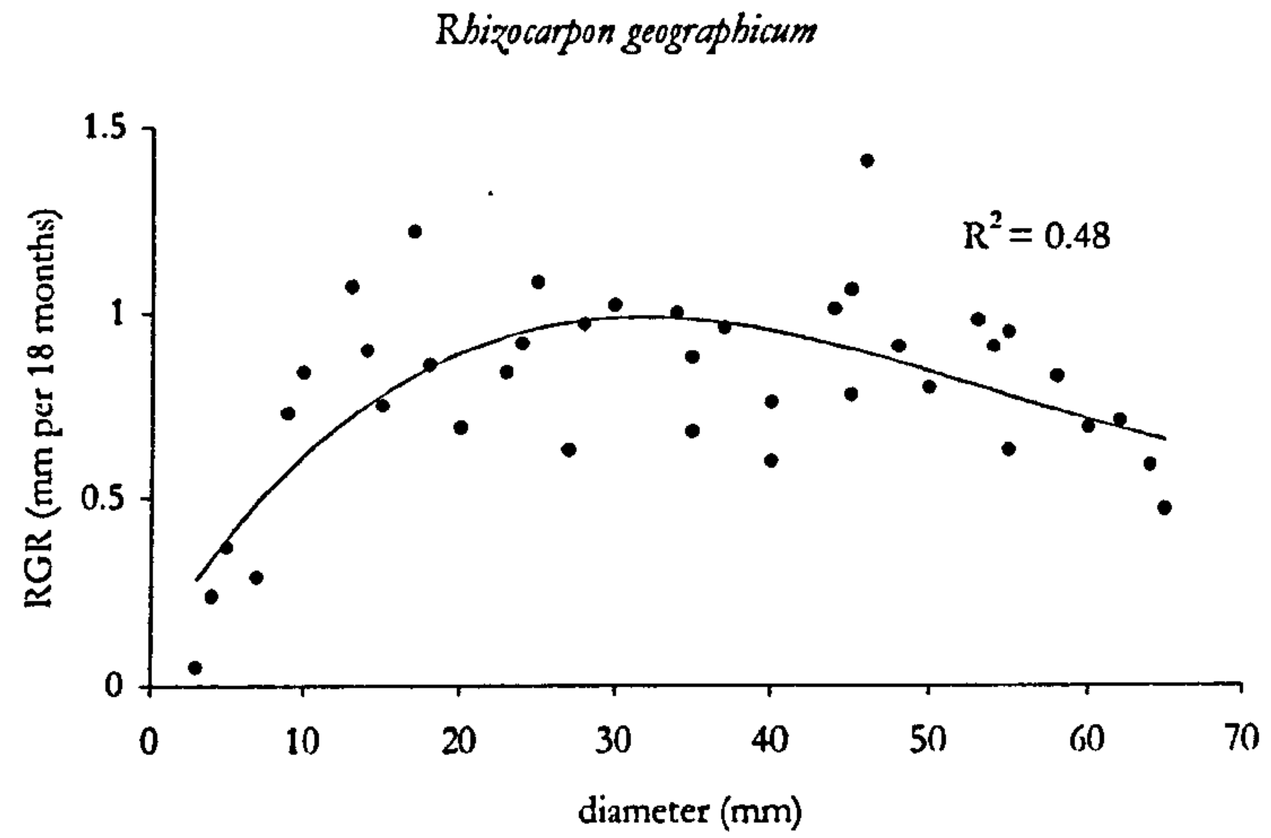
Once maturity is reached, lichen growth rates slow down. The definition of ***maturity*** in terms of age is far from unanimous, however most lichens enter a phase of declining growth rate once they have reached a certain size. Very few direct studies have detected the slowing of radial growth in crustose lichens. Hale (1967)

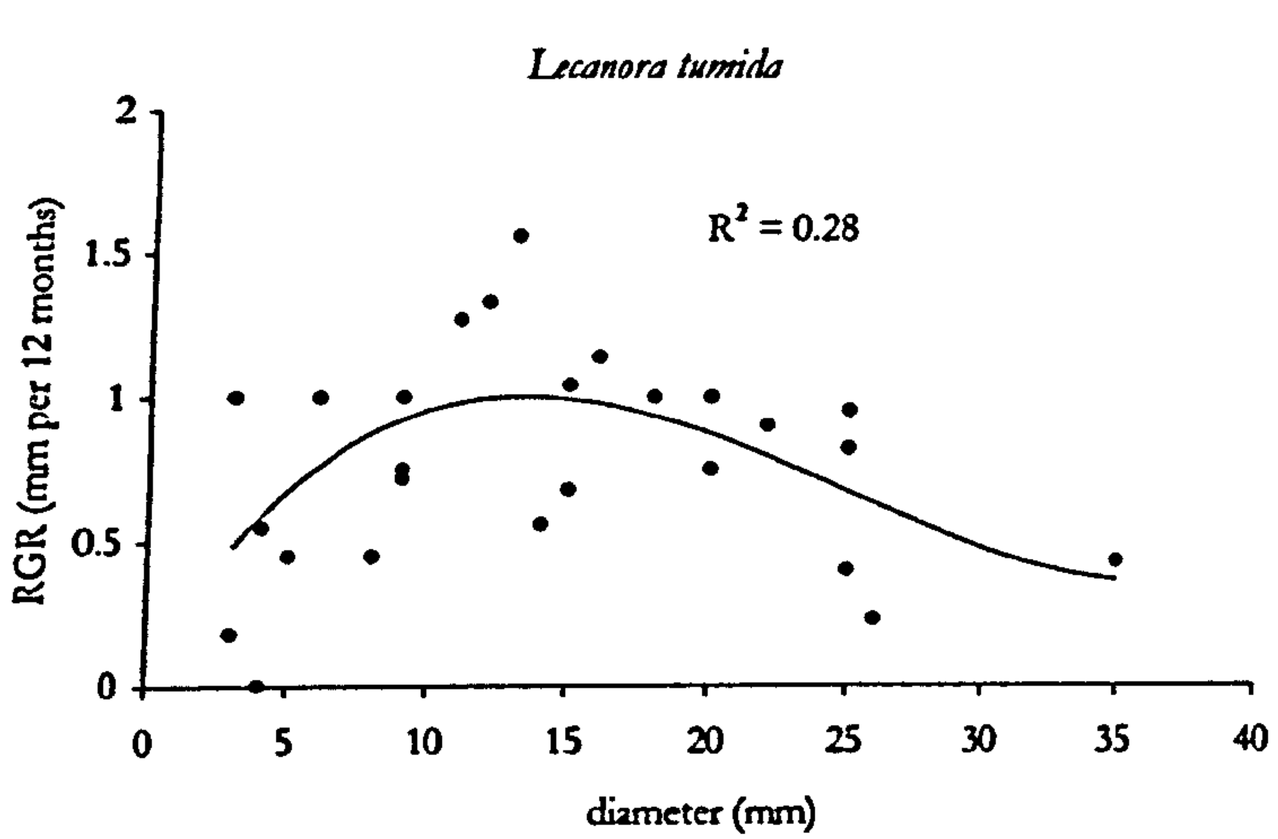
was possibly the first to observe this growth rate decline in foliose species. Working with thalli of *Parmelia conspersa* he detected the largest thalli (>120 mm) to be growing at significantly lower rates than those of small and medium size. Armstrong (1974) also observed declining growth rates in several saxicolous crustose lichens which he termed the '*postlinear*' phase. Direct studies on *Rhizocarpon geographicum* have led Armstrong (1983) to confirm that this species also enters a slower, '*postlinear*' growth phase. A recent study by Armstrong (2000, *unpublished*) has detected a declining radial growth rate in three other crustose species (*B. aethalea, R.obscuratum and L.tumida*) (Fig. 5.2). The point at which growth rates begin to decline varies in all three species and has not been determined precisely. Nevertheless, Armstrong's recent work does show that growth rate decline is most marked in *Rhizocarpon obscuratum* and least marked in *Rhizocarpon geographicum.*

The scarcity of data relating to this final phase of growth, and the physiological mechanisms involved, partly account for the controversy still surrounding it. Recently, a directly-measured study of growth in thalli of the *Rhizocarpon* subgenus by Matthews (1994), did not detect a decline in the growth rate of the largest thalli (>50 mm), although the use of an aggregate 'species' – including thalli of both Sections *Rhizocarpon* and *Alpicola* – does not allow direct comparison with Armstrong's findings (1983).

The length of this mature, so-called *postlinear*, phase is still uncertain. Some researchers have postulated that crustose lichens must have a maximum size at which growth ceases. It is difficult to assess the duration and rate of growth during the *postlinear* phase based on current knowledge. The range of extreme dimensions reported for various species confirms that growth is finite in many species (*Parmelia sulcata* = 90 mm, Beschel 1958; *Parmelia conspersa* = 120 mm, Hale 1973; *Parmelia caperata* = 200 mm, Hale 1973). However, no such maximum has been recorded for *Rhizocarpon*, with thalli of 150 mm being reported for Section *Rhizocarpon* (Proctor, 1983; Orombelli and Porter, 1983) and in excess of 450 mm for Section *Alpicola* (Denton and Karlén, 1973). Some thalli may continue to grow slowly, possibly at an exponentially declining rate for very long periods of time (Mottershead and White, 1972; Denton and Karlén, 1973; Matthews, 1974, 1994; Innes, 1985, 1988).







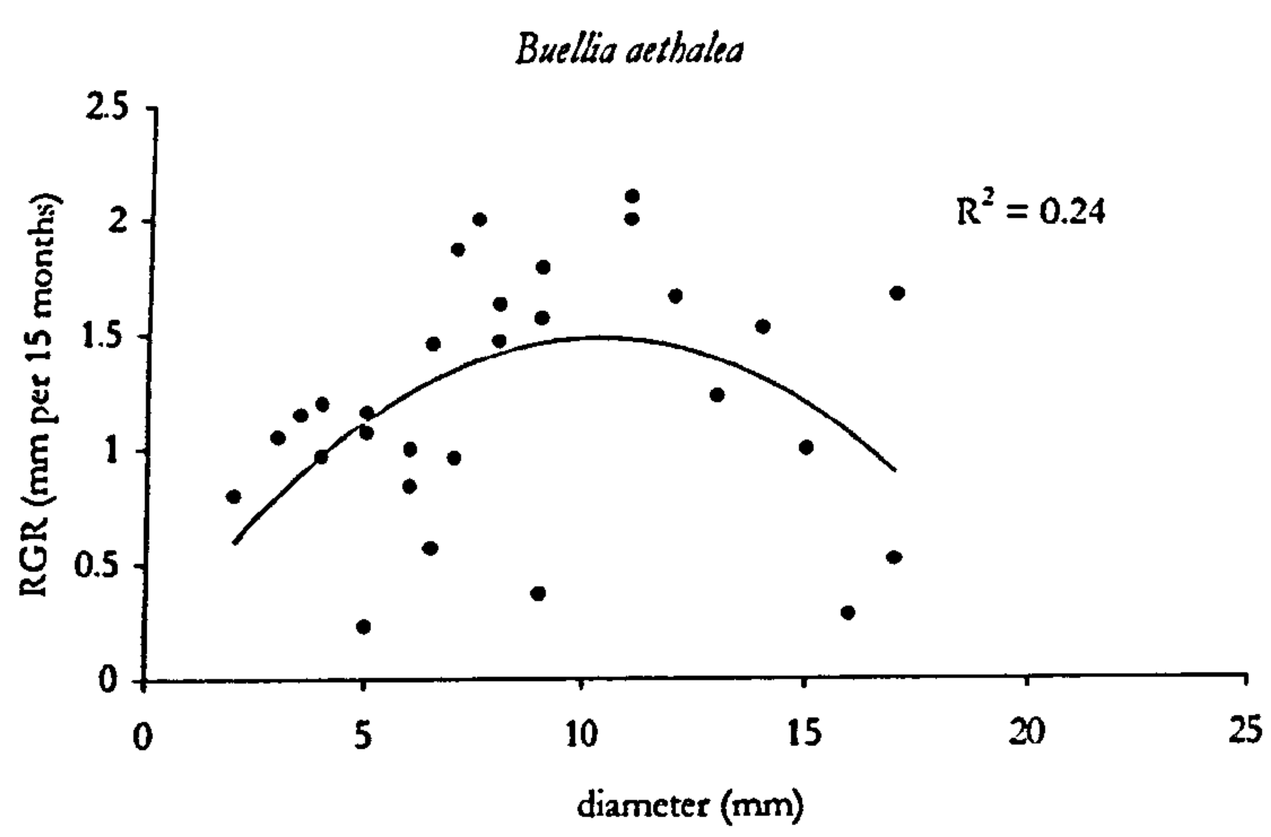


Figure 5.2 Directly-measured radial growth rates (RGR) of four crustose lichen species (from Armstrong, 2000 *unpub*.)

e. Senescence

Those lichens that have lived long and matured fully, ultimately, like all living organisms, reach the end of their growth cycle. In some species this ***senescence*** period is accompanied by disintegration of the thallus centre. Certain species will fragment entirely in old age with isolated patches able to re-generate and establish new colonies. Armstrong and Smith (1997) showed convincingly that thallus degeneration is an age-related process at least in certain species (eg. *Parmelia glabratula*). Other species, however, can experience disintegration of the thallus centre at various stages in their lifetime with little effect on their outward growth. Studies have shown that certain species, including *Rhizocarpon geographicum*, can continue to grow with their central portion removed (Armstrong, 1979; Armstrong and Smith, 1996). Thallus fragmentation is therefore not always an indicator of age and will not always affect radial growth rates. True senescence periods, accompanied by no further appreciable growth, can last indefinitely until finally the thallus disintegrates or becomes detached from its substrate.

Summary

Lichen growth rates are not constant over time. Both direct and indirect measurements suggest that over long time-periods growth rates generally become slower as lichen thalli get older. Most studies show non-linear, three-phase or exponentially declining growth in long-living, crustose lichens. However, the subject is still the focus of debate and may be more complex than early studies claim. Direct measurements of growth in large, crustose thalli (>50 mm) are lacking. Those data that presently exist are equivocal; suggesting that a linear growth model may be more applicable in certain species and certain environments. Until further direct measurements – both in the laboratory and the field – are carried out the exact shape of a 'typical' lichen growth curve remains undetermined.

The role of climate on lichen growth

Lichen growth rates vary as a function of climate. Environmental factors, such as moisture availability, temperature, light intensity and ventilation, are known to affect photosynthesis, assimilation, respiration and, in turn, growth rates in lichens. This section briefly reviews the various studies on systematic variations in lichen growth rates.

Environmental factors affecting growth

Growth in lichens occurs, as in all organisms, when there is positive net assimilation. Assimilation is a physiological process and is therefore controlled by the available amounts of moisture, heat and light etc. Normally, assimilation, and hence growth, is the result of a combination of these factors. When one or more of these factors is deficient, growth will be curtailed. Hence, lichens of the same species will grow at different rates in different climates. Significant differences in growth are also to be expected for example in winter when temperature levels are in marked contrast to summer, in temperate regions. Rydzak (1961) was the first to detect both seasonal and inter-annual variations in the growth of crustose and foliose lichens, which he attributed to the severity of the Polish winters and the relatively mild, wet summers. Smith (1962) also detected seasonal variations in growth in relation to optimal photosynthetic activity during spring and autumn. Phillips (1963) detected *Parmelia conspersa* in Tennessee to be growing at over three times the rate of the same lichen in Connecticut (Hale, 1959). This difference in growth rate can be attributed, for the most part, to the warmer and wetter climate of Phillips's study site. Beschel's detailed, indirect, studies in the 1950s and 1960s underlined the important role played by climate on lichen growth. He found the same species of *Rhizocarpon* to be growing at a fraction of the rate in Greenland when compared to rates in the Swiss Alps (1958, 1961). In order to understand lichen growth variations, the response of the organism to certain environmental variables must be examined properly.

Light

Radiant energy is essential for photosynthesis in lichens as in all plants. Link *et al.* (1985) found that there was a dependence of photosynthesis in certain species on light (radiation) levels. However, it has also been shown that light saturation is

dependent on thallus water content (Lange, 1969) and temperature (Kershaw, 1985). Light transmission is more efficient through a wet thallus cortex (Ertl, 1951), which may partly explain Lange's (1969) findings.

Lichens growing at high altitude or in polar regions may be subject to very high light intensities. Exposed rock surfaces in high latitudes experience intense light levels during the long summer. The effects of such high radiation levels are difficult to quantify, although certain species thrive in such localities. Wiesner (1895) suggested that the need for light increases as temperatures decrease with latitude or altitude. In Arctic and Antarctic regions, crustose lichens may be dependent on high light intensities for warmth where air temperatures would otherwise prevent photosynthesis. Examples of photo-philous 'sun-lichens' are *Cladonia*, *Parmelia*, *Umbilicaria* and *Diploschistes*. The distribution of certain species, such as *Rhizocarpon geographicum* (L.) DC., may involve a subtle balance of light levels, rock temperatures, and moisture availability within a small area. Such factors could explain a lichen species' preference for shady localities in certain habitats whilst exposed high-energy-receiving surfaces are preferred in others (Beschel 1956; Armstrong, 1975; Benedict, 1990a; McCarthy, 1997, 1999).

The influence of light conditions on the physical appearance of certain lichen species is well documented. Smith (1962) describes how thalli of one species growing in full sunlight will take on a darker pigmentation than similar thalli growing in shady locations. This adaptation is thought to be a protective response to increased levels of ultra-violet radiation at high altitudes (Billings and Mooney, 1968). In certain species, however, the production of brightly-coloured acids is stimulated by high light levels and clear air (Kappen, 1973). *Rhizocarpon geographicum* (L.) DC., for example, is a vivid yellow colour at high altitudes whilst being shades of green at lower elevations.

Field studies have yet to show equivocally that increased light promotes growth. Benedict (1990a) found a strong correlation between the growth of *Xanthoparmelia* and the number of daylight hours, over a two-year period. Longer days combined with higher angle sunlight and a lack of snow-cover could explain increased growth during the summer. In practice, it is very difficult to isolate the effects of light on the thallus, especially in relation to heat and moisture levels. For example, any reduction

in growth caused by lowered light levels might be offset by increased moisture levels in a shaded habitat.

Cloudiness

Cloud cover effectively lowers air temperatures, increases air humidity and suppresses light and radiation levels. To date, no studies have concentrated on this meteorological variable, probably due to the difficulty involved in the quantitative assessment of cloudiness. However, a few workers have suggested that increased cloud cover in a particular season may promote lichen growth. Hausmann (1948), without any precise meteorological data, found that growth of *Parmelia centrifuga* was up to 80% greater during the cloudy summer of 1941 than in the preceding two, or subsequent four years. Rydzak (1961) found "best" growth during the years with higher summer relative humidity, suggesting that wetter, cloudier conditions promote lichen growth.

Moisture availability

Water appears to be the primary environmental factor affecting the growth potential of all lichens. Epilithic saxicolous species cannot gain moisture from the soil like other small hardy plants, hence their only source of water is that which precipitates from the atmosphere, either in the form of rain, snow, dew or vapour. Many studies describe a positive relationship between moisture and lichen growth, with low moisture levels resulting in low metabolic activity (eg. Lange, 1953; Ensgraber, 1954; Kershaw, 1972; Armstrong 1974, 1993; Paterson *et al.*, 1983). The relationship is not simply linear, however, and debate still surrounds the optimum saturation conditions under which lichens will grow. Beschel (1950) argued that "thoroughly saturated conditions" were imperative for growth, whereas Kershaw (1972), Farrar (1976) and Coxson *et al.* (1983) all agree that total saturation depresses the rate at which photosynthesis can occur. The optimal conditions are complex and thought to depend on the rate and frequency of wetting and drying of the lichen thallus (Armstrong, 1973). Generally, moisture is essential for respiration and assimilation, with higher growth rates resulting from increased wetness.

Rainfall

Several field studies have demonstrated the link between the radial growth rate of lichens and rainfall. Nienburg, (1919) proposed that excessive rainfall in May 1889 most likely accounted for the high growth rates measured in southern Germany that year. Karenlampi (1971) monitored the fruticose lichen *Cladonia* during one summer in Lapland. He found that the relative growth rate over the four-month period was best correlated with daily rainfall ($r = 0.973$). Armstrong's (1973) study in North Wales showed lichen growth to be bi-modal in 1971 with maximum growth occurring in spring and autumn when rainfall levels were highest. In the Rocky Mountains of Colorado, Benedict (1990a) showed growth to be related to the length of time the thallus was snow-free and receiving moisture. He argued that at higher altitudes where lying snow persists for many months, growth conditions were only satisfied during the summer, whilst in low-altitude, temperate environments growth would be influenced all-year round by precipitation and humidity.

A slightly different approach adopted by Lawrey and Hale (1977) examined, amongst other variables, the role of rainfall frequency throughout the year. A fair correlation was found between the radial growth rate of two species and the number of rain days in Maryland, USA. A similar study on two foliose species in southwest England drew comparable conclusions (Fisher and Proctor, 1978).

Indirect (lichenometric) growth measurements of lichens on dated surfaces in the Swiss Alps and later in west Greenland allowed Beschel (1950, 1956, 1961) to show that the growth rate of saxicolous lichens was inversely proportional to the aridity (or hygro-continentality) of an area. The extremely slow growth rate of *Rhizocarpon* (<5 mm/100 yr) in certain parts of west Greenland can be attributed to the very low precipitation levels in the most continental regions. A subsequent re-investigation of Beschel's study sites confirmed these findings (Ten Brink, 1973).

Humidity

Some lichens rely on fluctuations in humidity as a source of moisture. There is currently uncertainty about the role played by moisture derived directly from the atmosphere. For water to be absorbed, the hygric potential of the lichen matrix must be lower than that of the atmosphere. Such conditions will only occur at relatively low temperatures (<20°C) and high relative humidities (>75%). The most likely environments where these conditions are satisfied are in coastal deserts where

relative humidity can regularly exceed 80%. Generally, the uptake of water from unsaturated air is unusual and is not a primary factor affecting the photosynthesis and respiration of lichens outside narrow climatic zones. There is some evidence, however, to suggest that variations of humidity may be more attractive to lichen communities than the maintenance of constantly high or constantly low levels (Hale, 1954; Pearson, 1969).

Dew

Formed by fluctuations in humidity and air temperature, dew is another important source of moisture for lichen growth. Where rainfall is scarce, dew can provide a regular source of moisture for photosynthesis and respiration (Butin, 1954; Kappen, 1988). Redon and Lange (1983) found lichens to be concentrated in areas of preferential dew precipitation, with growth being improved by the moist conditions. Thallus metabolism can be enhanced as dew prevents moisture loss after a period of rain. However, dew may be less significant in temperate and boreal regions, and where rainfall is plentiful throughout the year (Hale, 1970). Few field studies have focused on the role played by dew in promoting lichen growth, probably due to the lack of meteorological measurements of this phenomenon and the great degree of variation imposed by micro-habitat on dew formation

Fog

Certain species of saxicolous lichen can grow in extremely arid geographical locations (mean annual pptn <100 mm) by utilising the moisture in fog and low clouds. Several workers have concentrated on the particular lichen biogeography of the coastal fog zone. Kappen (1988) and Lange *et al.* (1990,1991) have worked in the fog zone of the Namib Desert in order to understand better the relationship between lichen moisture, photosynthesis and growth in arid areas. Certain species, particularly the multi-branched fruticose lichens can make maximum use of the moisture from fog clouds. The more compact foliose and crustose varieties may derive only a small proportion of the moisture needed to photosynthesise most efficiently (Lange, 1990). Fog will, however, delay drying in temperate and high latitude regions thus prolonging the time period in which photosynthesis can occur. Ahmadjian (1967) states that this is a crucial factor and may be most important in promoting rapid lichen development. If it occurs with regular frequency, Kappen

(1988) argues, fog can be a more effective source of moisture than rainwater (especially in *anombrophytic* species). Furthermore, Bliss and Hadley (1964) point out that light intensities for certain alpine species are at optimal levels during foggy periods.

Temperature

The effect of temperature on lichen growth is less well understood than moisture, but potentially of equal importance. In physical terms, temperature can control the amount of moisture that is available either through evaporation or freezing. In reality, it is almost impossible to isolate the effects of temperature on an organism in its natural habitat. However, in laboratory experiments it has been noted that raised temperatures increase the rate of photosynthesis and respiration in most lichen species, up to a certain point. Coxson and Kershaw (1983a) reported a notable increase in nitrogen fixation (a respiration indicator) at between 5 and 6°C, with optimum conditions for photosynthesis in sub-arctic *Peltigera canina* at around 16°C. After these optimum conditions were reached, nitrogenase activity declined – eventually to zero – at a temperature of around 35°C. In a similar study, Coxson and Kershaw (1983b) found maximum rates of photosynthesis in *Rhizocarpon superficiale* at temperatures of between 0 and 10°C. These observations are in agreement with other studies (Ried, 1960; Lange, 1965; Kershaw, 1985) and show that certain species, at least, show some degree of temperature dependence.

Epilithic saxicolous lichens, those that grow on rock surfaces, are subject to much greater extremes of temperature than those growing on materials with lower heat capacities. For this reason, saxicolous species must be able to survive both very high and sub-freezing temperatures with a minimum of stress. Kershaw (1983) found rock-dwelling lichens frequently experienced temperatures 20°C in excess of air temperatures (1-2 m above ground) in high latitude regions. Lange (1953), subsequently followed by Kappen (1973), examined the ability of lichens to tolerate extremes of cold and heat. Kappen (1973) found that in a dry state many species from exposed habitats showed heat resistance up to 90-100°C, however this tolerance declined by 40-50°C when the thalli were wet. Kershaw (1985) demonstrated the insensitivity to heat stresses by exposing lichens to daily (12 hour) heat treatments of

up to 45°C, over a period of several weeks. Photosynthesis was reduced at high temperatures yet respiration rates remained generally unaffected.

Those lichens inhabiting polar regions (>66°N or S) must endure extremely low temperatures, combined with low light levels, for several months of the year. If dried and stored in the dark, certain polar species can tolerate temperatures as low as -196 °C for a few days (Kappen, 1973) and -60°C for a few years (Larson, 1978). In many cases, when gradually thawed, these specimens could resume normal photosynthesis within less than 24 hours (Kappen, 1973). Nash *et al.* (1987) examined the tolerance of the algal component within lichens to low temperatures. They found that *Trentopohlia* photobionts were less resistant to frost than species containing the alga *Trebouxia*, such as *Rhizocarpon geographicum* (L.) DC. The same levels of cold tolerance normally cannot be achieved by low latitude or warmth-adapted lichen species (Larson, 1983).

Although relatively few field studies on the thermal operating environment of lichens have been carried out, laboratory experiments have shown that overall, within certain optimum thresholds, respiration levels and enzyme activity increase with increasing temperature. It can be seen that lichens from different habitats exhibit different temperature optima. Maximum photosynthetic activity generally takes place between temperatures of about 5-20°C, although certain Polar species have adapted to operate most efficiently between 0 and 10°C (Kappen *et al.*, 1996).

Wind strength

Wind affects lichen growth for two different reasons: ventilation and mechanical abrasion. Exposed localities will experience higher wind speeds and are therefore better suited for more hardy plants. The time period that a lichen thallus remains moist after a rain shower will be determined by evaporative processes, which are greatly increased by elevated wind speeds. Exposed, well-ventilated lichens will experience shorter periods of moisture availability for photosynthesis and assimilation. The particular micro-climate (ie. light, wind speed, moisture, temperature, snowcover, etc.) of any individual specimen is governed fundamentally by its micro-habitat, making the specific affect of various meteorological variables extremely difficult to quantify.

The problem of wind abrasion is normally confined to the highest latitudes and altitudes. High-speed winds armed with fine-grained sediment or ice crystals can cause considerable stress to saxicolous lichens in unsheltered locations. Thus colonising lichens will automatically favour less windswept aspects and elevations (Benedict, 1967; Hooker, 1980). Evidence of actual lichen erosion by wind abrasion has, however, presently only been reported from Antactica (Fristrup, 1951; Juckes, 1969).

Snowcover

The effects of snowcover on lichen growth have been examined by numerous workers from both the biological and earth sciences. The issue is a complex one and has given rise to many, apparently contradictory, findings. The most widely held view is that photosynthetic activity and, hence, lichen growth is reduced, or even prevented, by burial under snowcover for prolonged periods (eg. Beschel, 1961; Rydzak, 1961; Reger and Péwé, 1969, Pitman, 1973; Flock, 1974; Crittenden and Kershaw, 1979; Koerner, 1980; MacFarlane and Kershaw, 1980; Porter, 1981; Innes, 1985; Martin *et al.*, 1994). Some workers have proposed that persistent snowbeds can completely kill-off the saxicolous lichen flora (Ives, 1962; Andrews *et al.*, 1976; Locke and Locke, 1977; Haeberli *et al.*, 1979; Karlén, 1979; Caseldine and Baker, 1998; Benedict 1990b, 1991, 1993, 1999). Lichen-free areas in certain periglacial regions are, therefore, said to be the result of 'snow kill' during past periods of climatic deterioration. Various workers, on the other hand, find no evidence of a correlation between snowcover duration and radial growth rates (Karenlampi, 1971; Hooker, 1980; Benedict, 1990a; Innes, 1985), stating instead that other environmental variables are probably more important for growth. Several authors have also argued the beneficial effects of snowcover in protecting lichens from severe weather (Lindsay, 1973; Hooker, 1980; Kappen, 1988) and even promoting growth in some cases (Larsen and Kershaw, 1975; Matthews, 1977; Haines-Young, 1983). It is clear that there is no consensus on the subject of lichen growth and snow cover.

Lichen zonation around semi-permanent snow patches was first noticed by Paterson (1951), although Darbishire (1905) found Antarctic lichens were only checked by the persistent snows and not the extremely low temperatures. Lichen size, abundance and diversity all increase with distance from the centre of a snow-patch which

Pitman (1973) attributed to the increased length of the growing season at the snow-patch margins. A physiological study by Crittenden and Kershaw (1979) showed that chemical reactions (nitrogenase fixation) were negligible in snow-covered lichens whilst, the following year, MacFarlane and Kershaw (1980) found photosynthesis effectively ceased at the low light levels typical of sub-nival conditions. These studies have been used by lichenometrists to validate their findings, whereby the suggestion is that the speed of lichen growth is related to the duration of snowcover. The problem is, however, somewhat more complicated.

Firstly, snow-depth is an important factor. The degree of metabolic activity in lichens is dependent on the amount of light penetrating the snowpack, which is proportional to snow depth. Kappen and Breuer (1991) have shown that photosynthesis can readily take place beneath snow depths of 15 cm (light intensity = 280μ.mol photons m^2s^{-1}). More recent measurements indicate that lichens are even able to photosynthesise at sub-zero temperatures beneath a snowcover, whilst apparently in a frozen state (Kappen, 1993). Conditions beneath a snow covering are normally both warmer and wetter than the surrounding air, which may allow certain lichen species to continue growing during cold, dry months (Benedict, 1990b). However, darkness, rather than freezing temperatures or moisture levels, will ultimately prevent photosynthesis beneath a deep (>50 cm) snowcover. Benedict (1990b) has even suggested that respiration can occur beneath deep snow, due to the presence of meltwater, but that photosynthesis will not proceed – resulting in lichen death if conditions are maintained for more than 40 weeks. In reality, it is extremely difficult to determine the exact role of snow due to the varying nature of the snowcover itself. Snow depth in the field will rarely be uniform and moisture, light and temperature conditions will all vary as a function of the time of day and the time of year.

Where lichen 'snow kill' has been proposed, various thresholds have been cited. Haeberli *et al.* (1979) claim that 12 snow-free weeks a year are required for '*Rhizocarpon geographicum* agg.' to survive in the Swiss Alps, which is very close to Benedict's (1990b) figure of 9-12 weeks derived in the Front Range of Colorado, USA. Benedict (1990b) also states that lichen death will occur within 5 to 8 years under such conditions, which would explain the presence of lichen-free areas dating from the 1950s and from previous centuries when snowcover was apparently more extensive than today. Innes (1985) questions the occurrence of 'snow kill' over

shorter timescales suggesting instead that more extensive snowcover – during the 'Little Ice Age' for example – probably resulted in a period of reduced lichen growth and an absence of colonisation. Kappen (1988) also attributes lichen-free areas to a lack of lichen colonisation rather than lichen eradication during severe climatic spells. Presently, no studies have found direct evidence of such a phenomenon. Furthermore, Kappen (1988) states that hardy, alpine lichens can survive in a dry, frozen state beneath snow for long periods, as shown by laboratory experiments (Nash *et al.*, 1987). The observation of lichen thalli on bedrock emerging from beneath receding glacier ice would tend to support such a view (Koerner, 1980).

Lichenologists have long been aware that certain *cyanophilic* lichens prefer to reside under the relative protection of persistent snowbeds (eg. Lynge, 1937, Gilbert, 1984; Gilbert *et al.*, 1992), particularly in alpine areas and around the margins of Greenland and Antarctica. In cold arid areas, melting snow may provide an important source of moisture (Bliss, 1956; Llano, 1959; Lindsay, 1977). Certain alpine crustose and foliose species are especially adapted for operating under seasonal snowpatches, responding rapidly when the melt season arrives (Ahmadjian, 1970; Crittenden and Kershaw, 1979; Kershaw, 1983; Crittenden, 1998). Matthews (1977) suggested that the increased moisture levels at the foot of many moraine slopes in southern Norway, resulting from longer-lying snow and lower windspeeds, may account for the apparent existence of larger and more numerous yellow-green *Rhizocarpon* thalli. This 'green zone', also identified by Haines-Young (1983), may be a result of lichens responding to micro-environmental change over time. The recognition of such lichen zones on moraines in other areas is not common in the scientific literature.

The problems associated with snowcover in relation to lichen growth are clearly complex and need to be examined further, by both lichenologists and earth scientists, before a clearer picture can be gained.

Nutrient supply

Lichens, unlike most other plants, can thrive in nutrient-deficient habitats. Normally, epilithic, saxicolous lichens extract mineral nutrients from their substrates by means of acids excreted from the fungal cortex (Ahmadjian, 1967). It is a slow process with the type and amount of mineral uptake depending on the composition of the rock on which it is growing. Artificially increasing the nutrient supply should have a positive

effect on lichen growth. However laboratory and field experiments have shown that the relationship is not a simple one. Hakulinen (1966) subjected foliose species growing in the field to a solution of bird excrement over an entire summer. He found that growth was noticeably greater in the nutrient-enriched plants than in those not receiving the solution. Unfortunately, Hakulinen was unable to state conclusively if the abnormal growth was due to the nutrient enrichment or merely resulted from added moisture. Jones and Platt (1969) performed the same experiment in the laboratory and found that nutrient-enriched lichens did grow slightly faster than those receiving only distilled water. Armstrong's (1984) results were less emphatic showing that two out of the five saxicolous species tested showed increased growth in response to bird droppings whilst two remained unaffected and one species showed a negative response. A more recent experiment (Armstrong, 1994) suggests that only certain *ornithocoprophlilous* (guano-loving) species show accelerated growth in response to the addition of bird droppings. This agrees well with the distribution of lichens in the field. The present author has noticed, however, that stony bird perches in Iceland, demarcated by the concentration of droppings, are often lichen-free, possibly due to the scratching of the rock surface by birds' feet.

Pollution

Lichens are sensitive to air pollution. The effect of industry on the lichen flora of many English and German cities was noticed as early as the 1850s. Certain lichen species have been used as reliable pollution monitors and the literature on this subject is extensive (eg. Ferry *et al.*, 1973; Nash and Gries, 1991). Lichenometrists need rarely consider the effects of pollution as the majority of their studies are conducted in clean air environments far away from urban areas (eg. Norway, Iceland, Rocky Mountains, Alaska, New Zealand, etc). Workers need to be aware, however, of the localised effects on lichen growth caused by industry and make note of any potential sources of pollution likely to affect the study area.

Continentality

The *continentality* of the climate, or the lack of oceanic influence, is reflected in a variety of the meteorological phenomena already mentioned (eg. rainfall, temperature range, humidity, cloudiness, etc.) and is not simple to define. Climatologists have used various indices to calculate the continentality of a region's

climate (Barry and Chorley, 1992). Generally, the continentality of an area's climate increases with distance from the open sea. Proximity to the ocean results in more equable, year-round air temperatures, and a generally wetter climate with precipitation distributed evenly throughout the year. The degree of continentality is, however, difficult to express quantitatively. Gams (1932) coined the concept of 'hygric continentality', which he defined as the quotient of mean annual precipitation (mm) and altitude (m). Beschel (1958, 1961), under the instruction of Professor Gams, studied *Rhizocarpon* growth rates in Greenland using the *indirect* approach. He found that there was a direct relationship between the '*lichen factor*' (maximum diameter of thalli growing on a 100-yr old surface) and the continentality of the climate in which they were growing. Beschel states (1961: 1048), in reference to lichen growth, that:

> "The effect of the climate is of paramount importance and the difference in the constant speed of diameter increase between a humid and an arid region may be a factor of twenty."

Re-measurements of lichen thalli carried out in 1970, 12 years after Beschel's original study, confirmed that growth of *Rhizocarpon* was indeed slower in the sites with a more continental climate (6-7 mm/100yr cf. 17-18 mm/100yr). These findings led Ten Brink (1973: 330) to reiterate Beschel's assertions, in that: "the slower rate of lichen growth...is a direct function of the site's very continental climate". The relationship is illustrated in Figure 5.3, taken from Beschel's, now classic, study of 1961.

Conclusions

Studies of lichens in their natural habitats have shown that wetter, more humid periods within the optimum temperature range generally yield the highest growth rates. Furthermore, prolonged snow cover and extremes of temperature, both high and low, are detrimental to lichen growth. No single climatic variable can be said to influence lichen growth, although metabolic activity can only occur when moisture is present. Separation of one variable from another in the field is almost impossible. Rather, a multi-variable parameter such as 'continentality' or 'oceanicity' correlates best with growth and presents a promising way forward in future studies examining the relationship between climate and lichen growth.

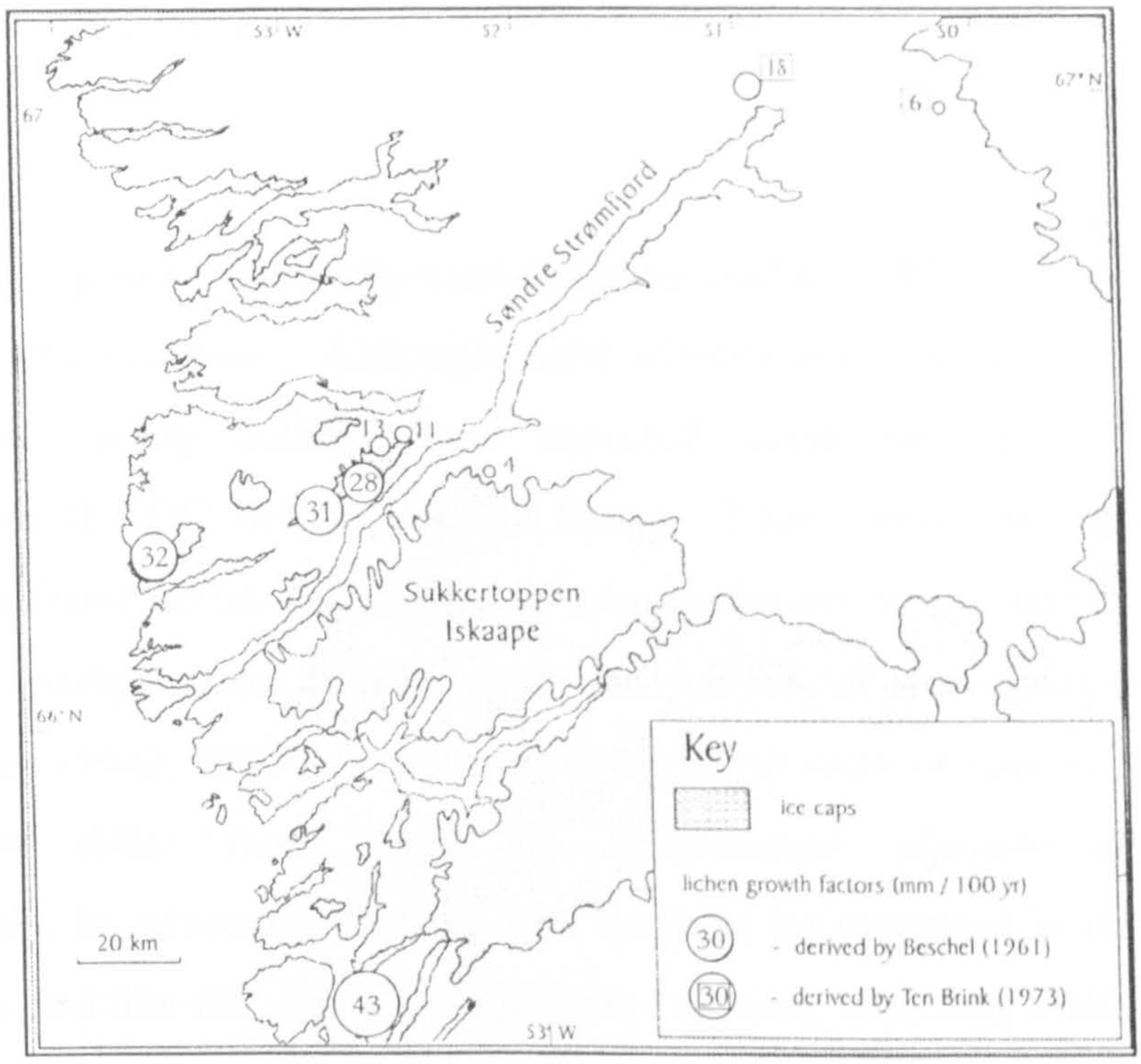
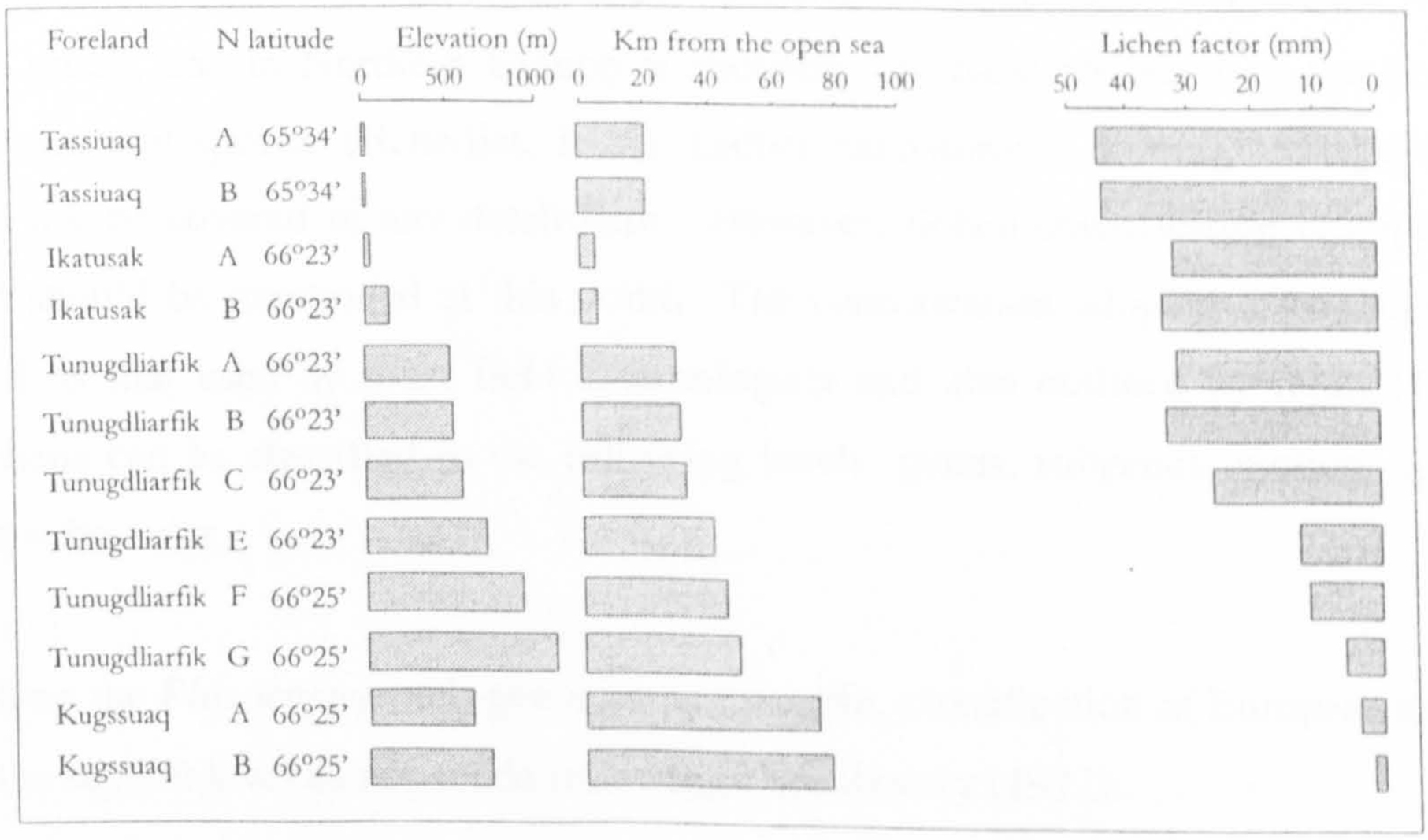


Figure 5.3 Lichen factors (ie. the maximum diameter in millimetres of thalli growing on 100-yr old surfaces) for *Rhizocarpon geographicum* in West Greenland.

The map (above) is based on work undertaken by Beschel (1958, 1961) and supplemented by Ten Brink (1973). Note the decrease in lichen factor with distance from the open ocean.



(Above) The *lichen factor* in relation to continentality in West Greenland. Redrawn directly from Figure 1 in Beschel (1961: 1048).

Taxonomy

The lichen group most frequently used in dating studies is *Rhizocarpon* Ram. em Th. Fr. subgen. *Rhizocarpon*. Although field identification is rarely possible to the species level, many authors have reported using the species *Rhizocarpon geographicum* (L.) DC. in lichenometric studies. More commonly, the lichen species chosen is referred to as '*Rhizocarpon geographicum agg.*', signifying that it is actually an amalgamation of more than one species, or sometimes more than one group. Innes (1982, 1983a) has shown that growth rates of species within Section *Rhizocarpon* differ from those of *Rhizocarpon alpicola* (Hepp.) Rabh. Consequently, he advocated that the two Sections be separated in dating studies to avoid errors and that the use of an 'aggregate' species, including thalli from different Sections, should be discouraged.

This section of the review deals primarily with the growth curve of lichen species within *Rhizocarpon* Section *Rhizocarpon* (formerly known as the *Geographicum* group (Runemark, 1956)) as classified by Cernohorsky (1977), and adopted by Innes (1982, et seq.). The species *Rhizocarpon geographicum* (L.) DC. is included within this group, and in Northern Europe is probably the most abundant of the Section *Rhizocarpon* species (Benedict, 1988). Lichen taxonomy is an enormous field and will not be covered in any detail here. However, lichen classification is important and should be mentioned at this point. The classification adopted throughout this work is that used by most field lichenologists and also outlined by Innes (1985). Lichens can be classified to the following levels: genus, subgenus, section, species and subspecies.

Taking the *Rhizocarpon* sub-genus as an example, classification of European species to the section level can be made following Cernohorsky (1977):

Section *Superficiale* (Runem.) Thoms.,
Section *Alpicola* (Runem.) Thoms.,
Section *Viridatrum* (Runem.) Thoms.,
Section *Rhizocarpon*

Section *Rhizocarpon* can be classified into the following species:

R. atroflavescens Lynge
R. cerpaticum Runem.
R. ferax H. Magn.
R. furax Poelt et V.Wirth
R. geographicum (L.) DC.
R. intermediellum Ras.
R. lecanarium Anders.
R. macrosporum Ras.
R. pulverulentum (Schaer.) Ras.
R. rapax V.Wirth et Poelt
R. riparium Ras.
R. saanense Ras.
R. sphaeroporum Ras.
R. sublucidum Ras.
R. tavaresii Ras.
R. tineii (Tornab.) Runem. (s.str.)

In terms of lichenometry, identification is realistically only possible to the Section level and, following recommendations by Innes (1985), workers should always specify the level of classification adopted within their study (ie: subgenus, section or species). Unfortunately, many workers (eg. Andrews and Webber, 1964; Matthews, 1974; Evans *et al.*, 1999) have used an aggregate of yellow-green *Rhizocarpon* 'species', often referred to as '*Rhizocarpon geographicum* sensu lato', making comparison with other studies difficult.

Growth rate studies

Direct method

In August 2000, only eight papers report growth measurements over a period of 12 months or more for species within Section *Rhizocarpon* :– Hausmann (1948), Beschel and Ten Brink (1973), Hooker (1980), Armstrong (1983), Proctor (1983), Rogerson *et al.*, (1986), Haworth *et al.*, (1986), Matthews (1994). With the exception of Matthews's study, all the measurements refer to thalli of '*Rhizocarpon geographicum*'. The existing studies will now be critically analysed in turn.

Ethel Hausmann (1948) was probably the first to publish information regarding the growth rate of '*Rhizocarpon geographicum*'. She did so in a two-page article in the *Torrey Botanical Club Bulletin*. Her observations of a single *R. geographicum* 'map lichen' growing near the summit of Mt. Monadnock, New Hampshire, showed that the radial growth rate of *R. geographicum* over a four-year period was slow, typically around 0.4 mm per year. She also made a basic physiological assertion that the rock-lichens in question grew fastest in the wettest year when the mountain was "more swathed in clouds" than normal.

Roland Beschel (1956, 1961) measured *Rhizocarpon* lichens in west Greenland. Ten Brink (1973) re-visited Beschel's control sites in west Greenland and determined the growth rate of several *Rhizocarpon geographicum* thalli over the intervening 15-year period. The results show that radial growth had been very slow, averaging 0.05 - 0.1 mm per year. No conclusions were made regarding the growth history of the plants.

Antarctic crustose lichens are presumed to be amongst the slowest-growing and, hence, longest-living organisms on Earth. Hooker's studies on the South Orkney Islands between 1972 and 1977 were published in 1980. He detected no measurable growth during a three-year period in any of the 63 *Rhizocarpon geographicum* thalli being monitored. However, after a subsequent three-year period, small amounts of radial growth were recorded, optimally around 0.1 mm per year. Using this data he estimated average radial growth rates in the maritime sub-Antarctic to be of the order of 4 mm per century. Hooker (1980) was unable to demonstrate any significant effect of size (age) on thallus growth rate.

Armstrong's (1983) observations on the growth of *Rhizocarpon geographicum* in North Wales comprise probably the most detailed study of this species so far conducted. He measured the radial growth of 39 thalli over a period of 1.5 years (Fig. 5.4). Armstrong's results detected three distinct phases of growth – an initial phase where the growth rate increases rapidly, a second phase in which the growth rate is high and relatively constant and a declining phase where growth rates fall. The statistical representation of this growth rate decline has been questioned (Innes, 1985). However, the radial growth rate of each of the thalli over 55 mm in diameter falls below the mean growth rate of thalli between 15 and 55 mm in diameter.

Armstrong's evidence suggests that a phase of declining growth occurs in thalli over 45-55 mm in diameter and hence that growth is a size (age) dependent phenomenon in *Rhizocarpon geographicum*.

Proctor (1983) studied the growth rates of thalli growing on the moraines of an Alpine glacier in Switzerland. He took photographs of numerous sites at four-year intervals and using these calculated the growth rates of 22 *Rhizocarpon geographicum* thalli. He then constructed a curve of mean radial growth increment against thallus radius showing that smaller (< c.10 mm) thalli grew significantly more slowly than larger thalli. Unfortunately, Proctor's dataset does not include any thalli of more than 36 mm in diameter. However, Proctor assumes constant growth rates in thalli over c.10 mm in diameter and extrapolates the linear part of his growth curve for thalli up to 60 mm in diameter, without any further measurements. The statement he then makes regarding the growth curve of *Rhizocarpon geographicum* could be seen as somewhat over-optimistic: "That lichen thalli can grow with a near-constant radial increment for periods of many years is now established beyond question" (1983: 258).

Rogerson *et al.* (1986) measured 7 thalli of *Rhizocarpon* Section *Rhizocarpon geographicum* growing on rocks in the coastal mountains of northern Labrador in 1978 and again in 1983. They derived diameter increases of 0.10 – 0.58 mm per year. An average diameter increase of 0.34 mm per year, over the five-year period, is equivalent to a radial growth rate of 0.17 mm per year. However, due to the small number of thalli measured, including only two greater than 50 mm in diameter, the study of Rogerson *et al.* sheds little new light on the relationship between thallus size and growth rate.

Haworth *et al.* (1986) measured 92 *Rhizocarpon geographicum* 'sensu lato' thalli over a 4-6-year period in the Central Brooks Range, Alaska (Fig. 5.4). Absolute radial growth rates ranged from 0.18 to 0 mm per year with variability being high. Their study, like Armstrong's (1983), also found evidence, though poorly defined, for an inverse relationship between thallus growth rate and thallus size. They reach the following conclusion (1986: 294):

"The growth curve developed for *Rhizocarpon geographicum* shows

continuously slowing increases in diameter throughout the life of the thallus."

They also state that lichen age is only one of many factors that will contribute to variations in growth rate, along with substrate lithology, micro-climate and competition between thalli.

Matthews (1994) presents direct measurements of 63 thalli of the *Rhizocarpon* subgenus (probably only species within Sections *Rhizocarpon* and *Alpicola*, (Matthews, *pers.comm*. May 2000) (Fig. 5.4). His measurements were conducted over a five-year interval on the outermost moraine at Nigardsbreen in southern Norway. Matthews finds no suggestion of a decline in growth rate with increasing thallus size. However, the extreme scatter within the dataset precludes the identification of any possible trends. Furthermore, the use of an "aggregate species" may cause any growth rate decline in *Rhizocarpon geographicum* to be hidden amongst data from species within different lichen groups (ie. *Rhizocarpon alpicola*; (see Innes, 1983a, 1985)). The apparently linear growth rate in *Rhizocarpon* thalli up to 120 mm in diameter should therefore be viewed with caution.

Although currently unpublished, research by Vanessa Winchester (*pers.comm*; August 2000) may prove very valuable. Winchester has recently measured thalli of *Rhizocarpon geographicum* on gravestones in North Wales. She deduced that maximum radial growth rates were approximately 1.4 mm/yr, over the four-year period (original measurements made in 1996). These findings represent the highest growth rates so far recorded for this species.

Indirect method

Earth scientists, interested in dating rock surfaces have contributed valuable information on lichen growth rates. Measuring the size of lichen colonies on surfaces of known age offers a way of studying the growth rates of slow-growing lichens over periods of many decades.

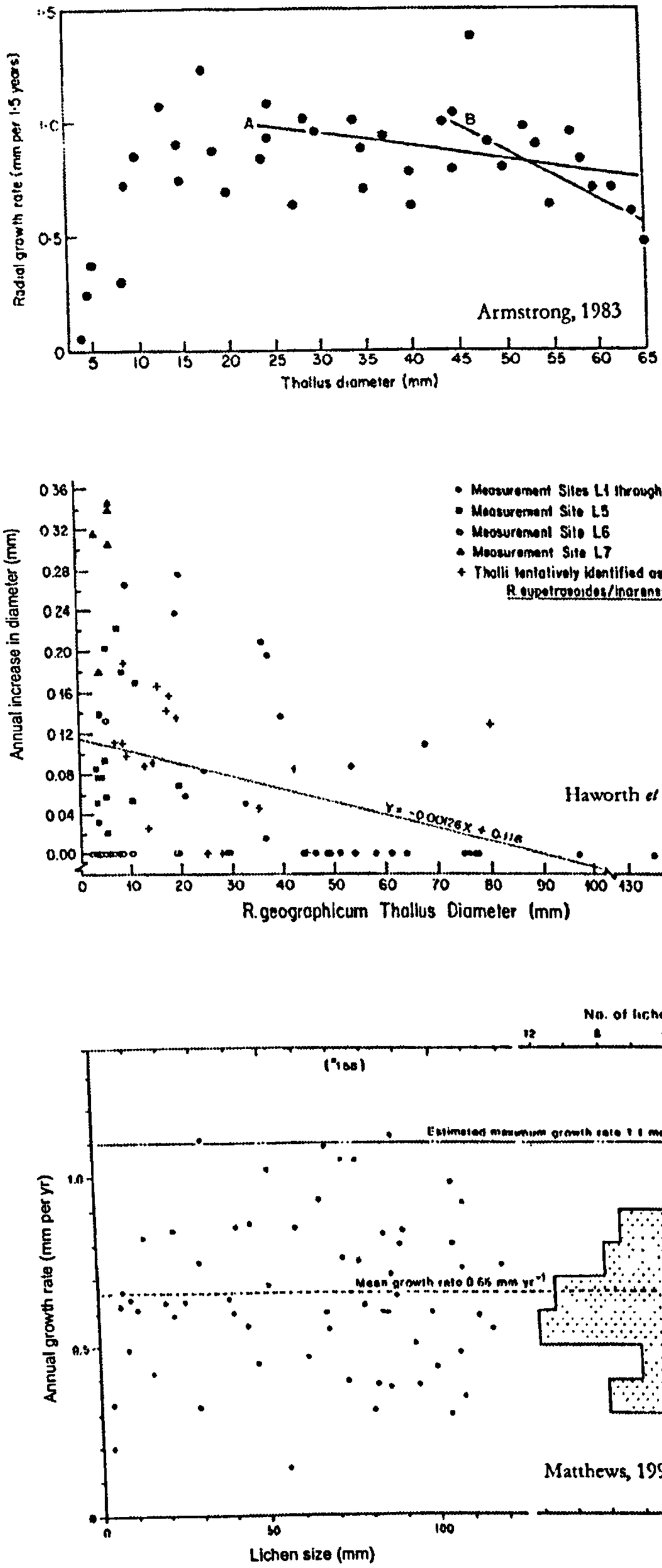


Figure 5.4 Directly-measured growth rates of *Rhizocarpon geographicum* thalli.

Note that the vertical axis is different in each study. Armstong (uppermost graph) measured radial growth (expressed in mm per 1.5 years); Haworth *et al.* (middle graph) and Matthews (lower graph) both measured diameter increase (in mm per year). Matthews's measurements were made on thalli from Section *Rhizocarpon* and Section *Alpicola* (Matthews, *pers. comm.*, 2000).

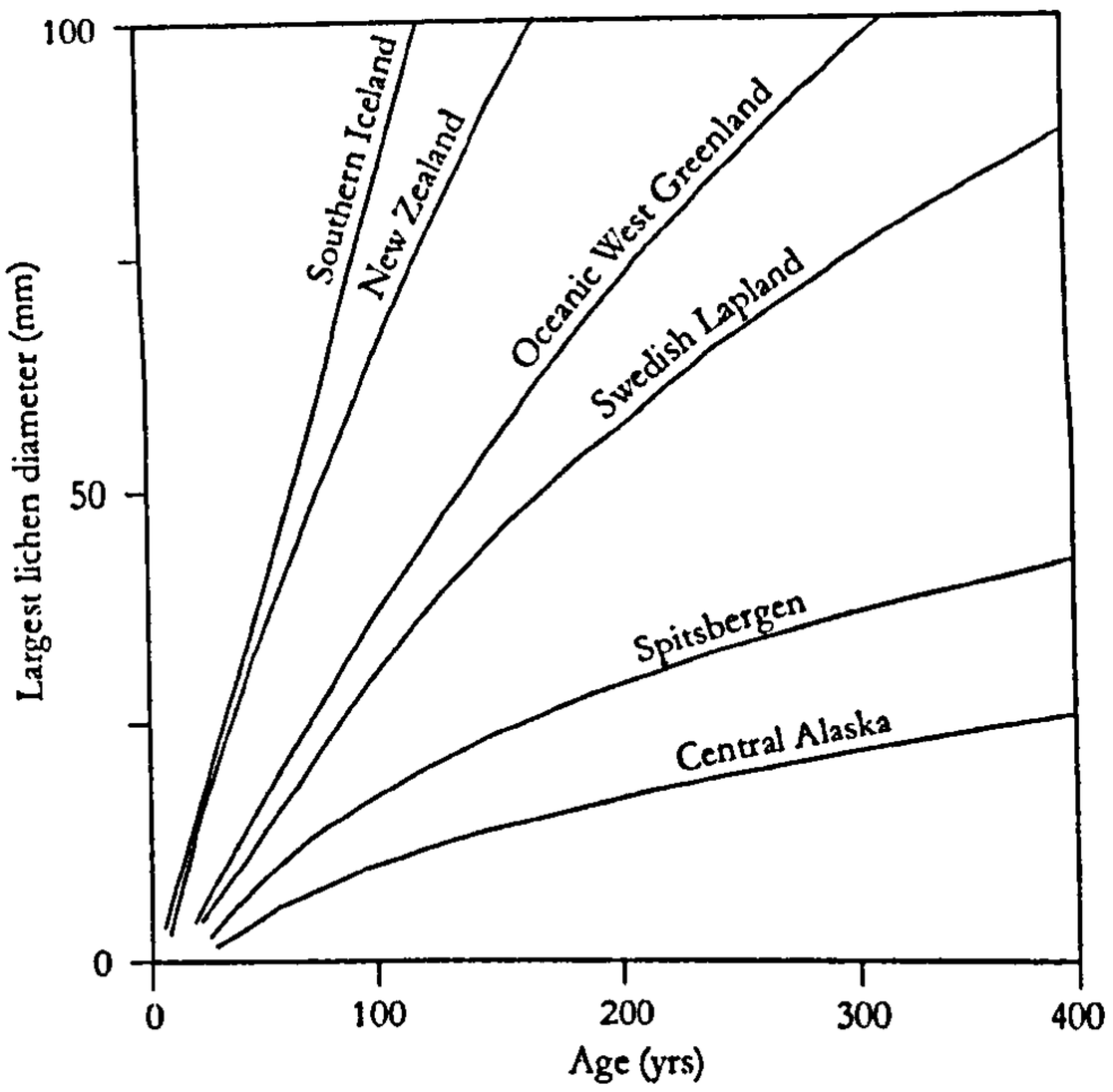


Figure 5.5 (above) Some curvilinear *Rhizocarpon geographicum* lichen dating curves.

Central Alaska - (Calkin and Ellis, 1980)
Spitsbergen - (Werner, 1990)
Swedish Lapland - (Denton and Karlen, 1973)
Oceanic West Greenland - (Beschel, 1961)
New Zealand - (Burrows and Lucas, 1967)
Iceland - (Evans *et al.*, 1999)

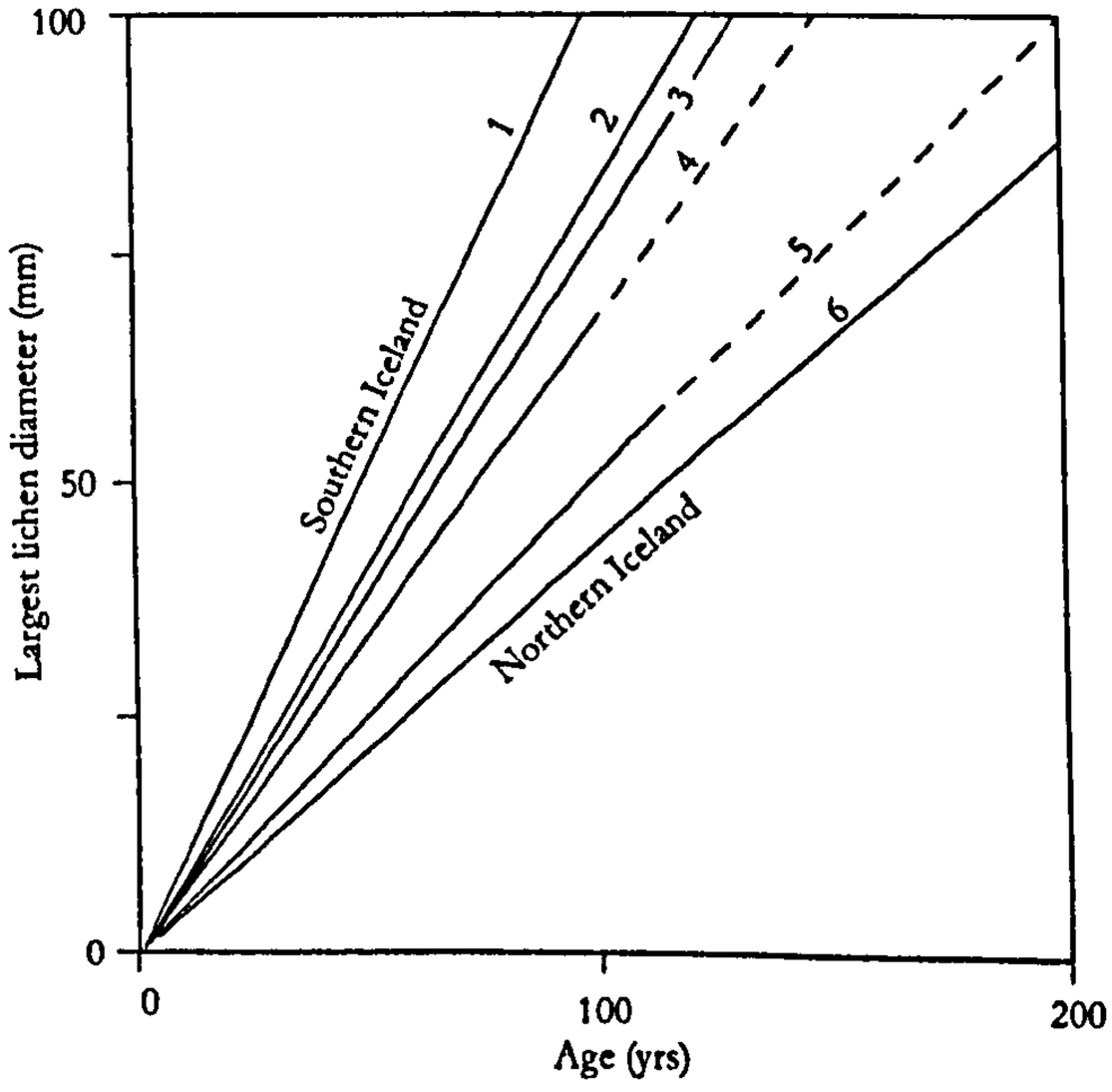


Figure 5.6 (above) Linear *Rhizocarpon geographicum* lichen dating 'curves' from Iceland. All have been normalised to remove any differences in lag-time. Dashed sections represent extrapolations beyond those published by the author(s).

1 - Breiðamerkurjökull (Gordon and Sharp, 1983)
2 - Suðursveit (Evans *et al.*, 1999)
3 - Sólheimajökull (Maizels and Dugmore, 1985)
4 - Eiríksjökull (Gudmundsson, 1998b)
5 - Brúárjökull (Evans *et al.*, 1999)
6 - Skiðadalur (Kugelmann, 1991)

Numerous authors have derived non-linear dating curves, using *Rhizocarpon* species, which 'flatten-off' over time (Fig. 5.5) (Beschel, 1958; Bornefeldt and Osterborg 1958; Miller, C.D., 1969; Miller, G.H. and Andrews, 1972; Mottershead and White, 1972; Denton and Karlen, 1973; Matthews, 1974; Innes, 1985; Benedict, 1985). Such curvilinear forms imply that lichen growth gradually slows down, probably as a function of thallus size. Those workers who have reportedly found a linear relationship between the age and size of *Rhizocarpon* lichens (eg. Burrows, 1975; Gordon and Sharp, 1983; Maizels and Dugmore, 1985; Evans *et al.*, 1999) over one or two centuries, have probably identified only part of a larger curve (Fig. 5.6).

Conclusions

Eight studies have published directly measured growth rates of *Rhizocarpon* lichens. Of these, five have examined the role of thallus size on growth rate. Two of these studies (Armstrong, 1983; Haworth *et al.*, 1986) show evidence for a decline in the radial growth rate of large thalli, whilst one does not include large thalli (Proctor, 1983), one is inconclusive (Rogerson *et al.*, 1986), and a fifth study shows no evidence for growth rate decline (Matthews, 1994). On balance, these studies appear to suggest that larger '*Rhizocarpon geographicum*' thalli grow more slowly than thalli of lesser dimensions. Some uncertainty still surrounds the exact shape of the growth curve, but recent measurements of other crustose species (Armstrong, 2000 *unpub.*) point towards a gradually decreasing, possibly asymptotic, form (Fig. 5.2).

Indirect lichenometric studies also appear to support a growth rate decline in larger thalli. The relationship between lichen size and age is **not** linear. Both direct and indirect measurements suggest that the radial growth rate of yellow-green *Rhizocarpon* lichens slows with time. Indirect studies still leave question marks surrounding the role played by climatic change and the effects of ecological competition. Further long-term direct observations are required from different climatic regions to show beyond reasonable doubt that age is the primary control on the radial growth rate of lichens within the *Rhizocarpon* family.

Lichenometric dating – sampling techniques

Lichenometric dating has been employed in numerous high altitude and high latitude locations throughout the world. In Iceland, lichenometry has been used extensively to date recently-deposited moraines, (Jaksch, 1970; 1975; Caseldine, 1983, 1991; Gordon and Sharp, 1983; Hjört *et al.*, 1985; Thompson, 1988; Häberle, 1991; Kugelmann, 1991; Stötter, 1991; Bradwell, 1998; Gudmundsson, 1998a, 1998b; Evans *et al.*, 1999; Kirkbride and Dugmore, 2001b), outwash deposits (Thompson and Jones, 1986; Maizels and Dugmore, 1985) and to estimate soil erosion rates (Buckland, 1994). Various different data collection techniques have been employed, including: measurement of the shortest (Thompson and Jones, 1986) or the maximum lichen diameter (Evans *et al.*, 1999), measuring the size of the 5, 10, or 20 largest lichens (Gordon and Sharp, 1983; Maizels and Dugmore, 1985; Evans *et al.*, 1999) or the largest single thallus (Kugelmann, 1991), and searching areas of different size (Thompson and Jones, 1986; cf. Gordon and Sharp, 1983). Caseldine (1991) advocated the use of size-frequency analysis to derive the age of a lichen population, rather than the use of one or a few thalli. All of these sampling strategies will have a major effect on the largest lichen recorded, the reported lichen 'growth' rate and, consequently, the lichenometric age of the surface on which they are growing (Tables 5.2 and 5.3).

Lichens within the slow-growing *Rhizocarpon* subgenus have been used in all the studies conducted in Iceland so far, although Buckland (1994) working in the northwest of Iceland also measured species of *Buellia* and *Placopsis*. Most workers have referred to the lichen being measured as '*Rhizocarpon geographicum* agg.' – an aggregate species including lichens from all four Sections within the *Rhizocarpon* subgenus. There is evidence that species within Sections *Rhizocarpon* and *Alpicola* grow at different rates (Innes, 1982, 1983a).

With the exception of Thompson and Jones's study (1986), all the published dating 'curves' have found a linear relationship between lichen age and lichen size in Iceland. Several dating curves have been extrapolated over the last 200 years or more (Maizels and Dugmore, 1985; Kugelmann, 1991; Häberle, 1991). Diametrical growth rates for '*Rhizocarpon geographicum*' of between 0.44 and 0.98 mm/year have been calculated from these dating 'curves'. However, the true growth rate of

lichens can only be determined by direct measurements carried out over periods of several years (Armstrong, 1974; Innes, 1985), yet, at present, no direct observations of lichen growth in Iceland have been published.

Innes scrutinised the different lichenometric procedures and made many suggestions based on his own findings in Norway and Scotland (1982, 1983a, 1984, 1985, 1986). In Iceland, Kirkbride and Dugmore (2001b) have recently compared results derived by three different lichenometric techniques on surfaces of known age. Their results show inconsistencies between age estimates of landforms derived using lichenometry (*largest lichen* and *1-in-1000* techniques) and those derived using tephrochronology. These findings would appear to undermine the use of traditional lichenometric techniques in Iceland over periods of more than ~100 years. However, the use of a *linear* lichenometric dating 'curve' by Kirkbride and Dugmore (2001b) over the last 200 years goes some way to explaining the apparent contradiction between the dates.

The area searched during a lichenometric survey also affects the size of the 'largest lichen' encountered. Several workers have suggested that the probability of finding the largest lichen on a substrate will increase as the sample area is enlarged (Matthews, 1975, 1977; Innes, 1984; Smirnova and Nikonov, 1990). Innes (1984) found that the largest lichen per quadrat increased with increasing quadrat size but that the rate of this increase decreased in a logarithmic fashion, probably due to the nature of lichen population distributions (Fig. 5.7). The diameter of the 'largest lichen' observed in a 256 m^2 quadrat is likely to be 10-15% greater than the 'largest lichen' in a search area one-quarter the size. Where the whole surface cannot be searched, as is often the case with large surfaces such as river terraces, raised beaches and lava flows, smaller quadrats of a pre-determined size should be used. Innes (1984) recommends a search area of no less than 400 m^2, although McCarthy (1999) has argued that standardisation of the search area between sites is more important than the actual size of the area itself.

In Iceland, lichenometric surveys have been conducted on various scales (Tables 5.2 and 5.3). Whilst Thompson and Jones (1986) preferred to search entire deposits for the largest lichens (the greatest with an area of 30,000 m^2), Sharp chose to sample within three quadrats of 50 m^2 on each surface (Gordon and Sharp, 1983). Although

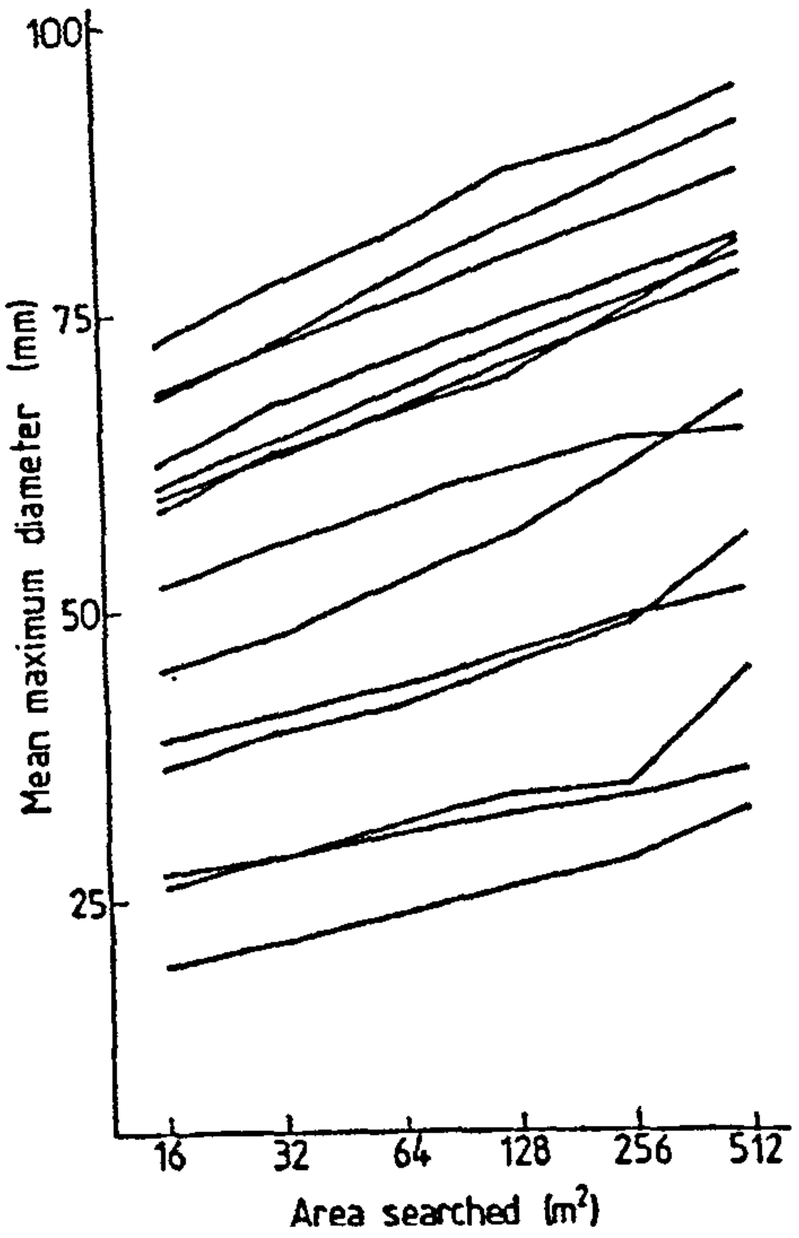
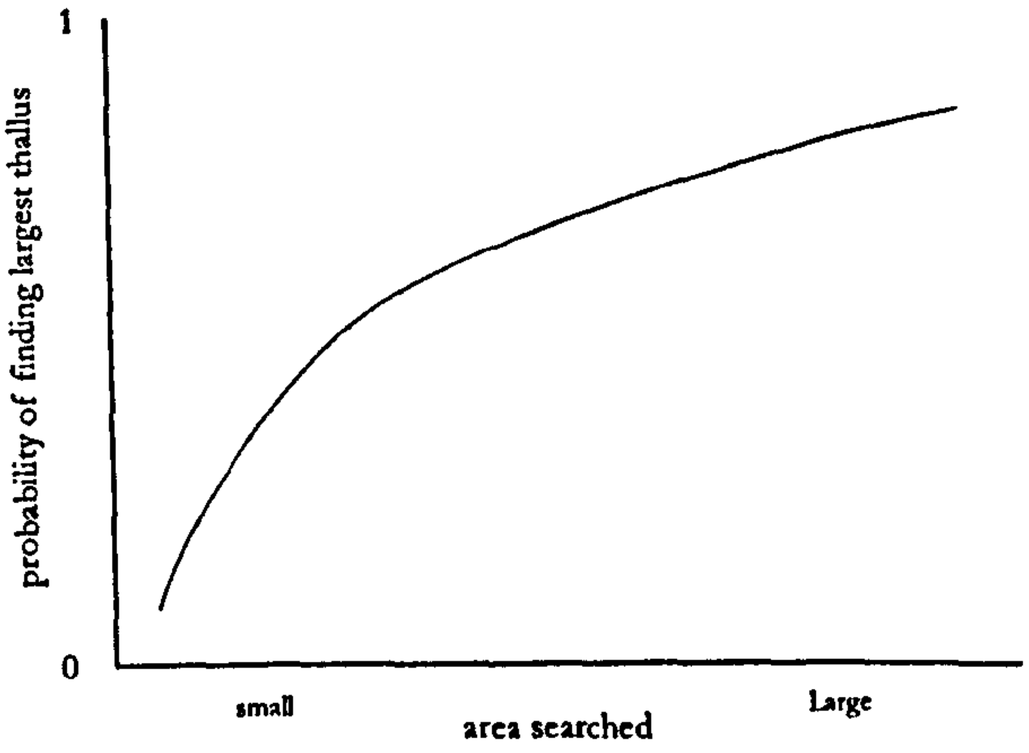


Figure 5.7 Increase in lichen diameter with increasing area searched

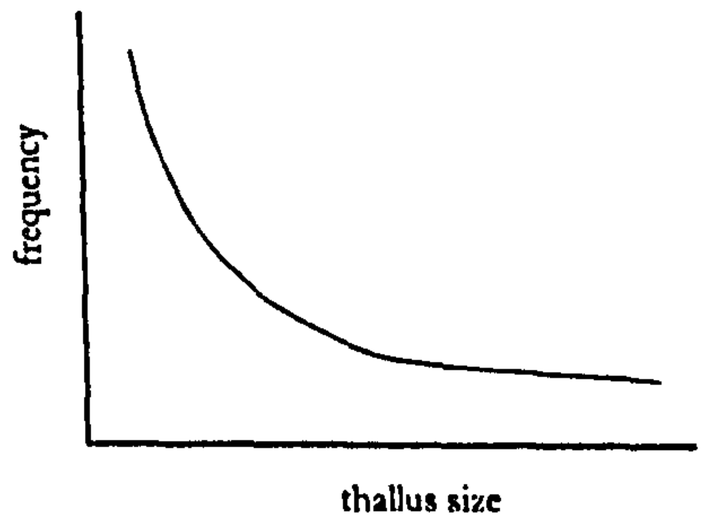
(after Innes, 1984; from studies conducted on moraines in southern Norway).

Note the horizontal scale is non-linear.



(above) Schematic representation of the relationship between sample area and the probability of finding the largest thallus.

As Innes (1984) pointed out, if a log-normal population distribution is assumed (right), the size of the largest thallus encountered increases with area, but the rate of this increase decreases in a logarithmic fashion.



(above) Typical log-normal lichen population distribution – as found by various workers (eg. Benedict, 1967; Andersen and Sollid, 1971; Topham, 1977)

the size of the largest lichen identified and hence the calculated magnitude of lichen 'growth' in each of the two surveys will be different, the form of both lichenometric dating curves should be the same irrespective of sample area (Matthews, 1974; Innes, 1985). Moreover, it is crucial that the technique adopted in the dating-curve calibration procedure matches that used in the lichenometric dating survey.

In summary, the use of the largest or the five largest lichen thalli growing on an entire surface is the most widely adopted lichenometric technique in Iceland and elsewhere (eg. Beschel, 1950, 1961; Denton and Karlén, 1973; Matthews, 1975, 1994; Kugelmann, 1991, Thompson and Jones, 1986; Evans et al, 1999). However, this technique relies heavily on identifying the largest, or several largest, thalli growing on surfaces of various size and age. In many studies this method is impractical and can produce unrepresentative results.

Building on the success of the 'size-frequency' technique, reported by Benedict (1985), Caseldine (1991) and, most recently, Bull and Brandon (1998), it is preferable to measure a large population of lichens on each surface. This technique enables a rigorous, more reliable, age estimate to be made whilst also allowing an assessment of the environmental history of the surface based on the lichen-population structure. To aid replication lichen measurements should involve a fixed sample size and should include every thallus (with the exception of those too small to identify). A detailed study of lichen populations in Iceland will hopefully yield important information on the age of various glacially-deposited features and, critically, also test the applicability of the technique in a maritime, sub-Arctic environment.

Table 5.3 Lichenometric dating studies conducted in Iceland

author(s)	date[1] (AD)	location	lichen species	lichen parameter	no. of lichens measured[2]	survey area (m^2)	calibration surfaces	oldest surface (AD)[3]	lag time (years)	growth rate (mm/yr)
Jaksch	1970	Sólheimajökull	*R.geog.*	long axis	1	entire[4]	gravestones	~1870	15	0.73[5]
Jaksch	1975	Flaájökull	*R.geog.*	long axis	1	entire	moraines	~1890	<15	0.65[5]
Caseldine	1983	Skiðadalur	*R.geog.* agg.	short axis	1, 5	80-690	farmsteads	1905	5-15	0.51
Gordon and Sharp	1980	Breiðamerkur.	*R.geog.* agg	short axis	1	1500	moraines	1894	5	0.675
		Skálafellsjökull	*R.geog.* agg	long axis	1	150	moraines	1887	15	0.769
				short axis	1	150	moraines	1887	15	0.987
Maizels and Dugmore	1983	Sólheimasandur	*R.geog*	long axis	1, 5, 25	200	moraines	1890	25	0.85[5]
Hjört *et al.*	1984	Hornstrandir	*R.geog*	long axis	5	entire	none	see footnote[6]	–	~0.5[6]
Thompson and Jones	1984	Öræfi	*R.geog.* agg.	short axis	5	entire	moraines	1870	–	0.585 - 0.725[7]
Kugelmann	1988	Skiðadalur	*R.geog.* agg.	long axis	1	entire	gravestones	1888	10	0.44
Häberle	1990	Barkárdalur	*R.geog.* agg	short axis	5	entire	gravestones	1880		
Gudmundsson	1991	Eiríksjökull	*R.geog.*	short axis	5	entire	gravestones	1890	19	0.67
Evans *et al.*	1993	Suðursveit	*R.geog.* s.l.	long axis	5	entire	moraines, lake shorelines	1887	6.5	0.80
							gravestones	1930	>5	0.56
Evans *et al.*	1995	Brúárjökull	*R.geog.* s.l.	long axis	5	entire	moraines	1890	16	0.51
Gudmundsson	1995	Öræfi	*R.geog.* agg	short axis	5	entire	see footnote[8]	1889	–	0.73
Bradwell	1996	Virkisjökull	*R.geog.* agg.	long axis	20	entire	moraines	1881	–	0.6[5]

Notes

1 - the year in which the lichenometric survey was carried out
2 - number of 'largest-lichens' averaged to derive age of surface (1 = only largest-lichen used)
3 - the oldest surface used in the construction of the calibration curve
4 - denotes that the entire surface was searched for the largest lichens
5 - these growth rates were originally expressed as 'mm per 100 yr'
6 - Hjört *et al.* (1985) did not produce a calibration curve and merely approximated a likely growth rate for lichens in Hornstrandir based on Caseldine (1983)
7 - temporally-variable growth rate that 'flattens-off' over time (~1930-1870)
8 - Gudmundsson (1998a) combined the growth rates for southeast Iceland derived by Gordon and Sharp (1983) and Thompson and Jones (1986).
R.geog = *Rhizocarpon geographicum* – where no level of classification is made.
R.geog. agg. = *Rhizocarpon geographicum* agg. – an aggregate species comprising all Sections within Subgenus *Rhizocarpon*.
R.geog. s.l. = *R.geographicum* sensu lato – ie. includes all species within Subgenus *Rhizocarpon*.

Table 5.4 Other lichenometric studies conducted in Iceland

author(s)	date[1]	location (AD)	lichen species[2]	technique employed[3]	lichen parameter	sample size or area[4]	calibration surfaces	oldest surface[5]	other comments
Caseldine	1988	Skiðadalur	*R.geog.* s.l.	SF	long axis	1000	debris flows	AD 1880s	Relative age-estimates only. Younger deposits have steeper SF gradients.
P.I. Buckland	1990	Reykjafjörður, Ofeigsfjörður	*R.geog.*	5LL	short axis	20 sq.m	none	–	No growth rates derived.
			Buellia spp.	5LL	short axis	20 sq.m	none	–	Linear growth rates, based on
			Dimelaena spp.	5LL	short axis	20 sq.m	none	–	rate derived for *R.geog.*
			Placopsis selida	5LL	short axis	20 sq.m	none	–	by Caseldine (1990), were adjusted for 3 other species.
Caseldine and Baker	1995	Skiðadalur	*R.geog.* s.l	SF	long axis	500-1000	moraines	>2800 BP	SF distributions used to detect 'snow kill' events. Lichens <10 mm not measured.
Kirkbride and Dugmore	1998	Gigjökull, Steinholtsjökull	*R.geog.*	SF, 5LL, LL	long axis	450-750	moraines, flood deposit, alluvial fans	AD 1721-1755	Assessment of various lichen-dating techniques on surfaces of known age (tephrochronology). Use of Maizels and Dugmore's (1985) dating curve, which assumes constant linear growth.

Notes

1 - the year in which the lichenometric survey was carried out
2 - *R.geog.* s.l. = Rhizocarpon geographicum sensu lato (ie. all species within the Rhizocarpon subgenus)
3 - SF = size-frequency 'population' study; 5LL = mean of the 5-largest lichens; LL = single largest lichen
4 - number of thalli measured (sample size), or area of sample site (m^2)
5 - oldest independently-dated surface used in the study

Chapter 6 : Results

Geomorphology

Introduction

In order to reconstruct the deglacial chronology of Lambatungnajökull it was important, firstly, to map the geomorphology in detail. The resulting map would act as a spatial framework on which to base the chronological investigations.

A geomorphological map was compiled from the most recent stereo aerial photographs of the area. Taken in August 1989, from a flying height of 6000 m, the photographs represent a scale of approximately 1:22,500. The photographic coverage is shown in Figure 6.1. The original prints were enlarged to allow mapping at a scale of approximately 1:11,000. Features of 2 m in length can be identified on these photographs. Steoreoscopic viewing allowed more complex landforms to be interpreted. The context of the detailed geomorphological map (Fig. 6.2, in back pocket) will now be addressed in more detail, focusing on each valley in turn

Glacial landforms in Skyndidalur

Skyndidalur contains numerous landforms of glacial origin. Moraines, constructional features resulting from sediment deposition at the ice-margin, are preserved on the valley floor and along the valley sides. Some of these moraine ridges are discontinuous and, in places, have been dissected by meltwater emanating from the glacier and streams draining from the surrounding hills. The *Skyndidalsá* river, which emerges directly from the glacier, is presently confined to a relatively narrow channel, 10-20 m wide, on the northern half of the valley floor. Many glacial landforms occur on the southern half of the valley floor. Closer to the glacier, meltwater flows in numerous, unconfined, small streams across almost the entire width of the valley. Amongst these outwash deposits

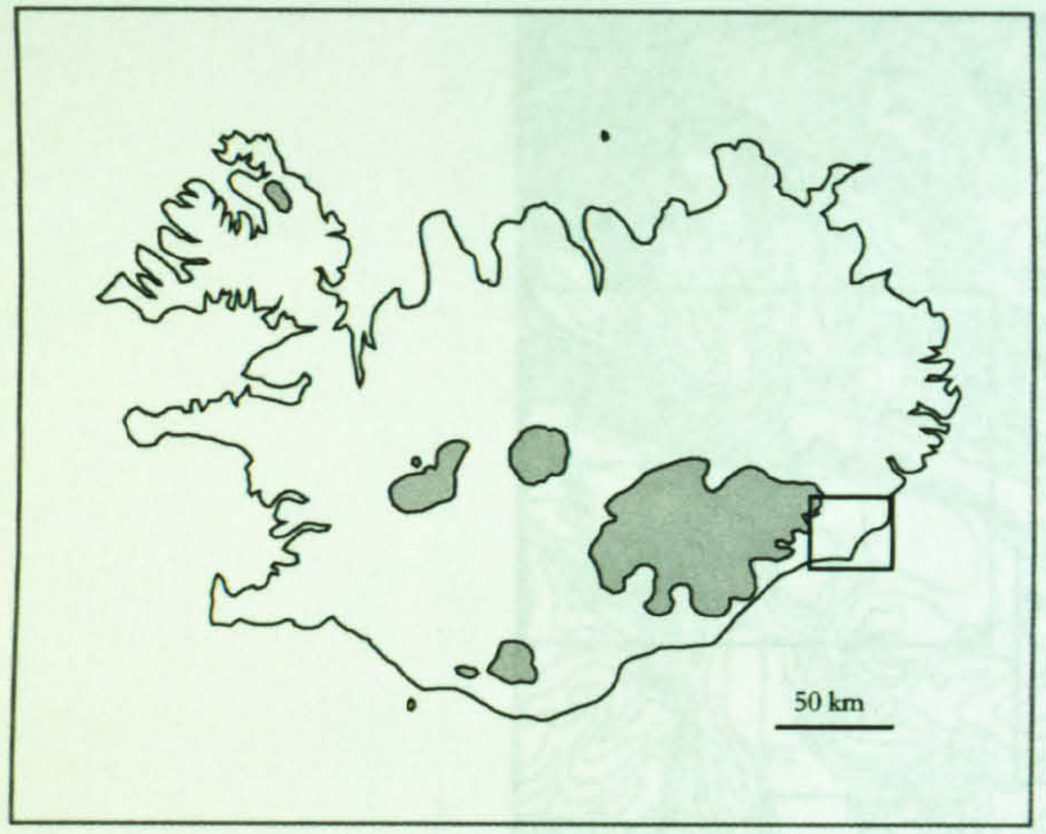
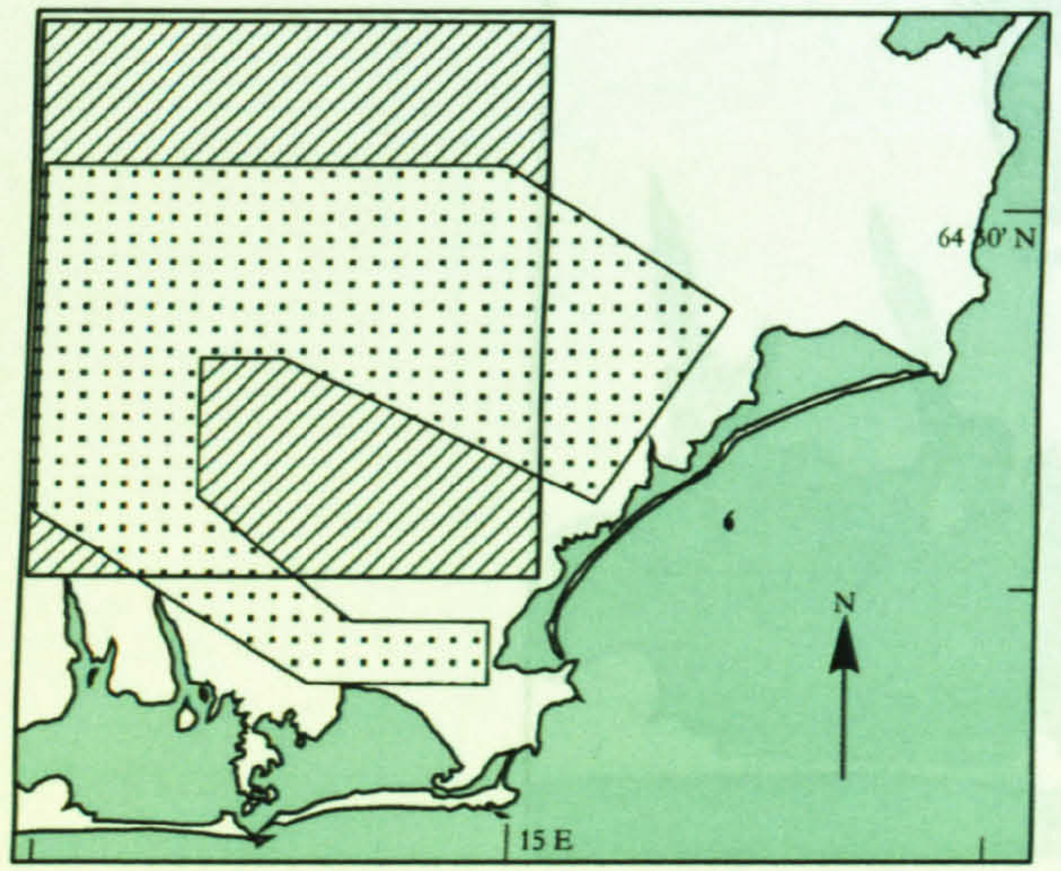
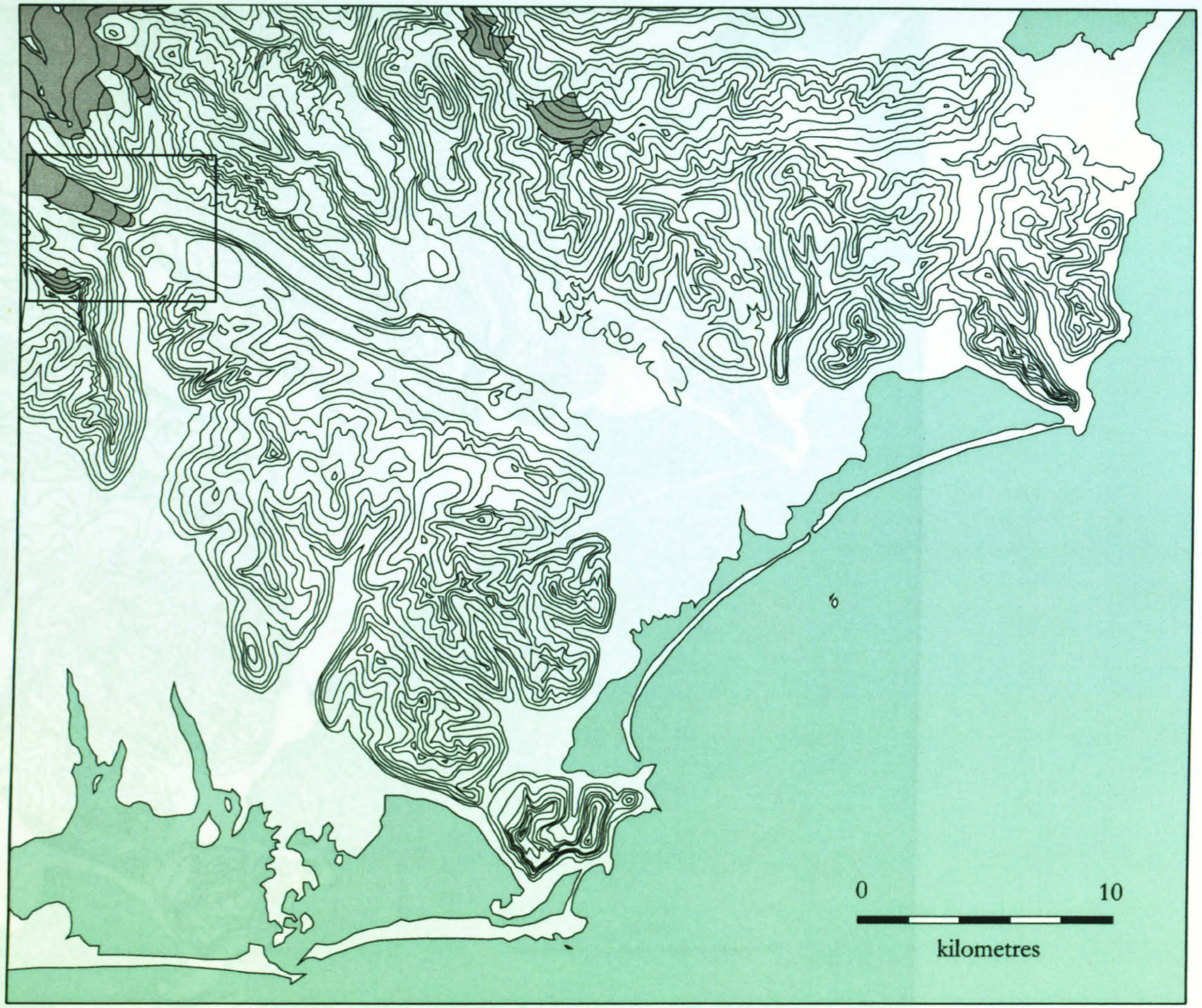


Figure 6.1 Location of the field study area in southeast Iceland (see inset, above). The hatched area (below) represents the terrain mapped from aerial photographs taken in 1982 and 1989. The stippled area was covered on foot and mapped during 1998-9.





(above) The field area stretches from the easternmost flanks of Vatnajökull to the North Atlantic Ocean and covers an area of roughly 400 km^2. Lambatungnajökull – the glacier at the centre of this research – is on the eastern edge of the map. Most of the detailed investigations were conducted in the vicinity of this glacier (within the small rectangle).

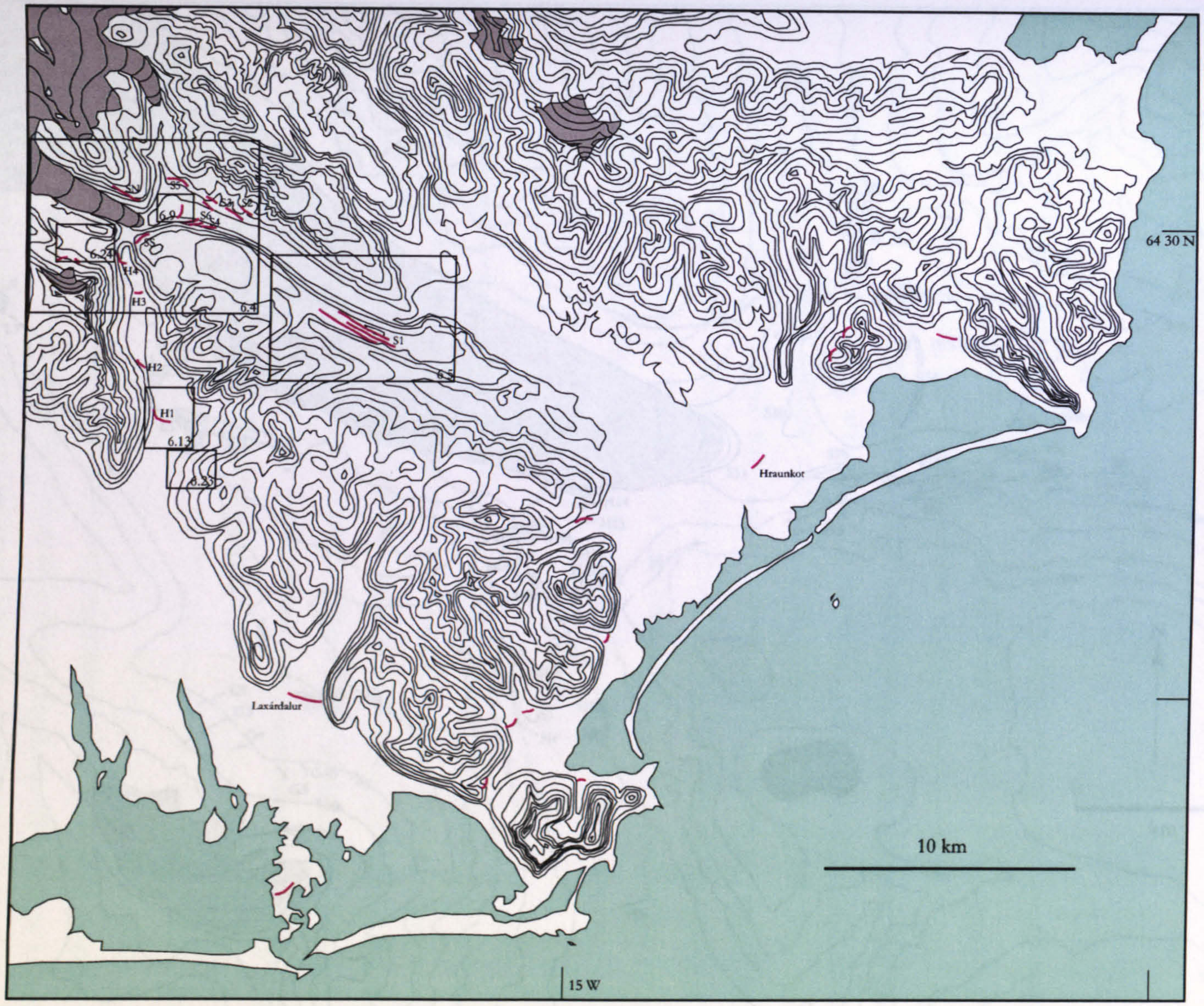


Figure 6.3 Moraines relating to the former extent of glaciers within the study area. Those allocated a number are referred to in more detail in this chapter. The smaller rectangles refer to subsequent figures within this chapter.

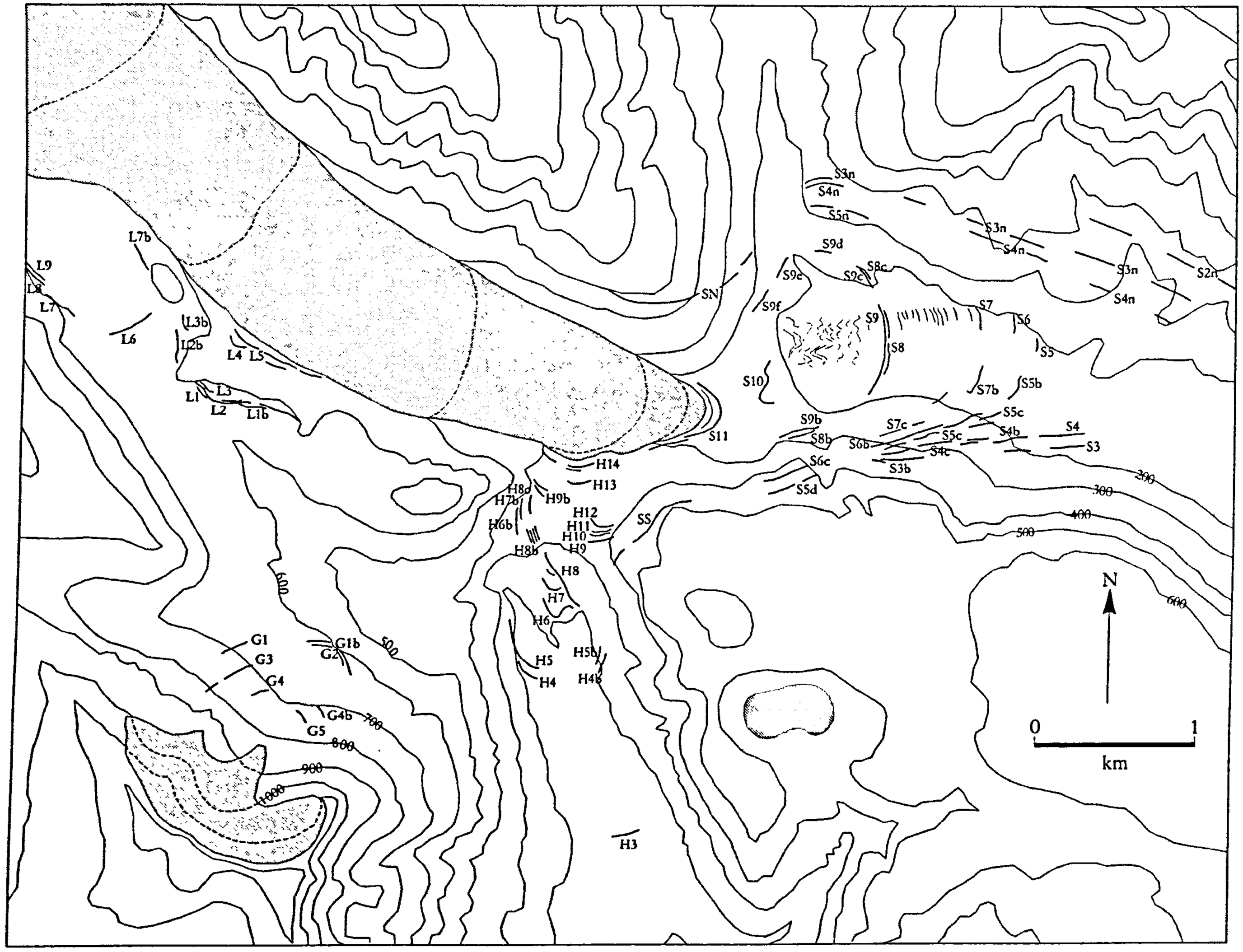


Figure 6.4 Moraines relating to former positions of Lambatungnajökull and the glacier on Grasgiljatindur. Contours are at 100 m vertical intervals. The glaciers are shown in their present-day positions (1998).

moraines are scarce. Moraine ridges have been mapped and numbered depending on their distance from the present day ice-front – starting with the most distant, as is conventional in geomorphological studies (Figs. 6.3 and 6.4). Preliminary geomorphic correlations between discontinuous sections of moraine are suggested on the basis of spatial distribution (Fig. 6.4)

Moraine S1

Situated 300-400 m above the valley-floor of Skyndidalur, on the narrow Dalsheiði plateau, Moraine *S1* is actually a number of indistinct, parallel, linear ridges (Fig. 6.5). The four ridges are laterally continuous over a distance of nearly 2000 m, trending almost parallel to the plateau edge. The outermost ridge (*S1a*) attains the greatest height, approximately 15 m above the surrounding plateau. All four are, however, subdued, low-elevation features with slopes of between 10 and 15°. The distance between the individual ridges, from crest to crest, averages 50-80 m. Large rounded palagonite boulders are strewn across the plateau but tend to cluster on ridge crests (Fig. 6.5). Soil cover is sparse. The thickest soils on the wind-swept plateau (<60 cm) were found in small hollows between ridges *S1a* and *S1b*.

Moraine S2

The outermost lateral moraine fragments in Skyndidalur are located on the northern side of the valley and are subdued features with rounded crests and asymmetrical cross profiles (Fig. 6.6). Typically the slopes of these features rest at 25°. The ridges have a thick vegetation cover and are soil-covered. The surface of this moraine is relatively free of stones and boulders. Access to this moraine was unfortunately prevented due to the impassable Skyndidalsá and Austurtungnaá rivers.

Moraine S3

A rounded asymmetrical feature, this well vegetated ridge is slightly smaller than moraine S2 and lies lower down the valley side (Fig. 6.6). At an elevation of 170-320 m asl, this linear, round-crested ridge continues along the valley side for 500 m in three discontinuous sections. The flanks of this lateral moraine slope at 20-25°. The ridge is

grass-covered and in places soils are over 1 m deep. A corresponding ridge was also identified on the northern side of Skyndidalur (Fig. 6.6). However, this feature, like moraine S2, was inaccessible during the 1998 and 1999 field seasons.

Moraine S4

A linear, sharp-crested ridge, reaching 15 m in height, this moraine is laterally persistent for over 100 m. With slopes of c. 30°, this ridge lacks vegetation and is largely free of soil (Fig. 6.7). Large rounded boulders and cobbles are strewn over the surface. The ridge terminates quite abruptly on the valley side at 160 m asl. No terminal moraine fragments on the valley floor correlate with this feature.

Moraine S5

Moraine S5 comprises two terminal fragments and four lateral ridge sections to the south of the Skyndidalsá river. The terminal moraine fragments are low elevation (<5 m), sharp-crested, features consisting of sub-rounded to sub-angular boulders and cobbles. Soil cover is thin and patchy. The lateral ridge sections associated with these terminal fragments are located between 420 and 180 m asl, attain a height of 5-8 m and are persistent for 100 m (Fig. 6.7). These sharp-crested, steeply-sloping ridges lack vegetation cover, particularly on the ice-proximal flank.

Moraine S6

Moraine S6 comprises two lateral ridges and a small terminal fragment. The lower lateral portion of the moraine, *S6b*, is the largest glacially-deposited feature on the proglacial foreland in Skyndidalur. Located 300 m asl and over 100 m in length, this sharp-crested valley-side ridge is asymmetrical in cross profile (Fig. 6.7). The ice-proximal slope is long (35 m), steep (35-40°) and unstable; whilst the ice-distal flank is short (5-10 m) and more gently sloping (25°). Much of the ice-proximal slope has slumped to reveal the internal sedimentology of the moraine – unstratified sub-rounded boulders and cobbles set within a finer matrix of sand and silt. The crest of the ridge runs broadly parallel to moraines S5, S4 and S3. A smaller lateral fragment of this moraine (*S6c*) occurs higher up the valley side at 400 m asl, somewhat closer to the

present-day glacier margin. A bouldery, discontinuous, low-elevation ridge fragment (*S6*) on the valley floor correlates with these lateral features.

Moraine S7

A distinct trimline on the oversteepened, ice-proximal flank of moraine S6 represents part of moraine S7. The main ridge fragments, however, are located on the valley floor at 180 m asl. The larger of the two (S7b) is 5-6 m high, persists for 100 m and is arcuate in plan. A smaller, bouldery, low-elevation, mound adjacent to the Skyndidalsá river is also part of this moraine (S7) (Fig. 6.8)

Arcuate *minor moraines*

A series of subdued minor moraines exist between moraine S7 and the next major ridge, S8 (Fig. 6.8). These closely-spaced rounded corrugations, barely sufficient to be termed 'ridges', are sub-parallel and arcuate in plan. The term *ribbed ground moraine* is probably a more appropriate description. Although the topographic expression of these minor moraines is less than 1 m in height, the features can be clearly seen when viewed from the air. Small pools of water tend to collect in the depressions between the minor ridges accentuating the ribbed pattern. These moraines are composed of a clay-rich sandy matrix containing poorly-sorted sub-rounded clasts.

Moraine S8 and Moraine S9

This large, arcuate ridge is the most conspicuous feature on the proglacial foreland. However, this moraine is actually a complex feature possessing two distinct crest lines – moraine S8 and moraine S9 (Fig. 6.8). The outermost ridge, S8, ranges from 4-7 m in height; the inner ridge is slightly smaller (3-5 m). Both ridges possess steep slopes (30-35°) and are lacking soil cover. In places, moraine S9 has overprinted the older moraine resulting in a single crest line. The maximum spacing between the two ridge crests is 8-10 m. South of the river Skyndidalsa, moraine S9 is well preserved; north of the river the moraine is more fragmentary. Several moraine arcs (*S9c*, *S9d*, *S9e*) correlate with the younger terminal feature. Two, small, lateral moraines (*S8b*, *S9b*) on the valley side, at 310 m and 320 asl, also correspond with the two valley-floor moraines.

Sawtooth *minor moraines*

Approximately 500-1000 m from the present-day glacier margin is a remarkably regular sequence of zigzag ridges comprised of low-elevation ridges (<2.5 m) and elongated hummocks (Fig. 6.9). The whole assemblage consists of up to twenty individual, interlocking lineations. The ridges nearest to the ice-margin are crenulated features with connected arcuate sections. An aerial photograph taken in 1945 shows that the morphology of the ridges mirrors the former outline of the ice-front (Fig. 6.9). These features resemble *sawtooth* moraines, characterised by downvalley-pointing teeth and upvalley-pointing notches, as first described at Bödalsbreen in Norway by Matthews *et al.*, (1979).

Moraine S10

Closer to the margin of Lambatungnajökull, where glacial meltwater flows in numerous unconfined channels, few moraines survive. 400 m from the present-day glacier margin is a conspicuous basaltic, bedrock knoll – 5-8 m in height and 20 m long. This bedrock feature is a streamlined form with steep, smooth sides and a rounded, convex, long profile. Striations – evidence of glacial abrasion – occur on the ice-proximal face; whilst the ice-distal face is fractured and has been subject to block removal or glacial quarrying. The feature is a *roche moutonnée*. Abutting against the ice-proximal flank of the bedrock knoll is a low-elevation, sharp-crested ridge. The ridge is sinuous and is broadly parallel to the present-day ice-margin.

Moraine S11

Adjacent to Lambatungnajökull, and covering an area of 100 x 500 m, is an ice-cored moraine complex. The ridges, rising up to 20 m in height, are steep-sided (30-45°), sharp-crested and arcuate in plan (Fig. 6.10). Glacier ice can be clearly seen in places where the debris cover has been removed. The overlying sediment consists of fine gravel, sand and silt. The debris mantle varies in thickness depending on topography. Sediment collects at the base of slopes and in small troughs. Typically the overburden is less than 0.2 m thick. Presently the rate of downmelting is very slow. No noticeable

change in moraine height occurred over the course of the field season (1998-1999), and very little modification of the ice-cored moraines has taken place since the 1980s (Fig. 6.10; aerial photographic evidence)

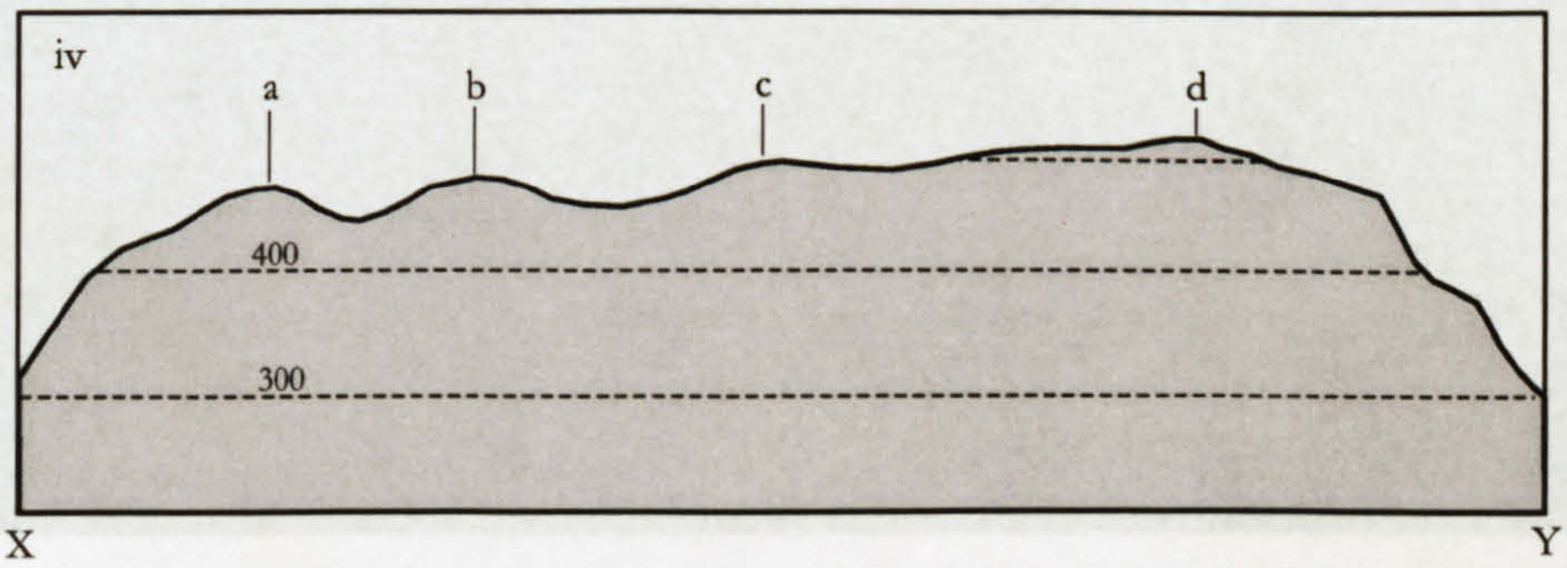
Other moraines in Skyndidalur

At an elevation of 400 m on the southern flank of Lambatungnahryggur is a large, linear, bouldery ridge with an asymmetrical cross-profile. The ice-proximal slope rests at an angle 25-30°; whilst the ice-distal flank slopes at c. 20°. This well-defined, lateral moraine, termed *SN*, continues downslope for a distance of 800 m, descending to 200 m asl. (Fig. 6.11). When viewed from the opposite side of the valley, the ridge appears to run parallel to the present-day glacier surface profile. The moraine has not been correlated with any other ridge fragments in Skyndidalur although must have been deposited when the glacier was at least 100 m thicker than it is now.

On the southern side of Skyndidalur, opposite to moraine SN, and at a similar altitude (360 m asl) lies another large, linear, bouldery ridge – moraine *SS* (Fig. 6.11). This lateral moraine has slopes resting at 30° and is well vegetated with mosses and lichens. The ridge is laterally persistent for over 500 m and continues into the pass between Skyndidalur and Hoffellsdalur. The ridge terminates abruptly where the high ground descends sharply into the valley below. This feature must represent an advanced stage of glaciation, similar to moraines S4 and SN.

Figure 6.5 Lateral moraines on Dalsheiði.

(i) The Dalsheiði plateau, looking west towards Lambatungnajökull. The photograph was taken from moraine c (see map, above) in August 1999.
(ii) Topographic map of lower Skyndidalur showing the extent of the plateau area and the moraines.
(iii) Vertical aerial photograph showing the sub-parallel, discontinuous ridges on Dalsheiði interpreted as lateral moraines. Note the rounded crests and orientation of the ridges parallel to the trend of the adjacent U-shaped valley.



(iv) Cross section through the Dalsheiði plateau from X to Y (see map). The moraine crests are marked *a-d*. Note that ridge *a* and *b* are more pronounced features.

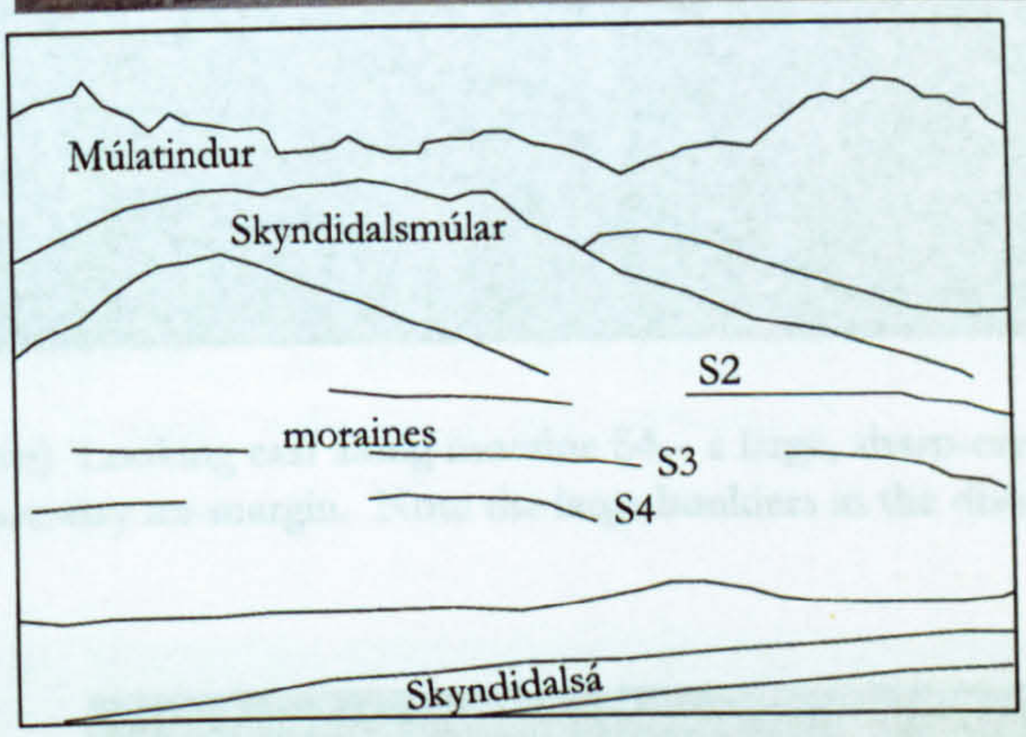


Figure 6.6 Moraines S2 and S3 in Skyndidalur.

(Above) Old lateral moraines on the northern side of Skyndidalur. Unfortunately these moraines were inaccessible and could not be studied in any detail. The outline sketch (left) highlights the elevation difference between surfaces S2 and S3 and the younger S4.

(above) Lateral moraines on the southern side of Skyndidalur. Moraine S3 is at a higher elevation and therefore older than moraines S4-S7. The small grassy mound in the foreground may be a fragment of an older moraine. Note the soil depth and the extensive vegetation cover on the outermost moraines compared to the soil-free, boulder-covered features nearer the valley floor.

Figure 6.7 Moraines S4, S5 and S6 in Skyndidalur.

(above) Looking east along moraine S4 – a large, sharp-crested, linear ridge lying 2.3 km fom the present-day ice-margin. Note the large boulders in the distance that have fallen from the valley side

(above) Moraine S6, in the foreground, is asymmetrical with a steep, unstable ice-proximal slope and shorter, more stable ice-distal slope. The glacier margin is located 1.5 km up-valley. Moraines S4 and S5, in the distance, run parallel to moraine S6. Note the lack of soil cover on these moraines.

Figure 6.8 Moraines S7, S8 and S9 and the associated minor moraines in Skyndidalur.

(above) The arcuate ridge in the centre is a complex feature possessing two distinct crest lines – moraine S8 and moraine S9. Note the ribbed appearance of the ground between moraines S7 and S8. The ridge in the foreground is moraine S6

(above) Up-valley of moraine S8-S9, lie a series of low-elevation, sub-parallel, zigzag ridges. On the ground these features are difficult to distinguish but when viewed from above a clear pattern emerges. The ridges are minor moraines, formed as the ice-margin retreated across the foreland from E to W.

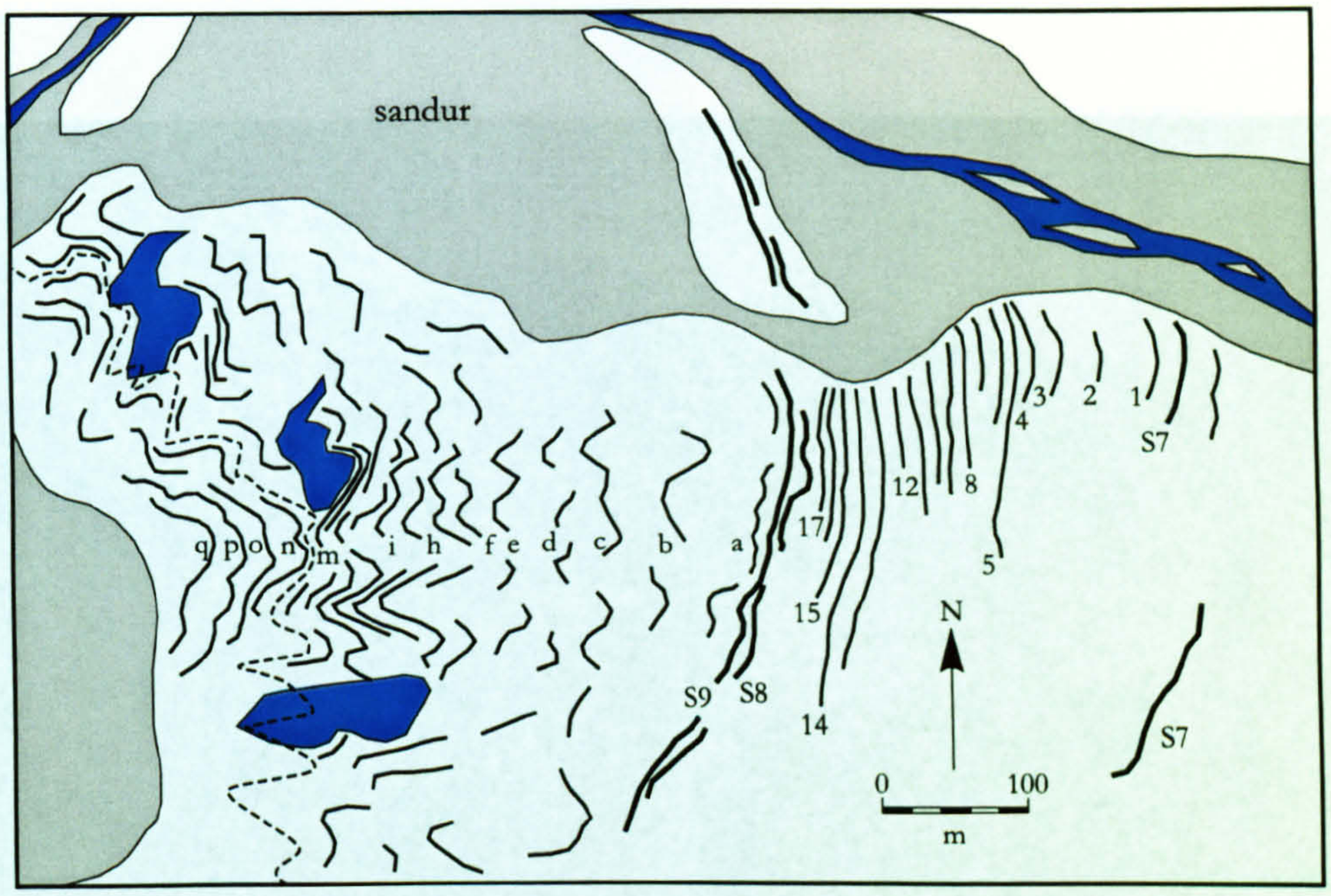
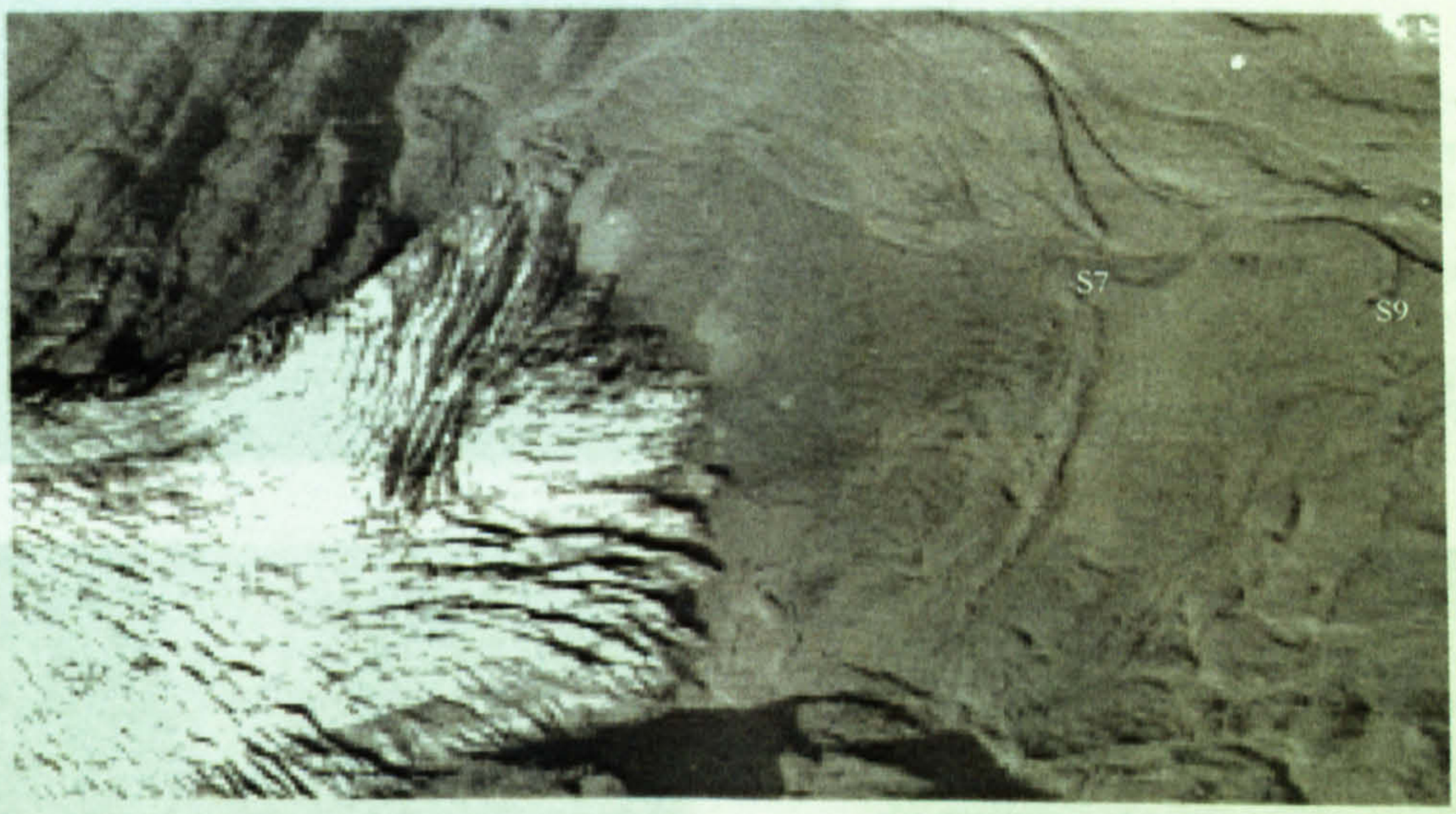


Figure 6.9 Map of minor moraines in Skyndidalur. Each ridge has been allocated a number or letter, according to convention. Note how the moraine morphology changes from *arcuate* to *sawtooth* over the sequence. Note also how the spacing between minor moraines varies. There are 17 ridges in the arcuate moraine sequence (*1-17*) and a further 17 minor sawtooth moraines (*a-q*). The dashed line represents the position of the ice margin in summer 1945. Blue areas are water bodies.



(above) Part of a vertical aerial photograph taken in September 1945 showing the minor moraines in Skyndidalur and the position of the ice-front. The scale is similar to the map above. Note the intensely crevassed glacier snout and the saw-tooth outline of the margin. Moraines S7 and S9 are highlighted.

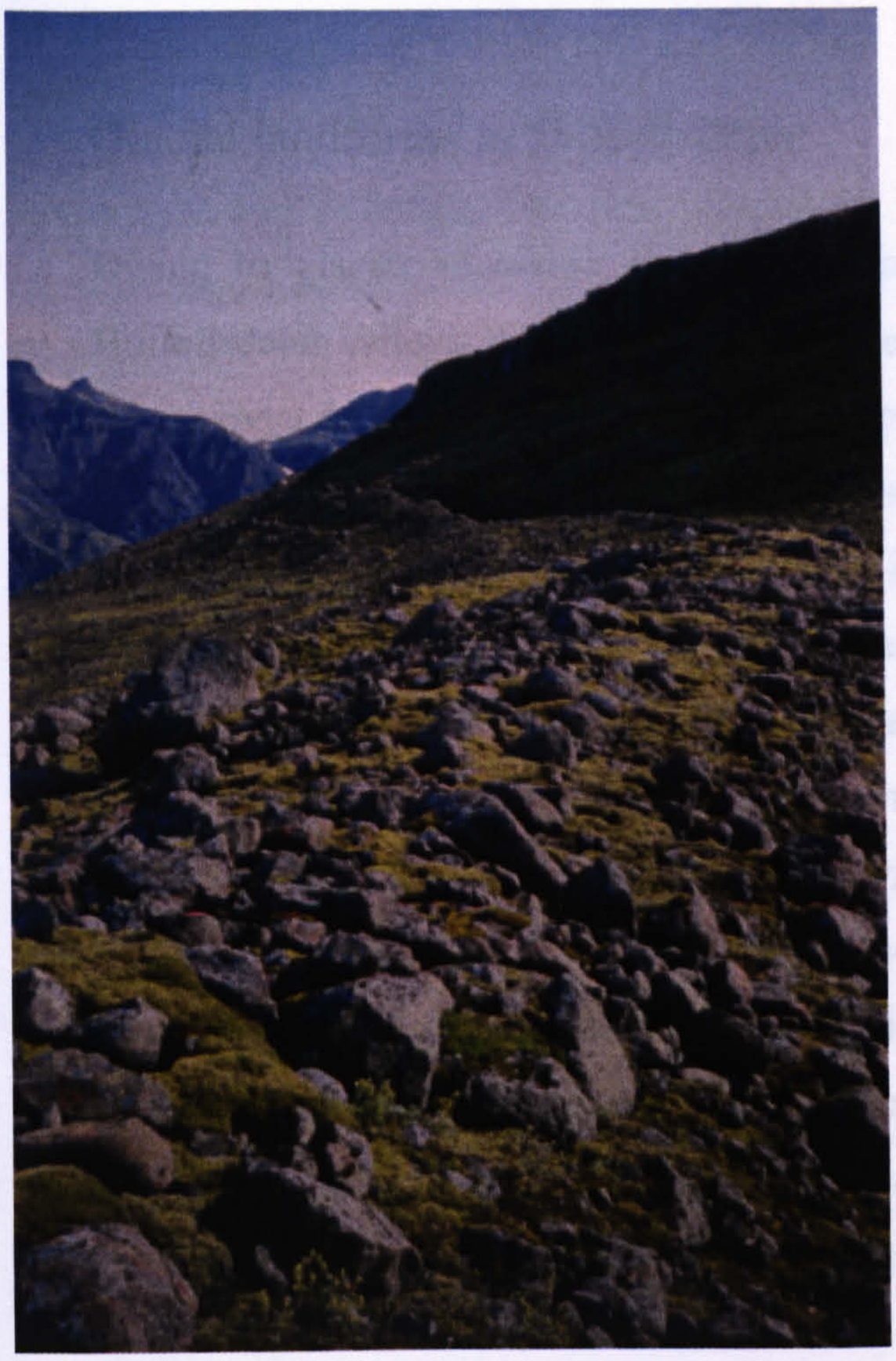
Figure 6.10 Ice-cored moraines at the terminus of Lambatungnajökull - July 1999.
(above) Note the height of the debris-covered ice (~20 m) relative to the person standing on the clean-ice margin.

(left) The terminus of Lambatungnajökull, in Skyndidalur.

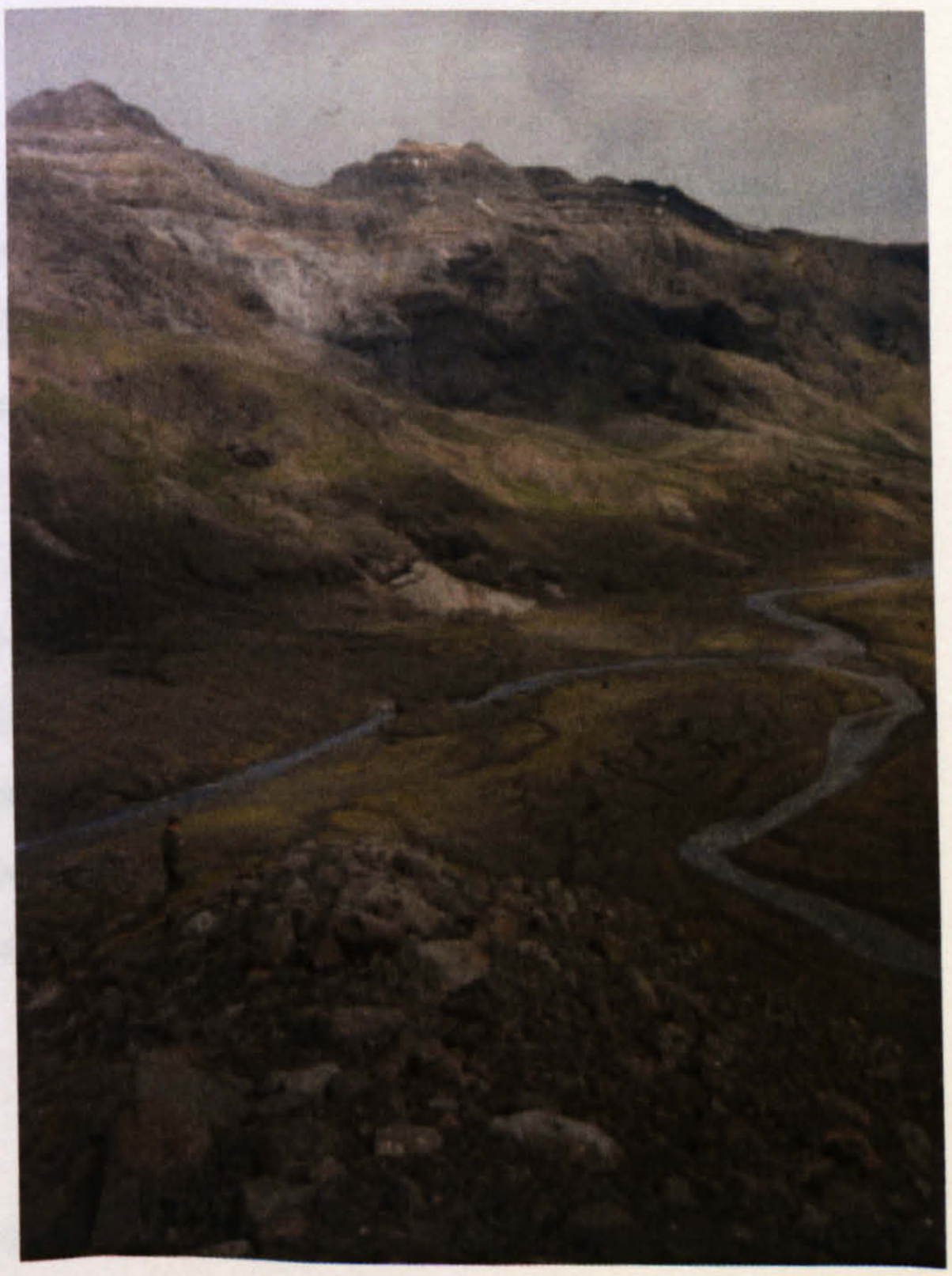
A complex series of arcuate, ice-cored push and dump moraines are evidence of ice-margin stagnation over a period of several years.

The photograph was taken in August 1999 from 400 m altitude. Compare the position of the glacier snout with the aerial photographs taken in September 1982 and 1989 (below)

Figure 6.11 Other moraines in Skyndidalur.



(left) Looking east along moraine SS – a large bouldery ridge located at 450 m asl, on the pass between Skyndidalur and Hoffellsdalur. The feature is laterally continuous for 400 m.



(right) Moraine SN, in the foreground, is a large, bouldery, poorly-vegetated ridge trending E–W. The moraine is located 100 m above the present-day glacier surface on the northern side of Skyndidalur. Note the person for scale.

Glacial landforms in Hoffellsdalur

Owing to glacier recession and the particular topography of the Skyndidalur and Hoffellsdalur valleys, the latter is now ice-free. However, a wealth of glacial landforms is testament to this valley being once occupied by an arm of Lambatungnajökull (Fig. 6.12). The absence of a glacially-fed river in Hoffellsdalur has resulted in a high degree of landform preservation, a situation rare in southeast Iceland. Presently, the valley contains only a small river, *Hoffellsá*, fed chiefly by rainwater runoff from the surrounding hills and melting snow in the hanging valley of *Fossdalur*. This river was once considerably larger and flowed in numerous, unconfined, channels. In the lower reaches of Hoffellsdalur, the valley floor is covered by an expanse of sandur.

At the head of the valley is a conspicuous complex of nested end moraines. A termino-lateral feature lies about 800 m outside this. Further down the valley are at least 3 more ridges or ridge fragments – the outermost of these lies over 7 km from the present-day ice-margin. As in Skyndidalur, these features have been numbered, starting with the most distant (oldest). However, at this stage, no correlation has been attempted with the moraines in Skyndidalur.

Moraine H1

Lying at 45-50 m above sea level, 7.2 km from the present-day glacier margin and approximately halfway along Hoffellsdalur, is a large constructive landform. This ridge is part of a termino-lateral moraine complex (Fig. 6.13). The lateral part of the feature is well preserved and forms an arcuate ridge over 20 m high in places with a rounded crest-line. The terminal part of the feature is more subdued, only 3-5 m in height, with slopes of 15-20°. The ridge has a dense covering of *Salix callicapaea*. Soils immediately inside this moraine (10-50 m to the north) are up to 2 m thick in places. A small stream, emanating from a steep gorge in the valley side, called *Selgil*, has deposited a sequence of alluvial fans. Morphological relationships suggest that the oldest of these surfaces must have been deposited prior to the construction of moraine H1 while the two younger fans represent a period of moraine dissection.(Fig. 6.13). These alluvial fans are now apparently inactive.

Moraine H2

Approximately 2000 m north of moraine H1 close to the western flank of the valley lies a linear, round-crested ridge, with slopes inclined at 25°. Only 50 m long and orientated N-S, this feature is a latero-teminal moraine fragment (Fig. 6.14). The ridge is grass-covered, although the soil cover is thinner than on moraine H1. Moraine H2 terminates quite abruptly and has no counterpart on the opposite side of the valley. The feature is composed of poorly-sorted boulders, cobbles and gravels set within a silt-rich sandy matrix. Boulders within the diamict are striated.

Moraine H3

A further 3000 m up-valley lies an inconspicuous, gently rounded rise. This feature, rising only 3-5 m above the valley floor at its highest point and over 10 m wide, and with slopes of between 5-15°, can barely be described as a 'ridge', rather a low elongate hillock. It is, however, clearly a constructional landform. Sections cut by the river Hoffellsá through the underlying sediment revealed a poorly sorted, bouldery, diamict with a sandy matrix. This low, elongate, rise is a partially-buried moraine fragment (Fig. 6.14).

Moraines H4 and H5

Close to the head of the valley are two large ridges lying only 20 m apart. Both are arcuate, termino-lateral moraines 5-12 m in height (Fig. 6.15). The innermost of the two ridges (H5) is more clearly defined and extends as a lateral valley-side expression for 200 m. Both ridges have quite rounded crests and are steep sided (~30°). The ice-proximal slope of the inner ridge (H5) has been over-steepened as a result of fluvial undercutting. Ridge H5 partly overlies a bedrock hummock. Both moraines (H4 and H5) are covered in boulders, the more angular of which probably derived from the steep rock faces above. Corresponding moraine fragments exist on the opposite side of the valley (*H4b*, *H5b*).

Moraines H6, H7 and H8

At the head of the valley, the ground rises to the pass into Skyndidalur. A series of nested end-moraine loops are located in this region, 100-400 m from the pass (Fig. 6.16). Sharp-crested, with steeply-sloping flanks (30-40°) these ridges are up to 10 m high in places. H6, H7 and H8 are considerable in size and comprised of sub-angular to sub-rounded boulder and cobble-grade material. The ice-proximal face of moraine H6 has been disturbed in places by the construction of a jeep track. The innermost ridge, H8, is the largest moraine in upper Hoffellsdalur. The ice-proximal flank of this feature slopes at 40° and descends over 30 m to the stream flowing in the gulley below. The crest of moraine H8 terminates at the foot of the cliff representing the head of the valley. Access to Skyndidalur and Fossdalur is via this ridge and along a steep ledge in the valley headwall and has been used by farmers since the 1930s (Thorarinsson, 1943). Smaller termino-lateral moraine fragments, situated in the pass between Skyndidalur and Hoffellsdalur, are correlated with moraines H6, H7 and H8 based on morphological grounds. One of these ridges, *H8b*, is a complex feature comprising at least four separate crest lines (Fig. 6.16).

Moraines H9, H10, H11 and H12

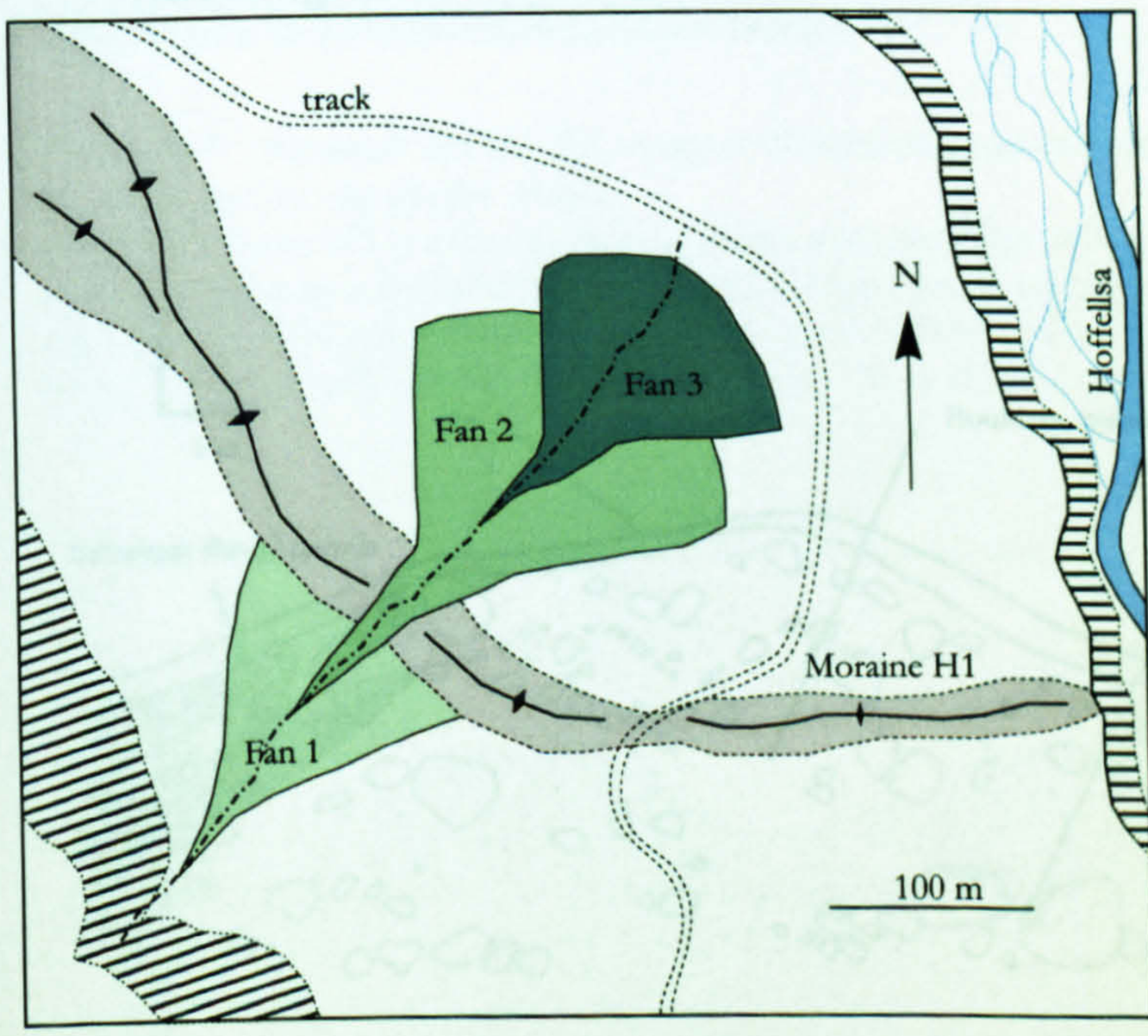
Situated in the pass, at an elevation of 350-380 m, is a moraine sequence consisting of four, closely-spaced, bouldery ridges (*H9, H10, H11, H12*). All are arcuate, low-elevation, poorly-vegetated features (Fig. 6.17). The outermost and innermost are slightly larger than the other two. A nearby ridge associated with moraine H9 or H10, tucked beneath an overhanging bedrock exposure, consists entirely of unweathered, angular, basaltic boulders without a matrix support. These moraines, H9-H12, on the northern edge of the pass into Hoffellsdalur, mark the glacier's final pause before retreating into the adjacent valley.

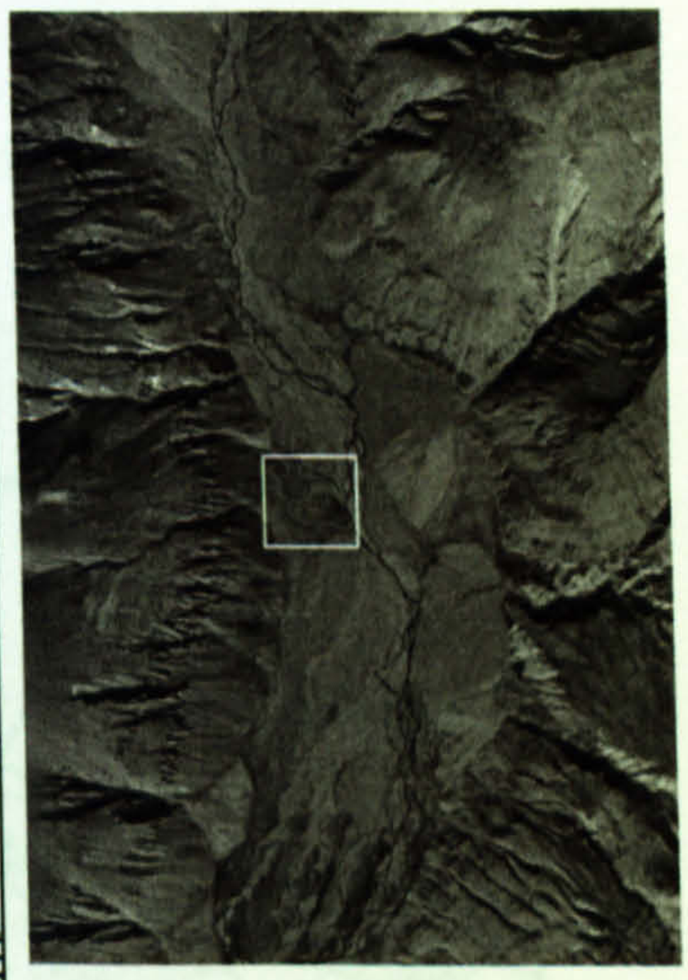
Figure 6.12 Moraines in Hoffellsdalur.

(above) A complex system of nested lateral and terminal moraines lie at the head of Hoffellsdalur – less than 1 km from the present-day ice-margin.

(above) Geomorphological evidence that Lambatungnajökull once occupied the valley of Hoffellsdalur. At least 5 moraine ridges, each relating to a different advance event, have been identified (H4-H8).

Figure 6.13 Moraine H1, lower Hoffellsdalur, located 45 m asl and ~7 km from the present-day ice-margin. The picture was taken from the highest point on the moraine. Note the Land Rover parked in the middle distance, adjacent to the moraine.

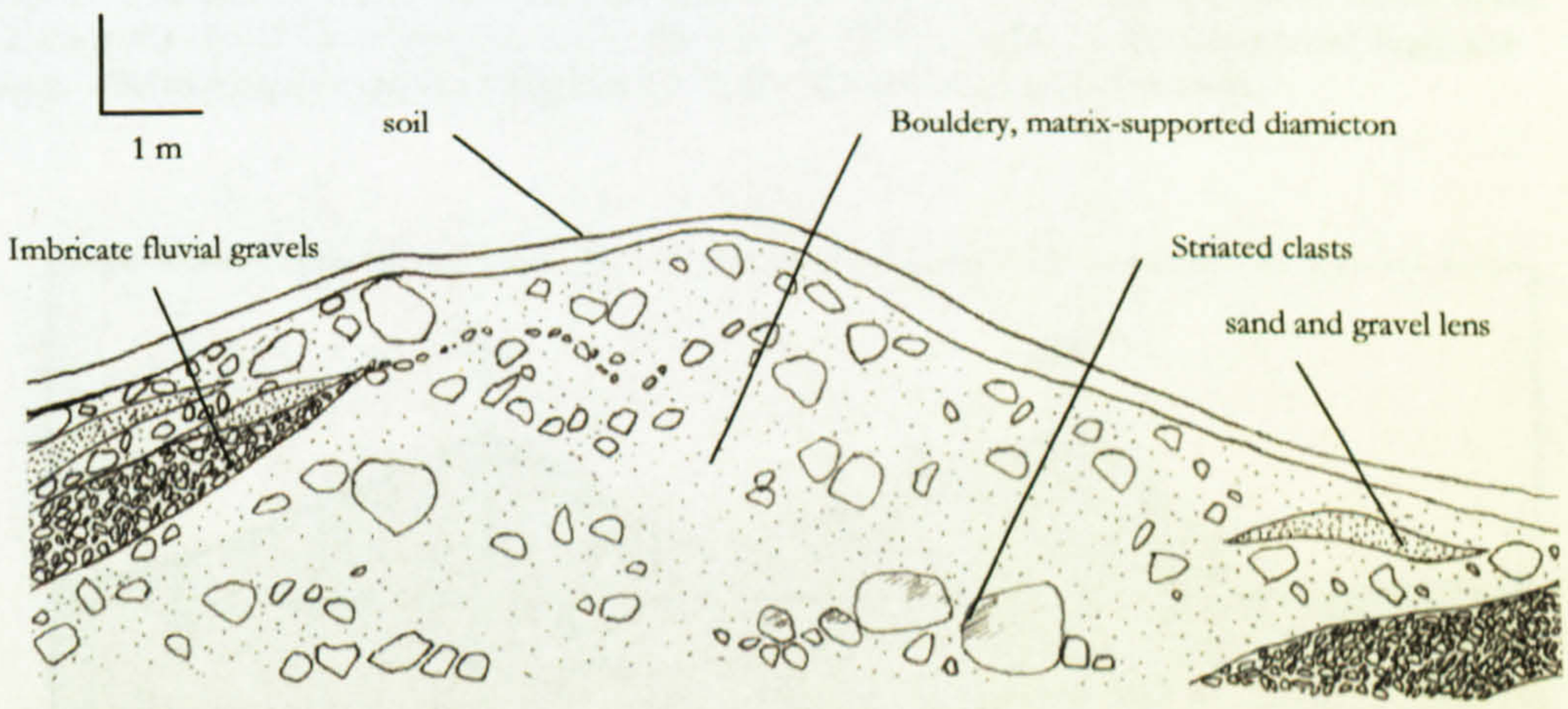




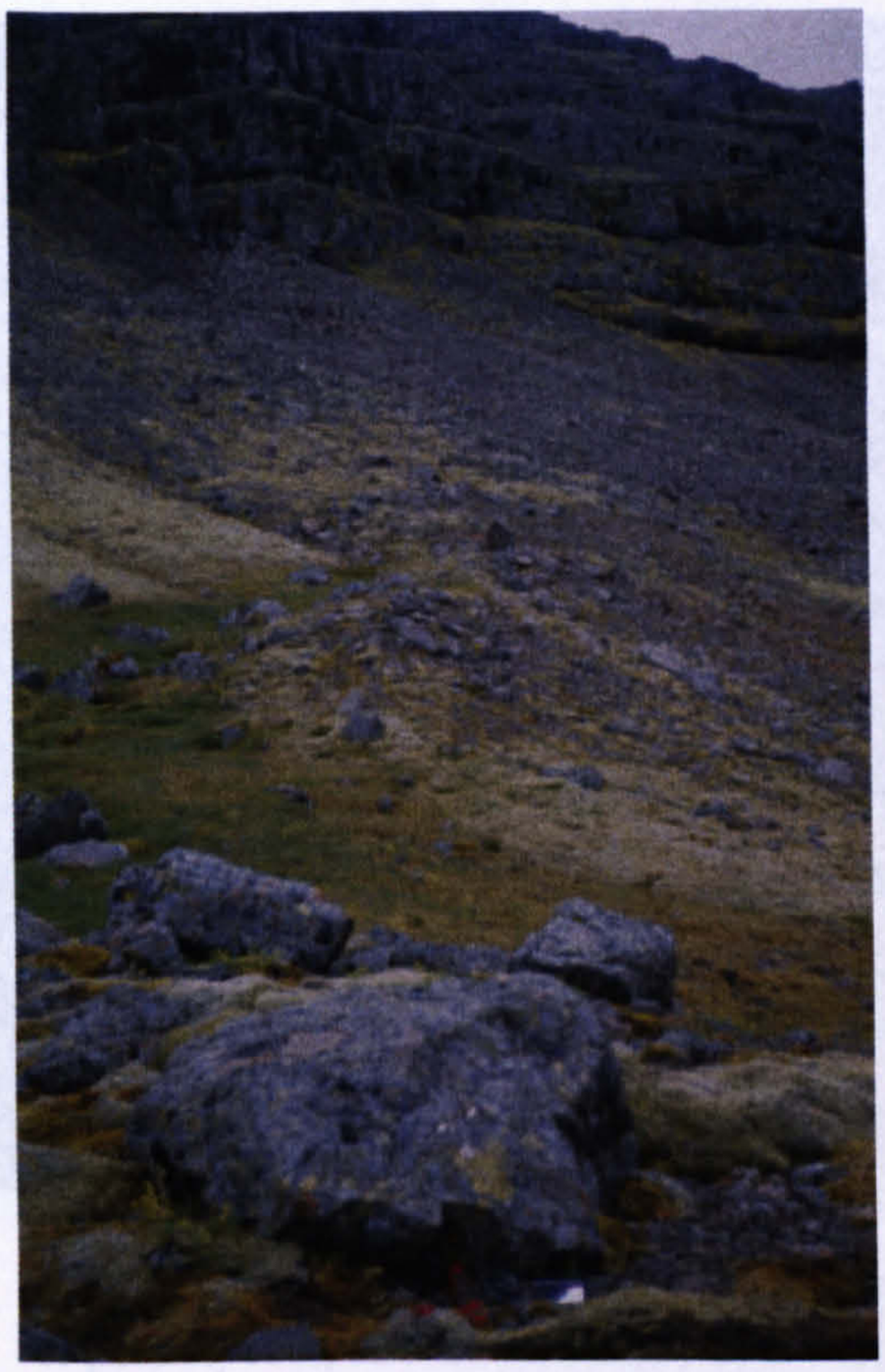
(Above) Aerial photograph of lower Hoffellsdalur. The geomorphological map (left) details the area inside the small square. Moraine H1 relates to the former extent of Lambatungnajökull. A sequence of alluvial fans have also been mapped. The oldest surface, Fan 1, was deposited before the ice-margin was located at moraine H1. Fans 2 and 3 are younger than the moraine.

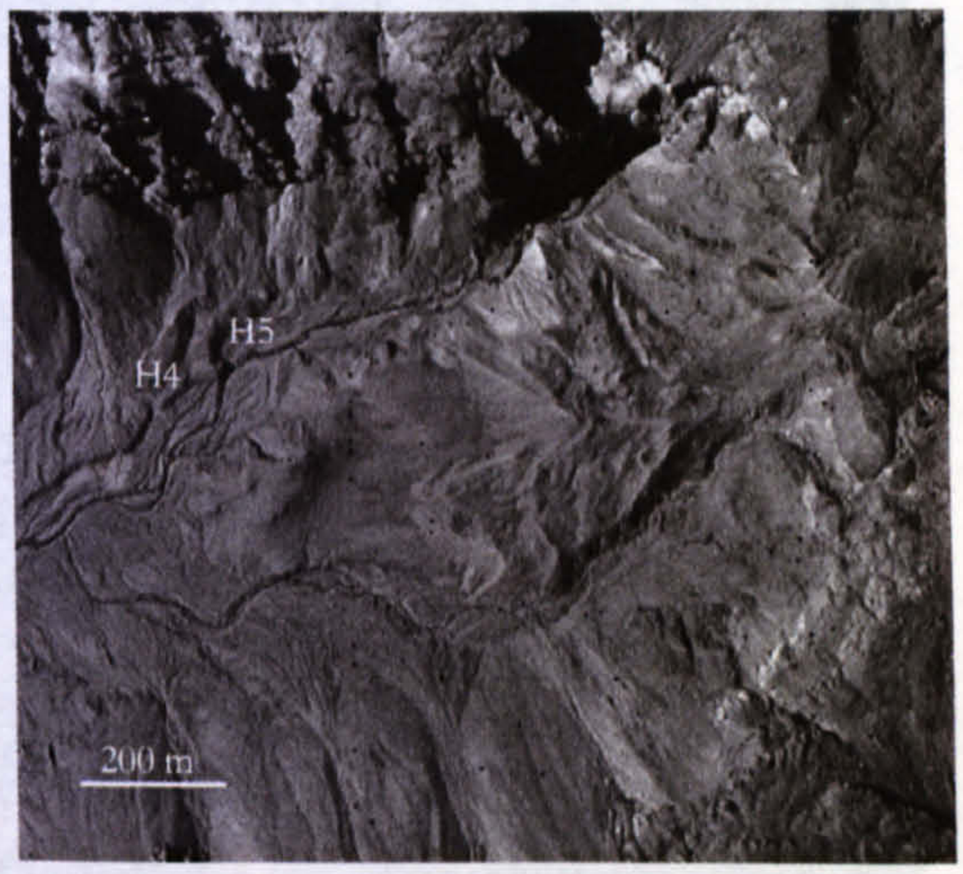
Figure 6.14 Moraines H2 and H3

(above) Moraine H2 (left of grey vehicle, middle distance) is a small, round-crested, moraine fragment located 3.5 km from the present-day margin of Lambatungnajökull. Note the typical road conditions in southeast Iceland!



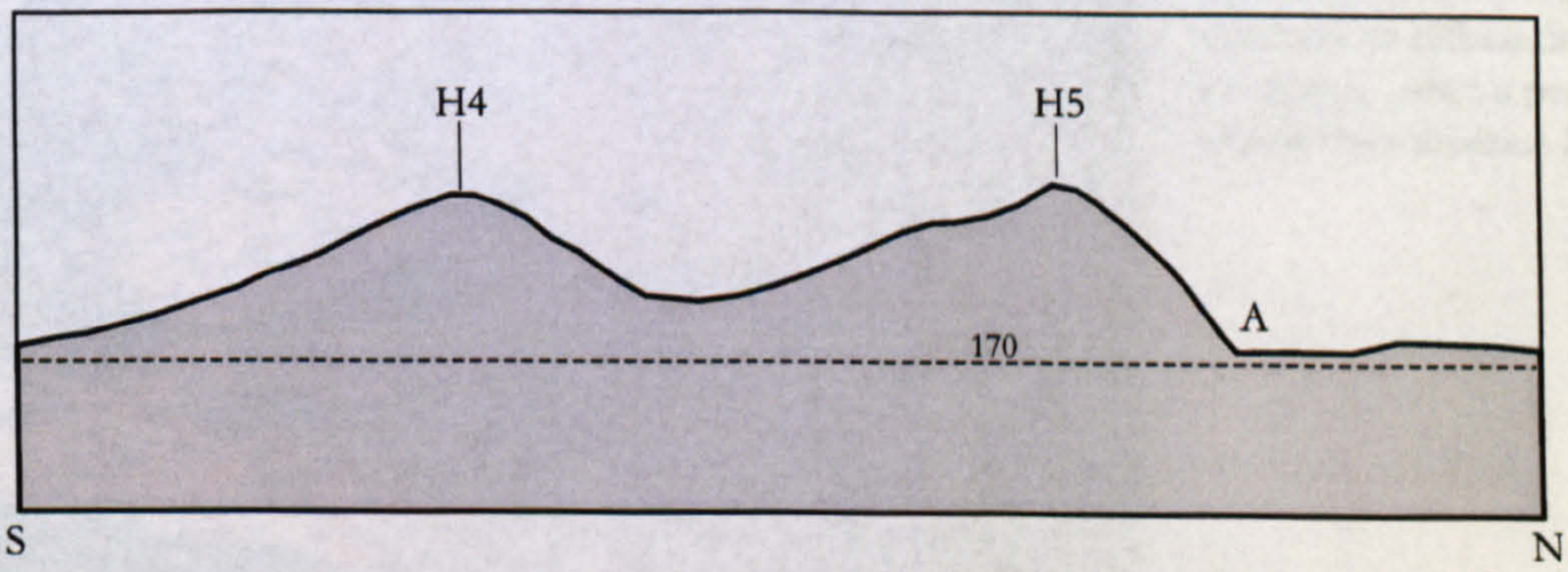
(above) A cross-section through the low grass-covered rise, moraine H3. Striated boulders set within a silty, clay-rich, matrix suggest a glacial origin for this feature.





(Above) Part of a vertical aerial photograph taken in 1982 showing the moraines in upper Hoffellsdalur. North is to the right. Moraine H4 and H5 are marked. H5 is a 100 m-long arcuate ridge which continues as a lateral expression for 200 m along the valley side. Large portions of both moraines have been eroded by the river *Hoffellsá*.

Figure 6.15 Moraines H4 and H5 in upper Hoffellsdalur are situated at 170 m asl and 1000 m from the present-day glacier margin.
(above) Moraine H5 is a clearly-defined, large, rounded ridge rising 12 m above the surrounding sandur. The picture looks NW along the ridge crest. Some of the large angular boulders in the distance may have fallen from the steep cliffs above. The boulder in the foreground supports large (>80 mm) yellow-green *Rhizocarpon* lichens. Note the compass for scale.



(iv) Cross section through moraines H4 and H5 from north to south. The proximal flank of moraine H5 has been oversteepened by fluvial erosion at the point marked A.

Figure 6.16 (above) Moraine H8b (middle distance) is a complex feature consisting of three, or possibly four, distinct crest lines. This moraine, situated on the pass between Skyndidalur and Hoffellsdalur, was deposited by the lateral margin of Lambatungnajökull. Moraine 6b and 7b are in the foreground.

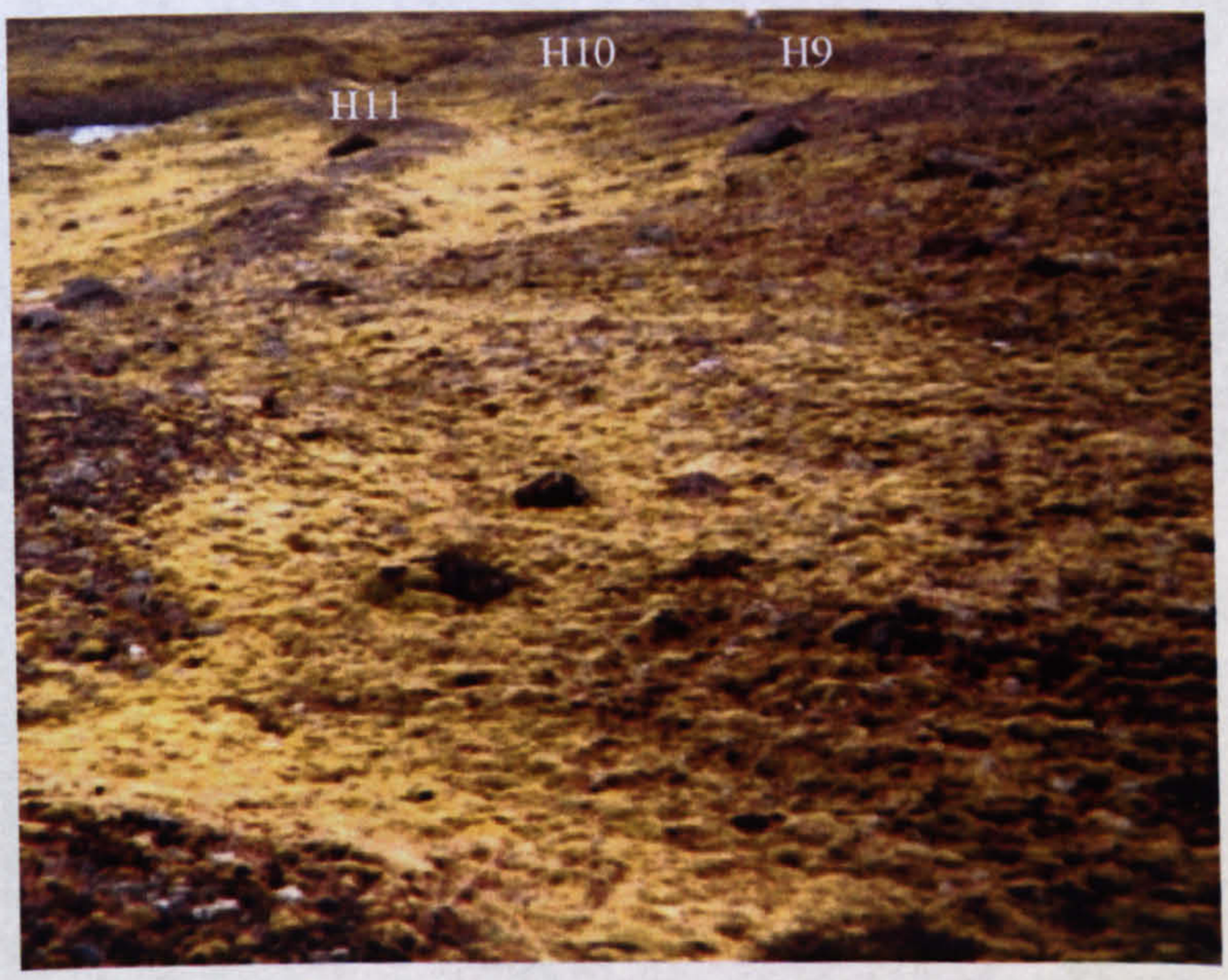


Figure 6.17 (left) Moraines H9, H10 and H11 are closely-spaced arcuate ridges c. 200 m in length. These low-elevation features represent small re-advances or stillstands of the ice-margin , after a period of retreat from moraine H8.

(left) 'Micro' moraine, <0.5 m high, formed by a combination of squeezing and 'bulldozing' processes. Minor fluctuations of the ice margin can result in features of this scale. Preservation potential of these ice-moulded forms is low. Note the logbook for scale.

(Picture taken 27.07.99)

Figure 6.18 Small-scale constructional features on recently-exposed ground adjacent to the southern margin of Lambatungnajökull.

(above) Emerging from beneath the margin of the glacier is a small ridge of unconsolidated sediment. Probably formed by squeezing of soft sediment into a cavity in the sole of the glacier, this feature extends for only 6-7 m. Note the high debris content of the overlying basal ice-facies. (Picture taken 29.07.99.)

Moraines H13 and H14

The ground falls steeply from the pass into Skyndidalur. The present-day glacier surface lies approximately 50 m below the level of the pass. Between the glacier margin and moraine H12 lie two, low-elevation (<3 m), sharp-crested ridges orientated parallel to the present-day ice-margin. Consisting of boulders and cobbles set within a clay-rich sandy diamict, this feature is asymmetric in cross-profile with a steep (40°), unstable ice-proximal slope. Minor, ice-sculpted forms – composed of unconsolidated silt, sand and gravel – are located in the 30 m-wide zone immediately adjacent to the glacier margin. These small-scale features (<0.5 m high) represent recent, minor, fluctuations of the ice margin of Lambatungnajökull (Fig. 6.18).

Glacial landforms in Fossdalur

Adjacent to Hoffellsdalur and flowing east-west, *Fossdalur* (sometimes referred to as *Vesturdalur*) is a classic U-shaped, hanging valley approximately 3 km in length. The valley is flanked to the north by the mountain of *Fossdalshnúta* (775 m) and by the impressive, ice-capped, *Grasgiljatindur* (1230 m), to the south. Two rivers, the *Fossá* and the *Vesturdalsá*, join approximately halfway down the valley. The river then tumbles over 150 m-high cliffs into Hoffellsdalur below. The waterfall, which is unnamed on maps, is most spectacular.

Seemingly cut off from other valleys, a narrow path follows the cliff edge making Fossdalur accessible from the pass between Hoffellsdalur and Skyndidalur. Grass is plentiful in Fossdalur and the soil is thick (>1 m) in the valley bottom. The valley is home to many grazing sheep in summer. During much of the year, however, the valley is snow-covered. Reconnaissance during May 1999 found the late spring snowline to be at around 500 m asl. Several semi-permanent snow patches were still present three months later (August 20 1999) when the snowline in Fossdalur lay at c. 900 m asl (Fig. 6.19)

The glacial geomorphology in Fossdalur reflects the former extent of a small cirque glacier on *Grasgiljatindur* (Fig. 6.20). The northern flank of the mountain, Grasgiljatindur, supports a firn-field which feeds a small, steep, outlet glacier. This glacier is currently thin and measures less than 400 m in length. Debris covers the ice-surface in the ablation zone making the exact location of the glacier terminus unclear. The term *glacieret* is probably more suitable for this feature.

Moraines G1 - G5

All five moraines associated with the glacier on Grasgiljatindur consist of arcuate, low-elevation, bouldery ridge fragments. The outermost two moraines, G1 and G2, lie at 620 m asl and are 2-3 m high with low-angled flanks (15-20°). The crests of these two features are sub-rounded. The vegetation consists of *Rhacomitrium* moss and lichens. Ridge G3 lies at 700 m asl, is 3 m high and 100 m long, and is the most conspicuous of the five moraines (Fig. 6.21). Only small fragments of the innermost moraines (G4 and G5) were identified. These represent the most recent advance of the small cirque glacier. The ground adjacent to the glacier snout is strewn with sub-rounded cobbles and boulders.

Moraines L1 - L5

Lambatungnajökull was once considerably thicker than it is now. At an elevation of 600 m asl, 60 m above the present-day surface of the ice, the southern margin of the glacier deposited a series of moraines on the valley side (Fig. 6.21). These moraines can only be reached via Fossdalur and are almost 5 km from the glacier terminus in Skyndidalur. The outermost of these arcuate, low-elevation features, *L1*, has been overridden by a later glacier advance. This subsequently-deposited ridge (*L2*) has an arcuate plan form and extends for 200 m. A further ridge, *L3*, is located inside L2. The only vegetation on these moraines are lichens. The distance between all three ridge crests is nowhere more than 30 m. Parallel to the glacier margin, 60 m below L1, are two recently-deposited moraines, L*4* and L5.

Figure 6.19 Annual and seasonal variations in snowcover

(above) Variation in the size of snowpatch A, in upper Fossdalur (600 m asl), between 1998 and 1999.

Seasonal variation of snowcover in Fossdalur (spring and summer 1999)

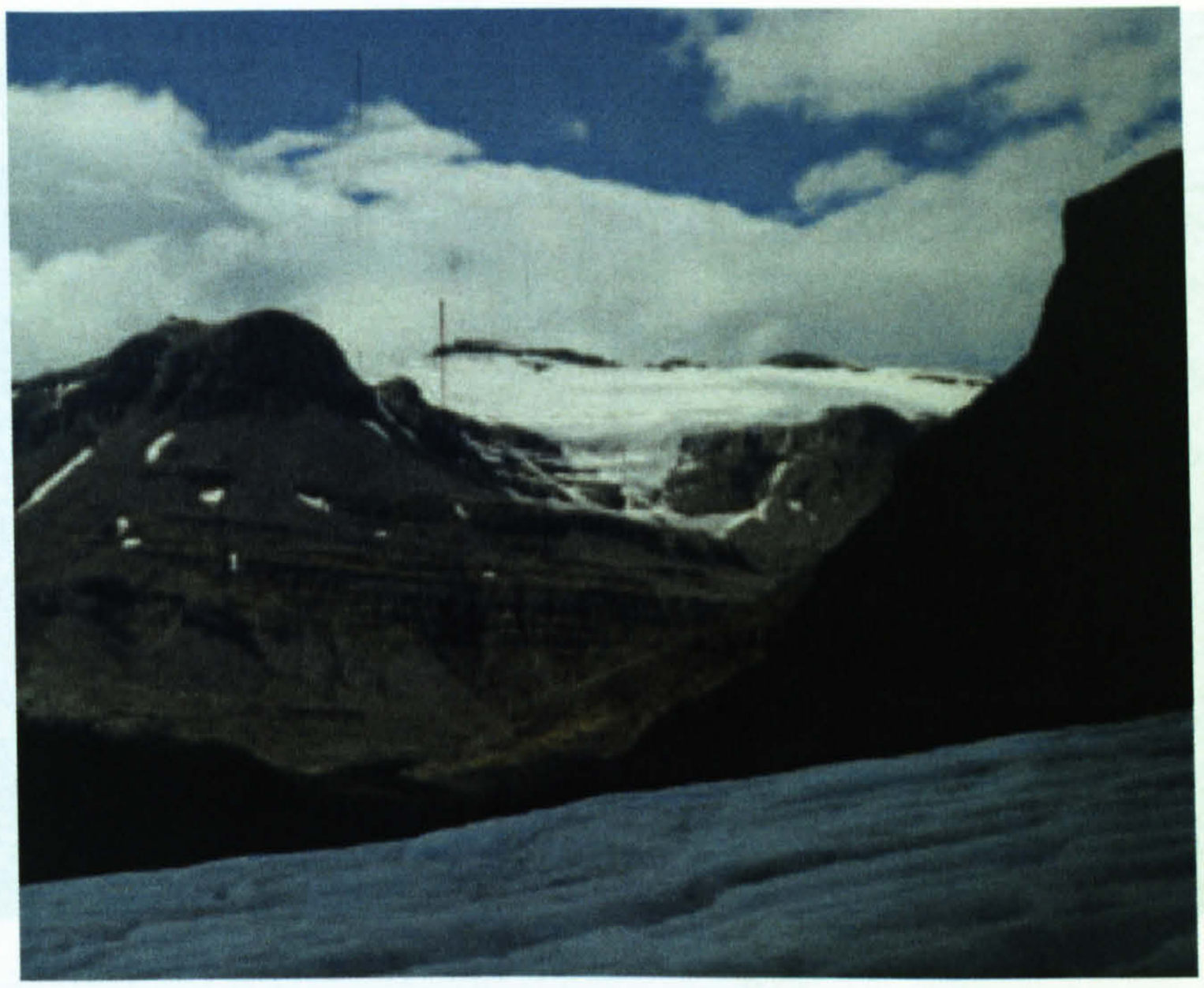


Figure 6.20 Moraines in Fossdalur associated with the glacier on Grasgiljatindur

(above) Grasgiljatindur rises to 1230 m and supports a firn field. A small, unnamed, glacier occupies the cirque in the northern side of the mountain. Lambatungnajökull is in the foreground

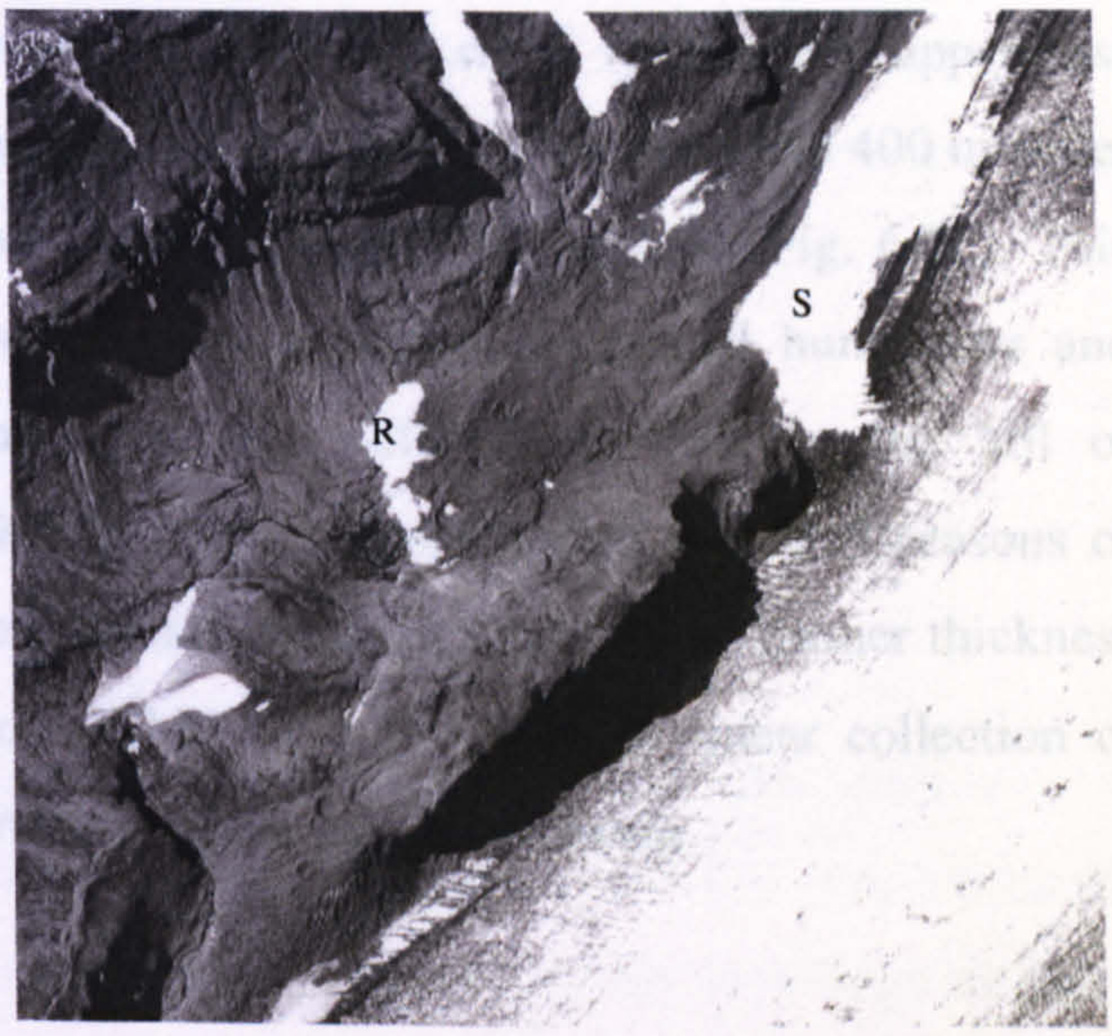
(above) Moraines on the valley side relate to the former extent of the glacier on Grasgiljatindur. Moraine G3 is pictured. This low-elevation, bouldery ridge lies 800 m from the present-day ice margin.

Figure 6.21 Moraines in Fossdalur associated with Lambatungnajökull

(above) A series of moraines deposited by the southern margin of Lambatungnajökull, referred to as L1-L5.

(above) A lobe of ice, ~1 km long and 400 m wide, once fed by Lambatungnajökull, is now debris-covered and stagnant. Several lateral moraines relate to the former thickness and areal extent of the ice-lobe (L6-12). (Also see Figure 6.22.)

Figure 6.22 Dead-ice lobe in upper Fossdalur



(right) Aerial photograph taken in 1982 showing part of the southern margin of Lambatungnajokull and the lobe of debris-covered ice. A large area of glacier ice, flanked by the snowpatches *R* and *S*, is currently buried beneath a sediment mantle 0.5 - 1 m thick. Note the hummocky appearance of the ice-cored terrain. Very little surface lowering of the debris-covered ice has occurred since 1945.

At the margins, where debris cover is thin, the ice-core is melting more rapidly (i). Mass movement processes, such as gravity-driven slumps, are common on the steeply-sloping ice-margins (ii). The ice-cored terrain is dissected by tunnels and vertical shafts, forming a karst-like topography. In photograph *iii* the ice has melted to reveal a portal. Sediment has fallen from the surface above to form a *dump cone*. Note the sedimentology of the dump cone and the total thickness of the buried ice mass (4-5 m). Photographs taken in July 1999.

Debris-covered ice lobe

A few hundred metres beyond the source of the Fossá river, in the very uppermost reaches of Fossdalur, is an interesting feature. A lobe of ice, 1 km long and 400 m wide, once fed by Lambatungnajökull, is now debris-covered and stagnant (Fig. 6.22). This area of ice-cored terrain – characterised by low-elevation, pyramidal hummocks and shallow sediment-filled hollows – occupies the valley floor between the hill of Gunsteinsfjall and the valley side. The area was mapped during the field seasons of 1998 and 1999. Several, poorly-defined, lateral moraines relate to the former thickness and areal extent of this dead-ice lobe (L7-L9). L6 is a broadly linear collection of mounds and hummocks marking the terminus of the ice-cored terrain.

Other glacial landforms in Hornafjörður and Lón

Approximately 22 km from the present-day snout of Lambatungnajökull, only 3 km from the North Atlantic coast, and at an elevation of 20 m asl is a conspicuous area of raised ground. The ground rises 10-15 m higher than the surrounding *Jokulsá í Lón* sandur plain and slopes in two opposing directions:- towards the sea and inland. The whole feature is c. 800 m in length and forms a broad, low-elevation, ridge parallel to the coastline. The land is known as *Hraunkot* and is currently farmed. Large boulders of basalt and palagonite occur in the fields, some of which display striations. In places, the soil cover is thin, although no exposures of the underlying sediment were observed. The morphology of this feature suggests that it is a moraine fragment.

Terminal moraines at similar altitudes – 15-30 m asl – are found at the entrance to several valleys in Hornafjörður. Laxárdalur, Endalausidalur, Gjáðadalur, Þorgeirsstaður, and Kastardalur on Vestrahorn all possess conspicuous ridges in their lowermost reaches (Fig. 6.3). The largest of these is at the mouth of Laxárdalur. The moraine is 10-20 m high and c. 1000 m long with a rounded crest and gentle slopes of 10-20°. The moraines at Endalausidalur are smaller and consist of several sub-parallel ridges. The moraine in Gjáðadalur is a termino-lateral feature c. 250 m long. A small area of hummocks and

hollows occurs in Kastardalur c, 1200 m from the low-elevation terminal moraine. Finally, two large ridges on the coastal plain 1-2 km east of Reydarártindur are also believed to be moraines. These were first identified by Nordðahl and Einarsson (1988), along with other moraines in Altafjörjður and Berufjörður to the north.

Non-glacial landforms

Relict rockglaciers

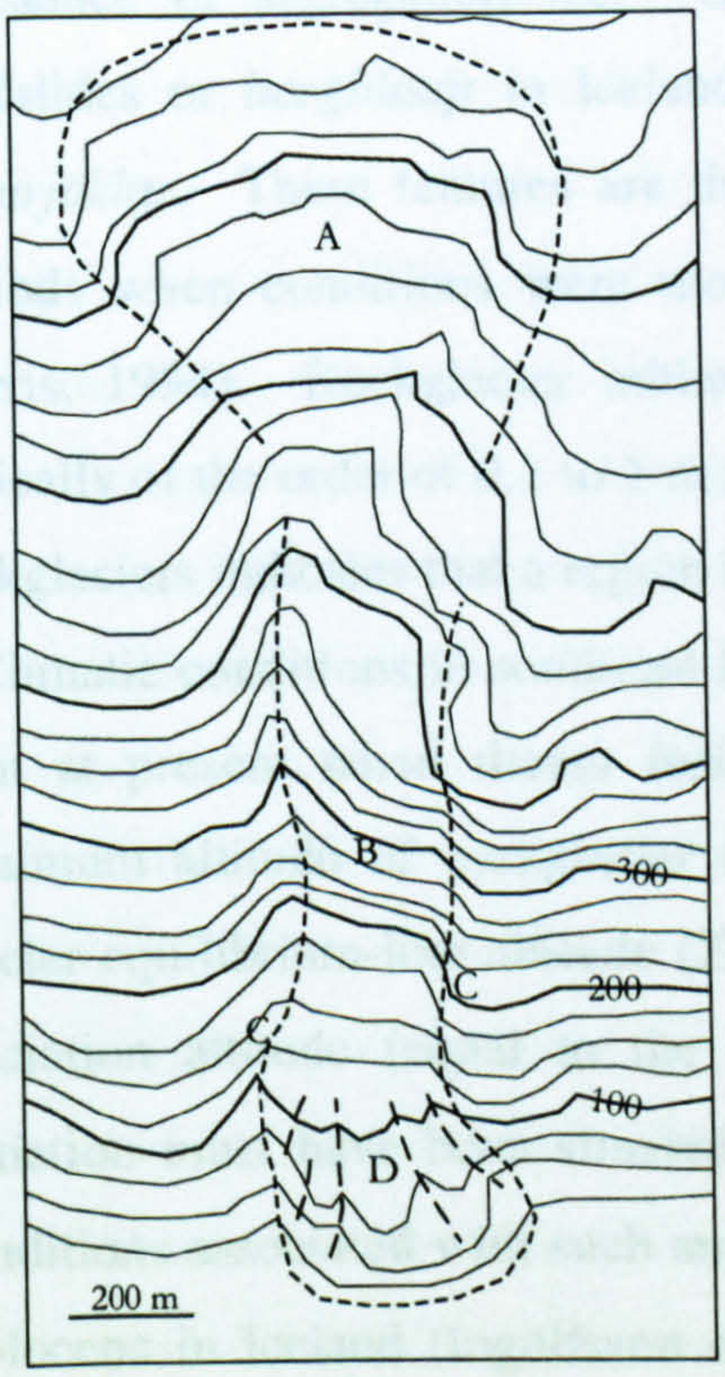
Situated in lower Hoffellsdalur, over 9 km from the present-day terminus of Lambatungnajökull and 2000 m downvalley of moraine H1, is an enigmatic geomorphological feature (Fig. 6.23). Located on the valley side is a large elongate depression, orientated downslope and stretching over 1000 m, terminating at 25 m asl. This depression, known as *Gjúfrabotn*, is delineated by distinct lateral ridges and has a clearly defined backwall at ~500 m asl. At the foot of the depression, on the valley floor, is a large accumulation of debris that rises 40-50 m above the surrounding sandur. This debris accumulation displays longitudinal ridges and depressions. The debris appears to have derived from the depression above. The whole landform is comprised of sub-angular, poorly sorted, matrix-supported diamict.

A similar debris tongue has been identified at 500-700 m asl on Setbergsheiði, 2km east of Gjúlfrabotn (Fig. 6.23). Unconfined by lateral walls and with a surface slope of of 10-15°, this debris accumulation is generally 300-400 m wide and broadens downslope. Longitudinal furrows on the lobate terminus splay out giving an impression of flow.

The Setbergsheiði feature is probably a relict, talus-derived, rockglacier and if proven would be the first rockglacier to be identified in this part of Iceland. The genesis of the Gljúlfrabotn landform is less clear. A detailed geological and stratigraphical investigation would be required to determine the exact formation and probable age of such a feature. However, based on purely morphological criteria – plan shape, lobate terminus and longitudinal furrows – the landform is probably the result of mass movement processes either in the form of a landslide or as a now-relict, talus-derived, rockglacier.

Figure 6.23 The possible relict rockglacier in Hoffellsdalur.

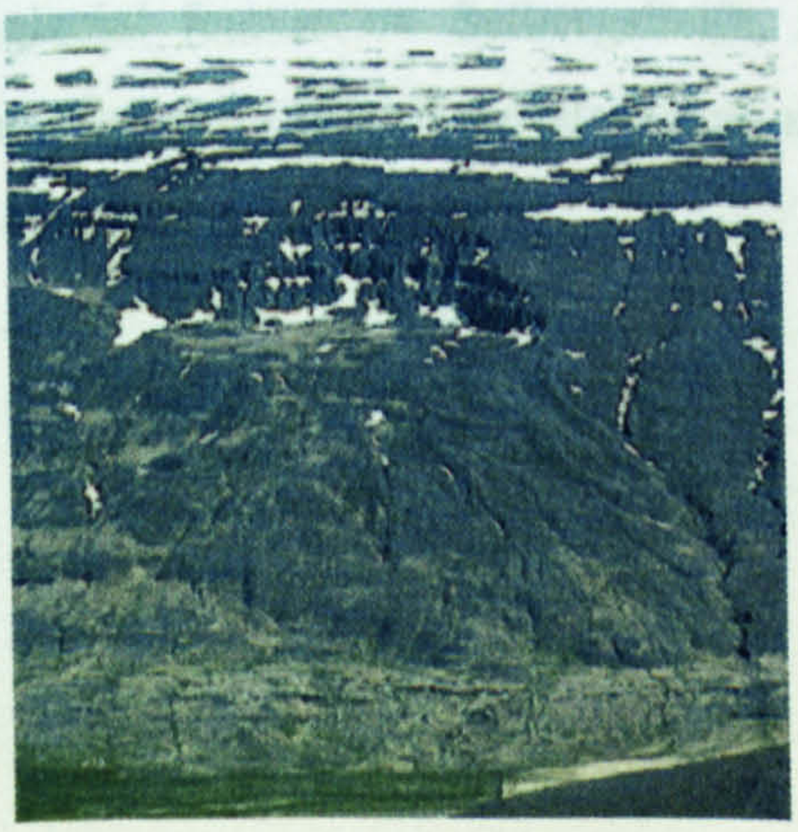
(above) The valley-side depression known as *Gljúfrabotn* in lower Hoffellsdalur. Note the steep lateral ramparts and large, lobate sediment accumulation at the base of the feature.

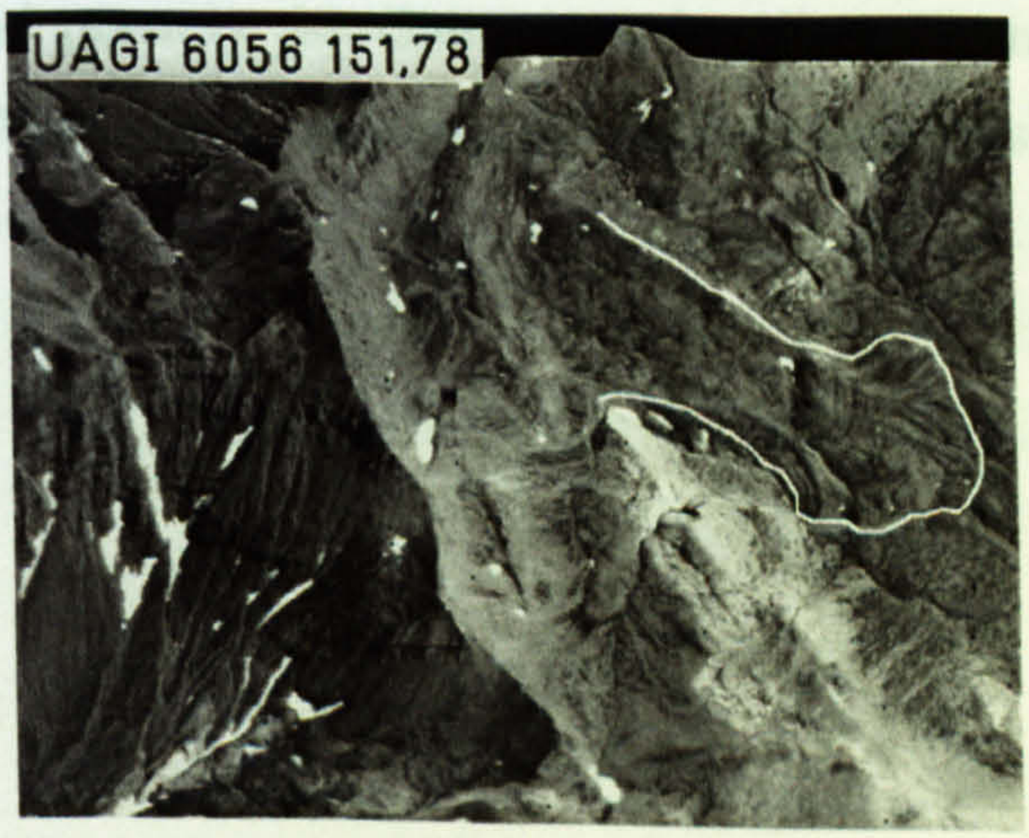


(left) Detailed topographic map of Gljúfrabotn. Contours are at 25 m vertical intervals. North is to the right. The main parts of the rockglacier feature are discernible:

A - cirque-like 'accumulation' area
B - earthflow
C - lateral ramparts
D - lobate 'terminus' with axial furrows

(Below) Morphologically similar valley-side features elsewhere in Iceland. The colour photograph shows a landform in Vatnsdalur, northern Iceland, which A. Gudmundsson (1995) believes to be a relict, talus-derived, rockglacier. The black-and-white aerial photograph shows the area immediately north of Gljúfrabotn in Hoffellsdalur. This probable rockglacier (highlighted) is identified here for the first time.





Palaeo-environmental implications

Rockglaciers are rare in southern Iceland. Numerous studies have been conducted on debris-covered glaciers and ice-cored rockglaciers in the Tröllaskagi highlands of northern Iceland (Whalley, 1974; Martin and Whalley, 1987; Martin *et al.*, 1994; A. Gudmundsson, 1995). In contrast, only one previous report of a rockglacier in the southeast of Iceland exists (Eyles, 1978). The identification of relict rockglaciers is fraught with difficulties. In northern and eastern Iceland, Gudmundsson (1995, 2000) has identified many features that he believes to be relict rockglaciers of various ages. Morphological and stratigraphic evidence has revealed that, although no longer active, these features formed as the result of downslope creep processes, probably with the assistance of segregation ice. Gudmundsson (1995, 2000) argues that numerous landslides or *berghlaup* in Iceland should be re-interpreted as relict rockglaciers or *erðarjöklar.* These features are thought to have formed during colder, drier climatic periods when conditions were more favourable for ice preservation (Ballantyne and Harris, 1994). Rockglacier initiation may take several centuries and flow rates are typically of the order of 0.1 to 1 m per year (Humlum, 1998b). The preservation of relict rockglaciers indicates that a region has not been subsequently overrun by glacier ice.

Climatic conditions in southeast Iceland must have been considerably cooler and drier than at present when theses features formed. Humlum (1998b) suggests that the minimum altitude of rockglacier initiation is similar to, but slightly lower than, the glacier equilibrium-line altitude (ELA) in maritime climates. By implication, the local glaciation altitude (equal to the ELA) on Setbergsheiði at the time of rockglacier initiation must have been situated at, or slightly above, 600-700 m asl. The climatic conditions associated with such an ELA have probably not been satisfied during the late Holocene in Iceland (Ingolfsson *et al.*, 1997). It is more likely that the now-inactive periglacial feature on Setbergsheiði, and possibly the Gjúlfrabotn landform, formed during the early part of the Holocene and has not been overrun by advancing glaciers in the intervening millennia.

Turf-banked and Stone-banked terraces

Figure 6.24 shows good examples of soli-(geli-)fluction landforms in Fossdalur. These *turf-banked terraces* are comprised of a small vertical step or *riser* (<1 m in height) and a gently-sloping *tread.* Gradually, with increasing altitude, the terraces become *stone-banked,* lacking soil cover and vegetation, and consisting of coarse, frost-shattered, angular material. The turf-banked terraces consist of soil and colluvial debris with occasional sub-angular clasts. A small pit dug near the toe of one terrace in August 1999 revealed a thin surface *Ah* soil layer overlying a diamict – angular gravel and cobble-sized clasts within a fine-grained sand-loam matrix. Below this, at c. 0.6 m depth, was another weakly developed *Ah* layer overlying a stony horizon. The lower organic horizon is believed to represent a soil layer that has been overrun by the gradual downslope advance of the turf-banked lobe (Fig. 6.25).

Palaeo-environmental implications

Low elevation, lobate tongues of soil and rock are common features on low angle (2-20°) hillslopes in southeast Iceland. Solifluction, defined as the slow flowing of water-saturated regolith from higher to lower ground (Andersson, 1906), works in combination with gelifluction and frost creep processes in cold climates to produce a range of hillslope landforms (Benedict, 1976). Ice lenses in the soil supply water and create discontinuities that reduce the cohesive strength of the regolith and promote downslope mobility (Williams, 1959).

The excavation of sediment from a turf-banked terrace in August 1999 revealed no interstitial ground ice within the soil. However, freezing of the sediment during the winter cannot be ruled out. Alternatively, the terraces may have developed during a former period of cooler climate when gelifluction and frost creep processes were more prevalent.

Benedict (1976) classified solifluction features on the grounds of topographic expression, amongst other things. The lobes in Fossdalur possess steep, bulging, arcuate fronts and are typically overlapping. This mobile-looking appearance suggests that solifluction processes are still acting on these slopes. Future studies of soil movement

rates in this humid, alpine environment are needed to confirm the 'active' status of these periglacial features.

Patterned ground

Sorted stone patterns occur on various scales, particularly in Fossdalur and the surrounding hills. On level ground these patterns are equant and polygonal in form; on gently-sloping ground they are elongate; whilst on steeper slopes linear features predominate. The coverage of patterned ground in the study area is shown on the main geomorphic map (Fig. 6.2). The altitude range of these features in Fossdalur is from c. 400-1000 m. Sorted nets vary in diameter from 1.5 m polygons on *Gunnsteinsfjall* in upper Fossdalur to small features less than 0.5 m in diameter on *Vesturdalsheiði*. Well-developed stone stripes with a spacing of ~1 m were found on Dalsheiði. Some examples of patterned ground are shown in Figure 6.26.

Palaeo-environmental implications

The presence of patterned ground implies cryogenic processes are at work in the soil. The formation of segregation and needle ice allows small stones to be heaved vertically and displaced laterally in the soil (Washburn, 1956). Over longer periods, this frost heave results in sorting at the surface. Large stones become concentrated in zones at the edge of polygons, whilst the central zones consist of finer material largely free of stones. The features are known to form in areas where the sub-soil remains permanently frozen (Washburn, 1969, 1980), and are therefore often indicative of permafrost. However, patterned ground is also reported in areas where the ground thaws completely during the summer months (Caine, 1963; Ballantyne, 1987).

The diameter of the polygons, or spacing of the stripes, constituting the sorted network is believed to reflect the depth of freezing during the winter (Goldthwaite, 1976; Washburn, 1980). In Fossdalur, patterned ground is not believed to indicate perennially frozen ground, rather more marginal conditions where the surface soil may freeze during the coldest months and thaw completely in summer. Most of the stone patterns observed are believed to be presently active, especially those in higher altitude

localities. However, it is probable that these features developed during cooler periods over the last several millennia. Patterned ground formation in Iceland may have been accelerated in recent centuries during the so-called *Little Ice Age* – when soil probably remained frozen for longer, more often. Using lichenometry, Cook-Talbot (1991) found sorted stone polygons in the Jotunheim mountains of Norway were inactive, and had remained largely undisturbed for over a century. A similar study in upland Iceland – such as in the Lambatungnajökull area – would be of great palaeo-environmental interest. Presently, the relative degree of patterned ground 'activity' in Fossdalur, and other parts of maritime southeast Iceland, is unknown.

Landforms relating to mass movement

Geomorphological features resulting directly from mass-movement processes are found on the slopes surrounding Fossdalur. These *debris flows* are all situated on steep valley sides (generally 35-45°) or below bedrock faces. The flows have parallel lateral leveés, a depressed axial furrow and a terminal debris accumulation. The sediment eroded to form the furrow is subsequently re-deposited at the base of the slope forming a low, arcuate ridge or lobate mound (Fig. 6.27). The debris in these ridges is poorly-sorted and consists of clast-supported, gravel and cobble grade material. Boulders were found at the foot of the debris accumulation. A high degree of particle orientation was noticed in the lateral leveés.

Environmental interpretation

Debris flows can be triggered in numerous ways. In Fossdalur these features are probably the result of heavy rainstorms causing the rapid transportation of loose material downslope. In northern Scandinavia, Rapp and Nyberg (1981) attributed similar features to high intensity summer rainstorms. However in southeast Iceland optimum conditions for debris flow formation are in spring or late autumn when precipitation is greatest. It is worth mentioning that the role played by avalanching snow, particularly in upland areas (over 600 m asl) could be quite significant.

Figure 6.24 Turf-banked terraces in Fossdalur – evidence of solifluction.

(left) Spatial distribution of solifluction features (lobes and terraces) in Fossdalur.

The field of view of the photograph (above) is shown as a triangle. North is to the right.

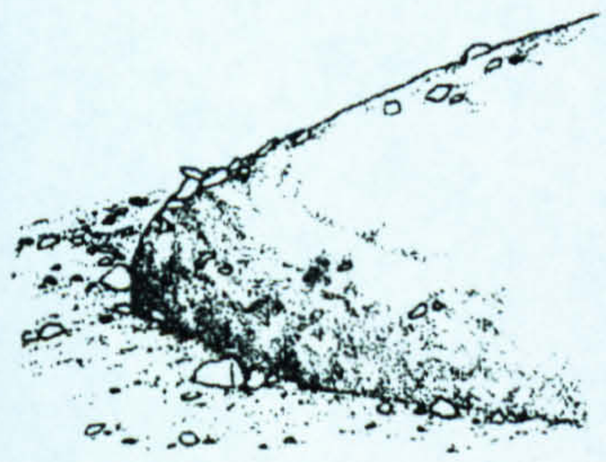
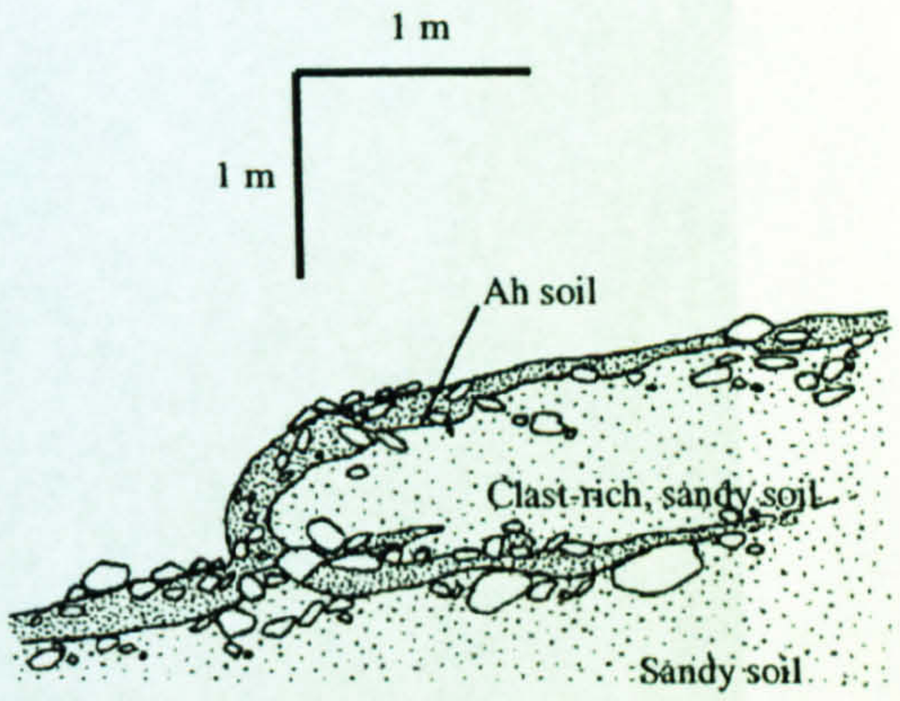
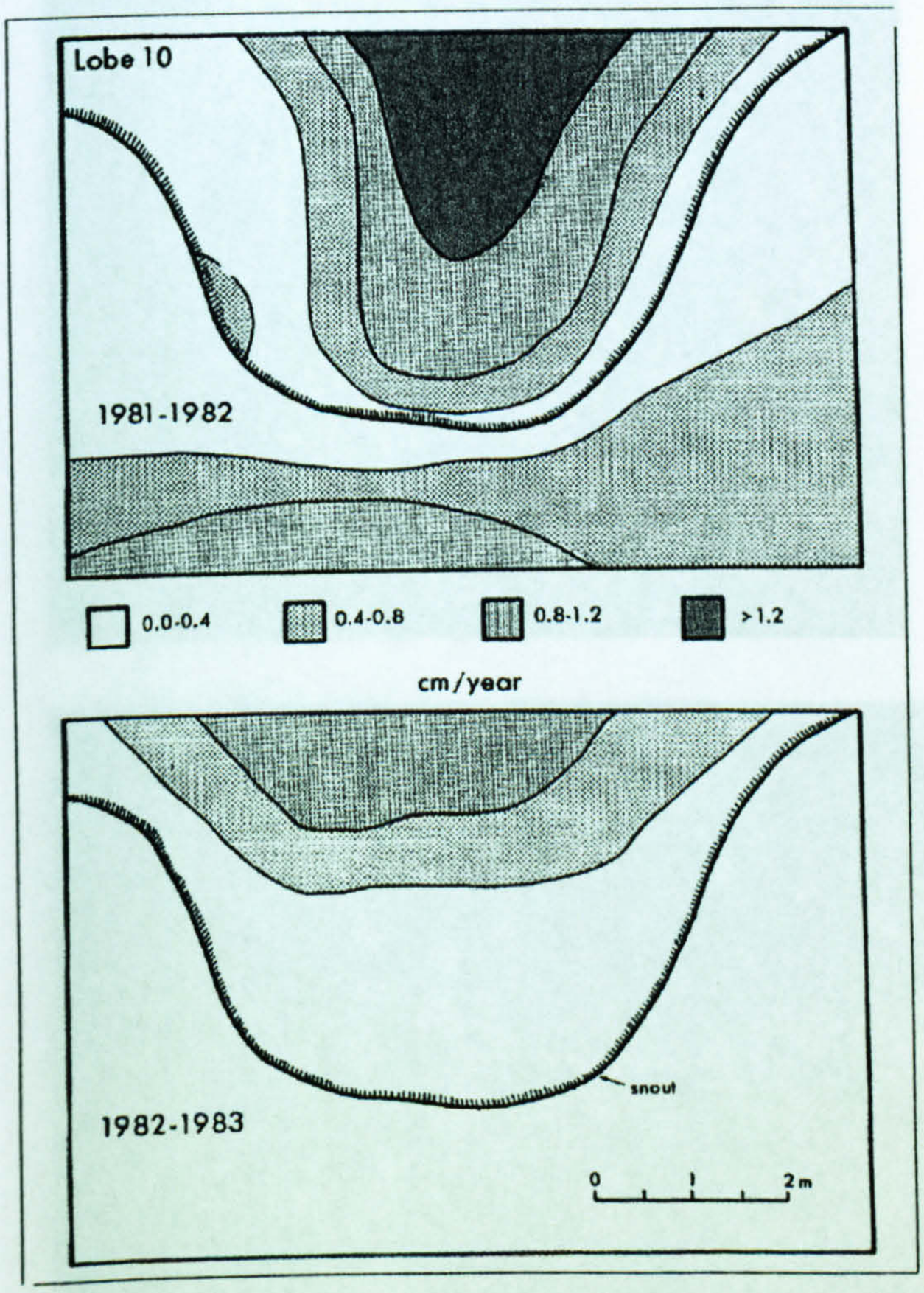


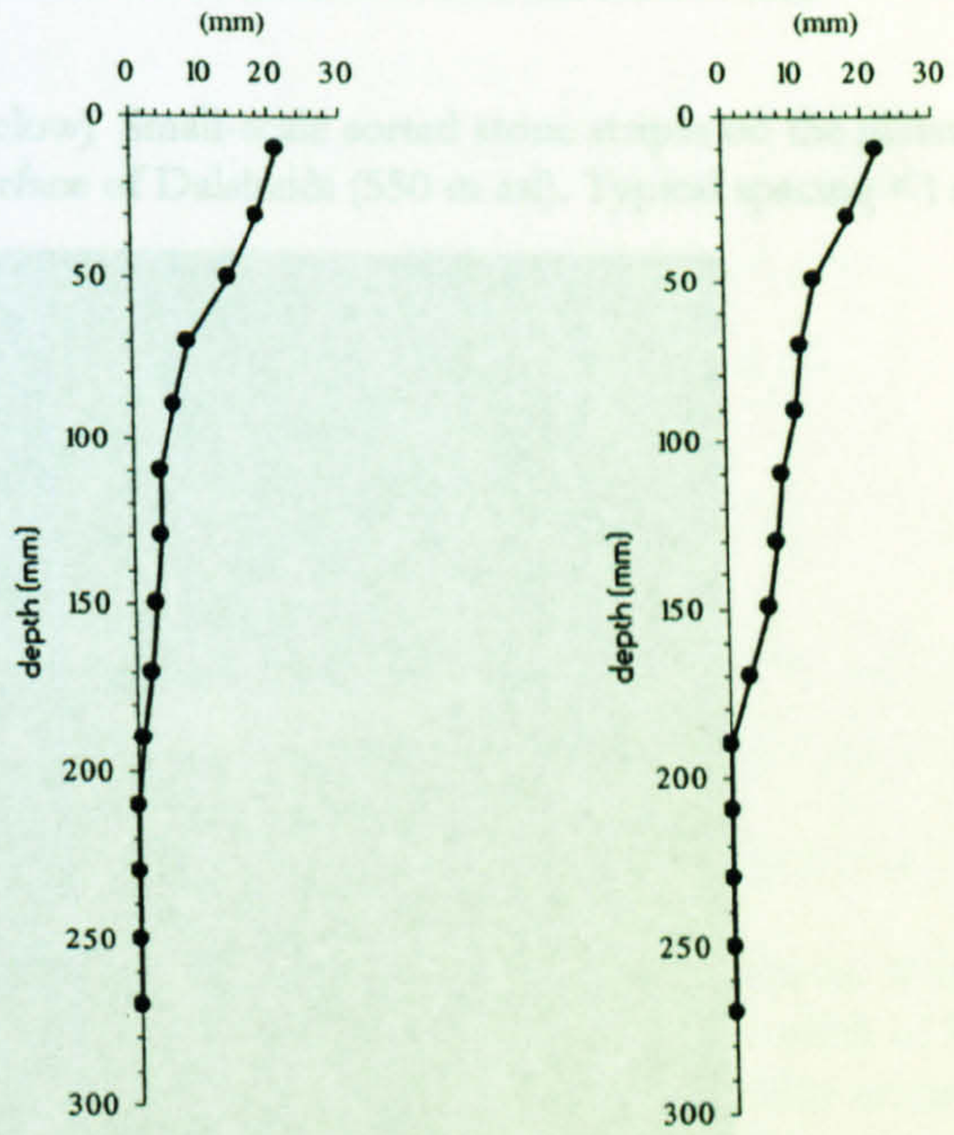
Figure 6.25 The morphology (left) and idealised sedimentology (below) of turf-banked solifluction terraces in Fossdalur, southeast Iceland. Both are to the same scale.



Modified from Benedict (1976)



(above) Observed variations in surface movement of turf-banked solifluction lobes in the Canadian Rocky Mountains (from Smith, 1988; Fig.5). Smith (1988) used Rudberg pillars and inclinometers in the soil to detect surface movement over a 2-year period.



(above) Rates of soil movement on two turf-banked terraces in Öræfi, southeast Iceland. A - Illuklettur; B - Skaftafellsheiði (from Douglas and Harrison, 1996).

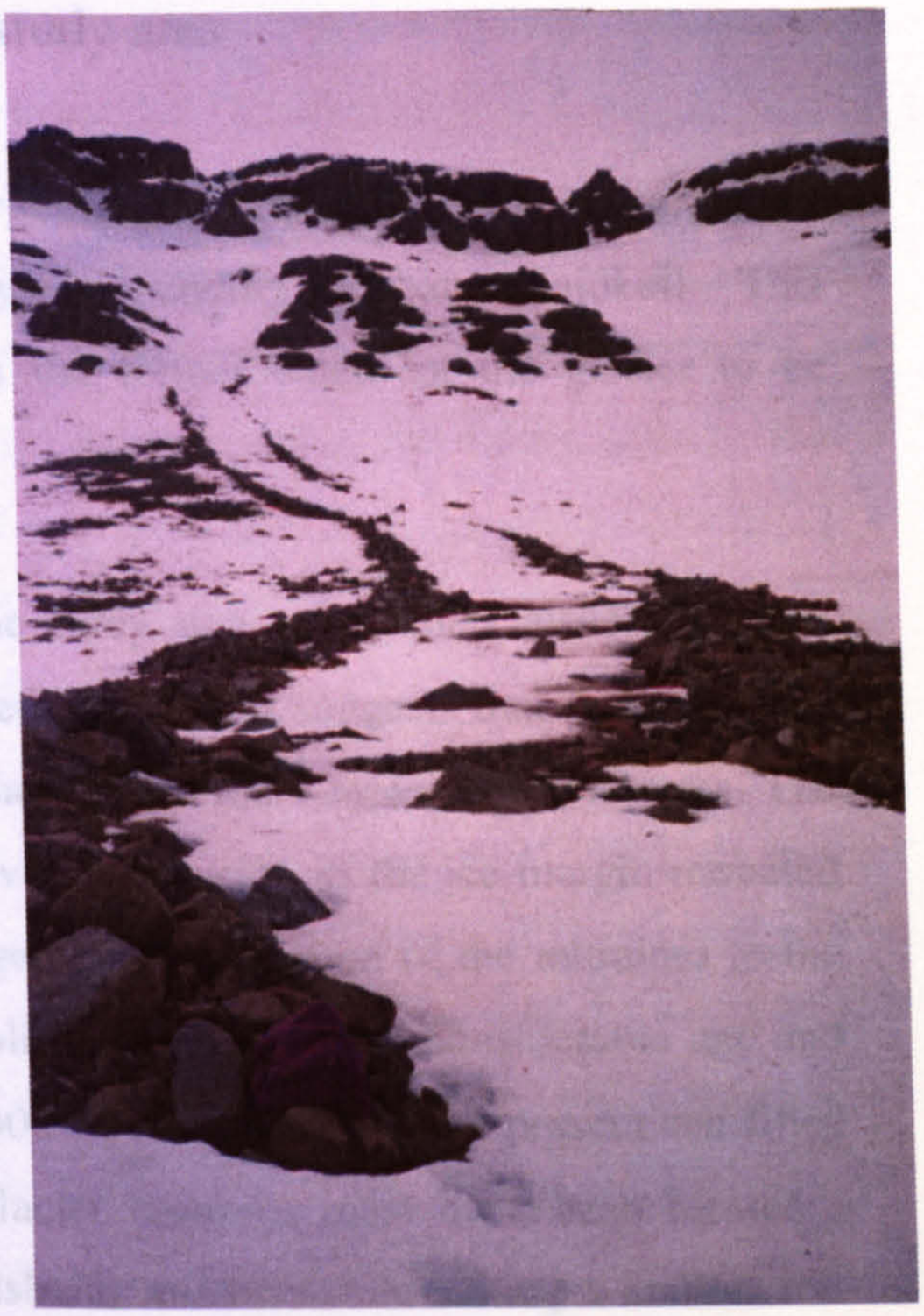
Figure 6.26 Patterned ground features

(left) Sorted stone polygons on the plateau surface of Gunnsteinsfjall, adjacent to Lambatungnajökull (700 m asl). Typical diameter of net 1.5 m.

(below) Small-scale sorted stone polygons in Fossdalur (600 m asl). Typical diameter 0.3 m.

(below) Small-scale sorted stone stripes on the plateau surface of Dalsheiði (550 m asl). Typical spacing <1 m.

Figure 6.27 Landforms resulting from mass movement



(right) Boulder tongue deposited by a debris flow. Located in Fossdalur at 700 m asl, the feature is c.100 long and only 5 m wide. Note the size of the debris relative to the rucksack (right). The photograph was taken in May 1999.

(left) Alpine debris flow located at an altitude of 2300 m asl., French Alps. The central channel has a width of 5-7 m. The flow is c. 150 m long. Note the similar morphology and sedimentology of this mass movement feature with that above. The photograph is taken from van Steijn (1988).

Interpretation of geomorphology in the study area

Moraines relating to numerous glacial advances or stillstands of the margin of Vatnajökull have been identified in the valleys surrounding Lambatungnajökull. This wealth of geomorphological evidence allows the former extent of the glacier to be reconstructed at various times in the past.

The highest altitude moraines identified in the study area are on the Dalsheiði plateau (S1) at 400-550 m asl. Their rounded, subdued appearance suggests that these features have been subjected to a long period of weathering and mass movement processes. The four discrete ridges represent a period of active deglaciation as the ice-margin retreated across the narrow plateau. Furthermore, the geomorphic relation of the moraines to the present-day snout of Lambatungnajökull implies that they are of considerable age and must date from a time when the glacier was 300-400 m thicker than at present and filled the valley entirely. At the same time the glacier terminus must have been located a considerable distance downvalley of the Dalsheiði moraines. Assuming a similar ice-surface profile in the ablation area to that of broad outlet glaciers currently draining from Vatnajökull, the palaeo-glacier probably terminated 10-15 km southeast of these lateral moraines (S1). The moraine fragment at *Hraunkot*, close to the coastline, is c. 14 km from the lowest extension of moraine S1a. The moraines on Dalsheiði have been tentatively correlated with this terminal moraine fragment and are believed to have formed during the same period of extensive glaciation

Moraine H1 in Hoffellsdalur is 7 km from the present-day margin of Lambatungnajökull. At an elevation of 50 m asl, slightly higher than the *Hraunkot* feature, this moraine represents the terminal position of Lambatungnajökull during a subsequent advance or stillstand. This is referred to as the *Hoffellsdalur I* stage.

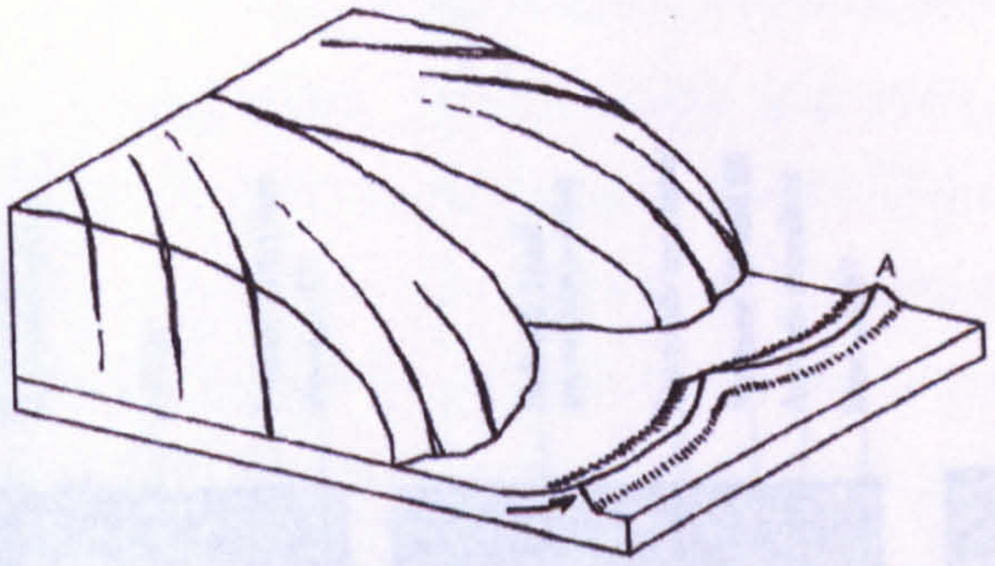
The next moraines encountered in Hoffellsdalur, H2 and H3, lie 2 km and 5 km north of moraine H1, respectively. The older termino-lateral moraine fragment, H2, at 100 m asl,

represents the *Hoffellsdalur II* stage of glaciation. H3, at 140 m asl, formed during a subsequent glacier advance or stillstand referred to as the *Hoffellsdalur III* stage.

The altitude of lateral moraines S2 and S3 in Skyndidalur suggest that they are considerably older than the landforms lying lower down the valley side and on the valley floor. The long-profile of these sub-parallel moraines suggests that at the time of their formation the glacier terminated approximately halfway down the valley, placing the terminus at an altitude of between 100-150 m asl. Moraine S2 is tentatively correlated with moraine H2 and moraine S3 is correlated with moraine H3 on the basis of altitude and distance from the present-day glacier terminus.

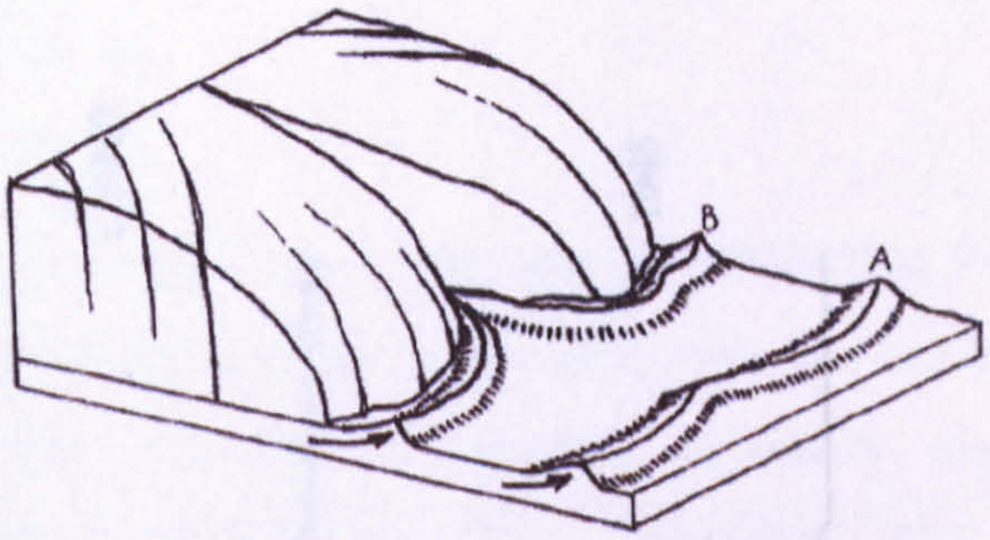
Younger moraines occur closer to the head of Hoffellsdalur. Moraines H4-H12 are well defined, relatively steep-sided, poorly vegetated features suggesting that they formed within the late Holocene. The two, arcuate, termino-lateral moraine fragments H4 and H5, at an altitude of 170 m asl, represent stages *Hoffellsdalur IV* and *V*. The outermost, poorly vegetated, steep-sided lateral moraine in Skyndidalur, S4, correlates with moraine H4 in Hoffellsdalur. The largest moraine at the head of Hoffellsdalur (H8) is similar in size and appearance to lateral moraine S6 in Syndidalur suggesting that they formed coevally.

A closely-spaced sequence of arcuate, low-elevation, moraines in the pass between Hoffellsdalur and Skyndidalur (H9-H12) resembles the pattern of ridges on the floor of Skyndidalur between moraines S7 and S9. Each of these minor moraine ridges probably formed by squeezing of underlying sediment during small-scale, possibly annual, fluctuations of the ice-margin. Immediately upvalley of the S8-S9 moraine complex is a further sequence of minor moraines. These *sawtooth* features probably formed by the same mechanism as the *arcuate* minor moraines and are thought to represent the former position of a highly crevassed ice-margin (Fig. 6.28). The difference in moraine morphology is a reflection of the crevassing that occurred at the ice-margin as the glacier flowed over a bedrock knoll (Figs. 6.29). Aerial photographs show that 13 of the sawtooth moraines had formed by AD 1945. It is likely that the minor arcuate moraines in Skyndidalur date from the early part of the 20th century.



Plan view

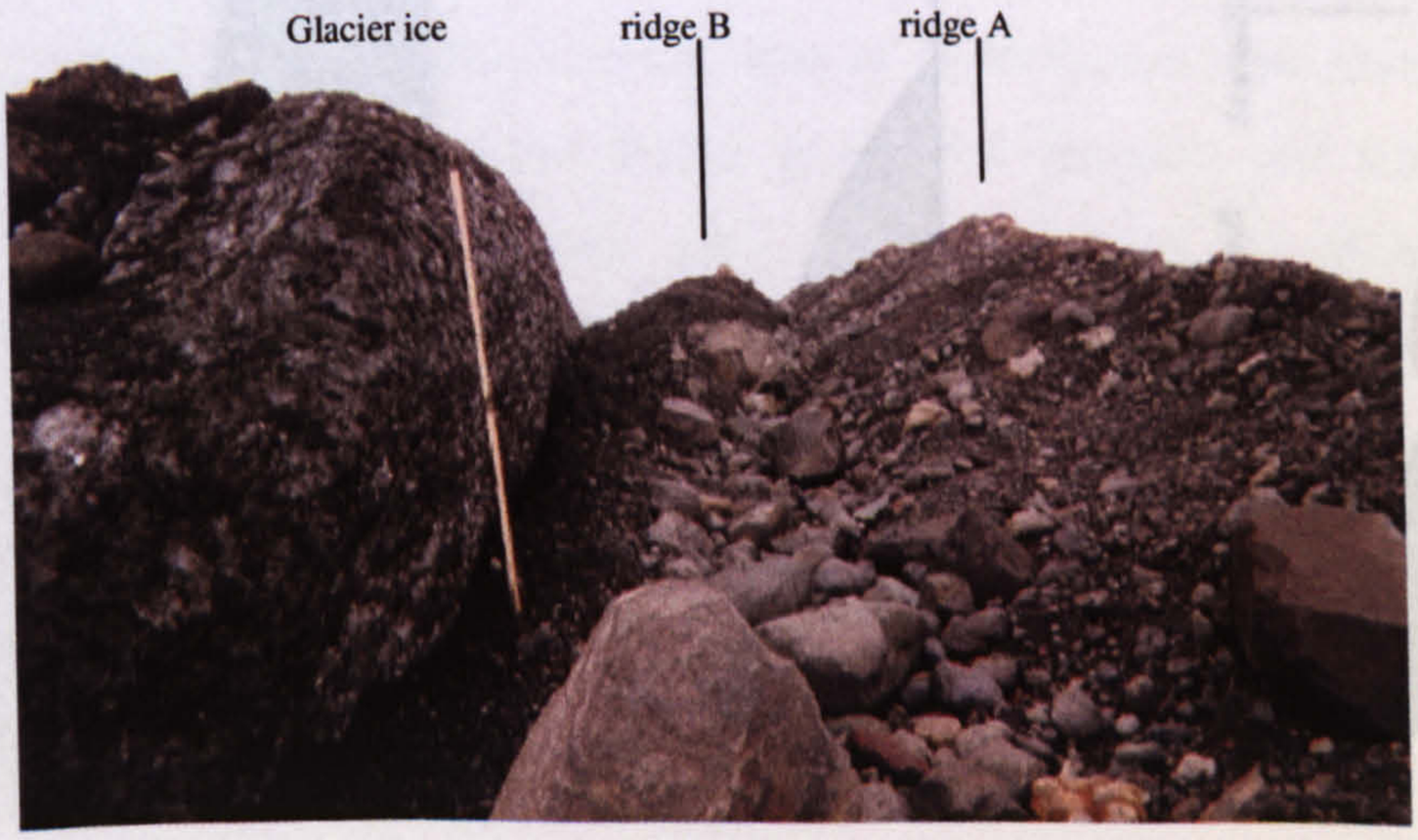
Late summer (following maximum retreat)



Plan view

Late spring (following winter re-advance)

Figure 6.28 (above) Minor, Saw-tooth, push moraines at Lambatungnajökull. Ridge A is formed during a slight winter advance. Ridge B is formed the following winter. Note how the moraine morphology mirrors the outline of the crevassed ice margin



(left) Minor push moraines forming at the margin of Fjallsjökull, southeast Iceland. Soft, deformable sediment is squeezed from beneath the toe of the glacier during small advance events. The photograph was taken in May 1999. Ridge B is still in contact with the glacier margin; ridge A was probably formed the previous winter. Note metre-stick for scale.

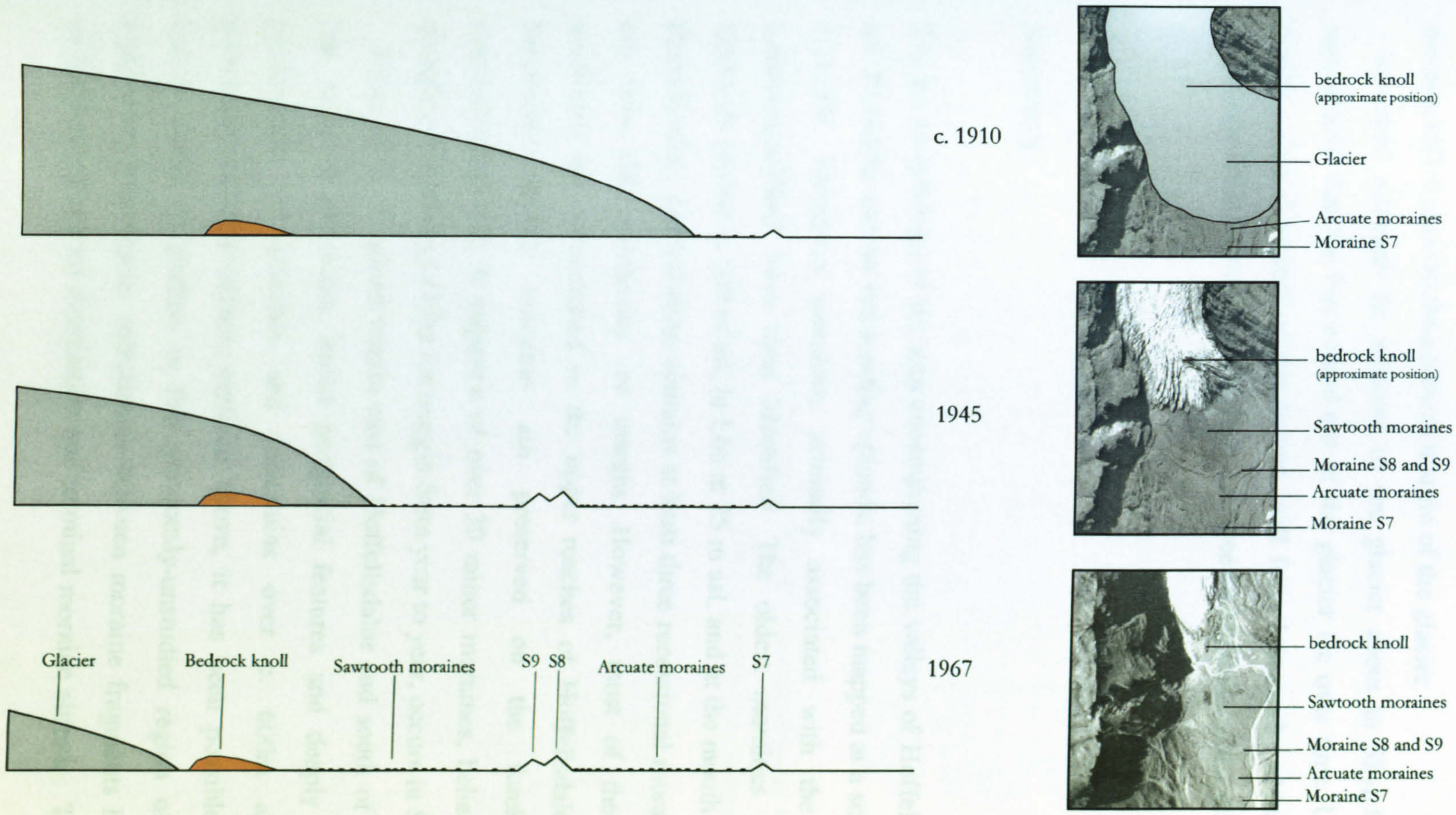


Figure 6.29 The relation between moraine morphology and glacier snout dynamics.

In the early 20th century, prior to c. 1930, the margin of Lambatungnajökull was free of crevasses and arcuate in plan (uppermost panel). By 1945 the ice-margin had become intensely crevassed and possessed a sawtooth outline. The distinctive zigzag morphology of the moraines deposited during this time reflects the crevassing that occurred as the ice flowed over a bedrock knoll (middle panel). As the glacier margin retreated the influence of the bedrock knoll was no longer felt by the glacier. By 1967 the glacier had completely revealed the bedrock feature. The glacier margin has been relatively free of crevasses and arcuate in plan since this time.

Moraine S10 next to the bedrock knoll in Skyndidalur represents a recent readvance of Lambatungnajokull between 1957 and 1967. This feature correlates with a push moraine (H13) close to the southern margin of the glacier.

Ice-cored moraines lie adjacent to the glacier snout in Skyndidalur. Most of the supraglacial material has melted out of the glacier ice over time. Ultimately the debris mantle, when deposited on the ground, will form hummocky *ablation moraine*. These ridges represent the most recent minor advance or stillstand of the glacier margin.

Summary

The geomorphology of the area encompassing the valleys of Hoffellsdalur, Skyndidalur and Fossdalur and the surrounding uplands, has been mapped at a scale of approximately 1:11,000. Numerous moraines, primarily associated with the former extent of Lambtungnajökull have been identified. The oldest moraines are located on the Dalsheiði plateau c. 500 m asl, in Lón at 25 m asl, and at the mouth of several valleys in Hornafjörður. Hoffellsdalur contains at least three recessional moraines at c. 7, 5, and 2 km from the present-day ice margin. However, most of the glacially-deposited landforms are concentrated in the upper reaches of Hoffellsdalur and Skyndidalur. Numerous smaller moraines are preserved on the foreland in front of Lambatungnajökull. A sequence of over 30 minor moraines, believed to represent the changing morphology of the ice margin from year to year, occurs in Skyndidalur

Large areas of upland terrain east of Hoffellsdalur and south of Skyndidalur exhibit few signs of glaciation. Relict periglacial features and deeply weathered bedrock predominate on plateaux and mountains over c. 600m asl. Using all the geomorphological evidence, outlined above, it has been possible to reconstruct the former extent of glaciers in this previously-unstudied region of southeast Iceland. Preliminary geomorphic correlations between moraine fragments have been suggested on the basis of spatial distribution and terminal moraine altitude. The following glacial

stages are identified. The extent of glaciers during these periods is reconstructed in Figures 6.30 – 6.39.

Table 6.1 Summary of glacier stages at Lambatungnajökull identified in this research.

Advance stage	*Moraines*		*Altitude of terminal moraines (m)*	*Palaeo-map*
	Hoffellsdalur	Skyndidalur		
Dalsheiði	–	S1	25	Figure 6.30
Hoffellsdalur I	H1	–	50	Figure 6.31
Hoffellsdalur II	H2	S2	100	–
Hoffellsdalur III	H3	S3	140	Figure 6.32
Hoffellsdalur IV	H4	S4	170	Figs 6.33, 6.34
Hoffellsdalur V	H5	–	170	Figure 6.35
Hoffellsdalur VI-VIII, Skyndidalur VII - X	H6 - H13	S5 - S10	>180	Figs. 6.36 - 39

The use of tephra layers and slow-growing lichens, examined in the following two chapters, will help to constrain the timing of these glacier fluctuations.

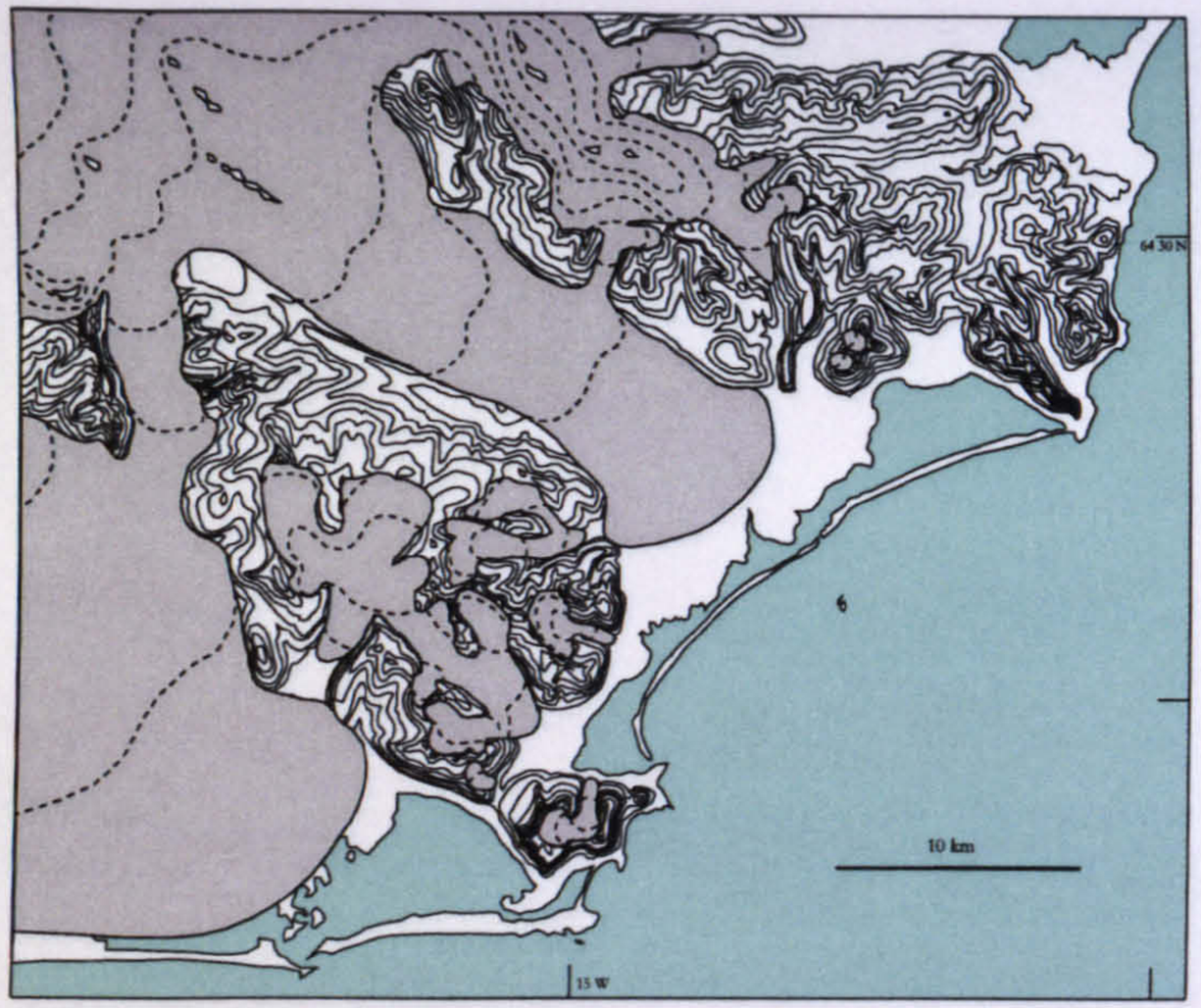


Figure 6.30 *Dalsheiði* stage

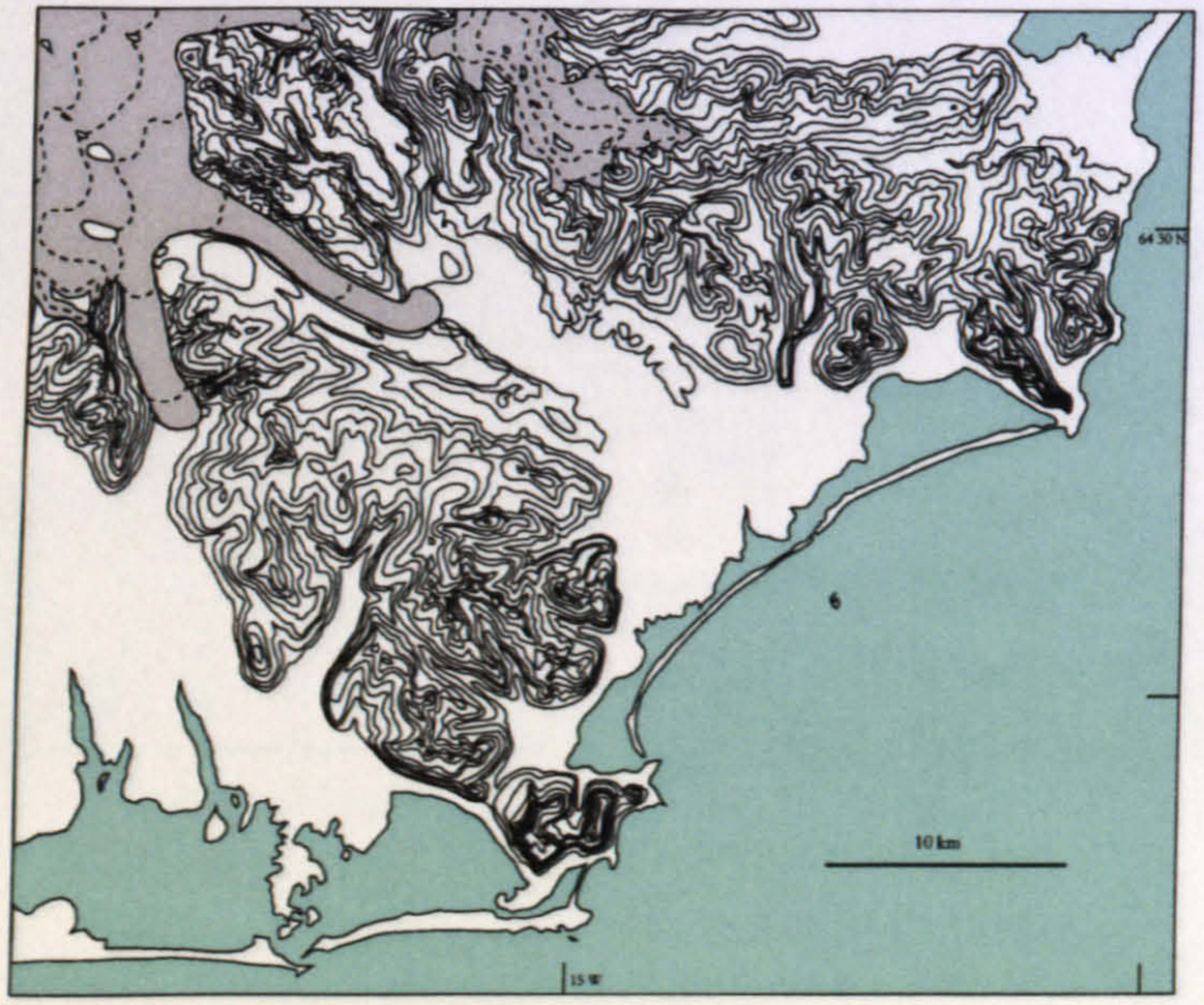


Figure 6.31 *Hoffellsdalur I* stage

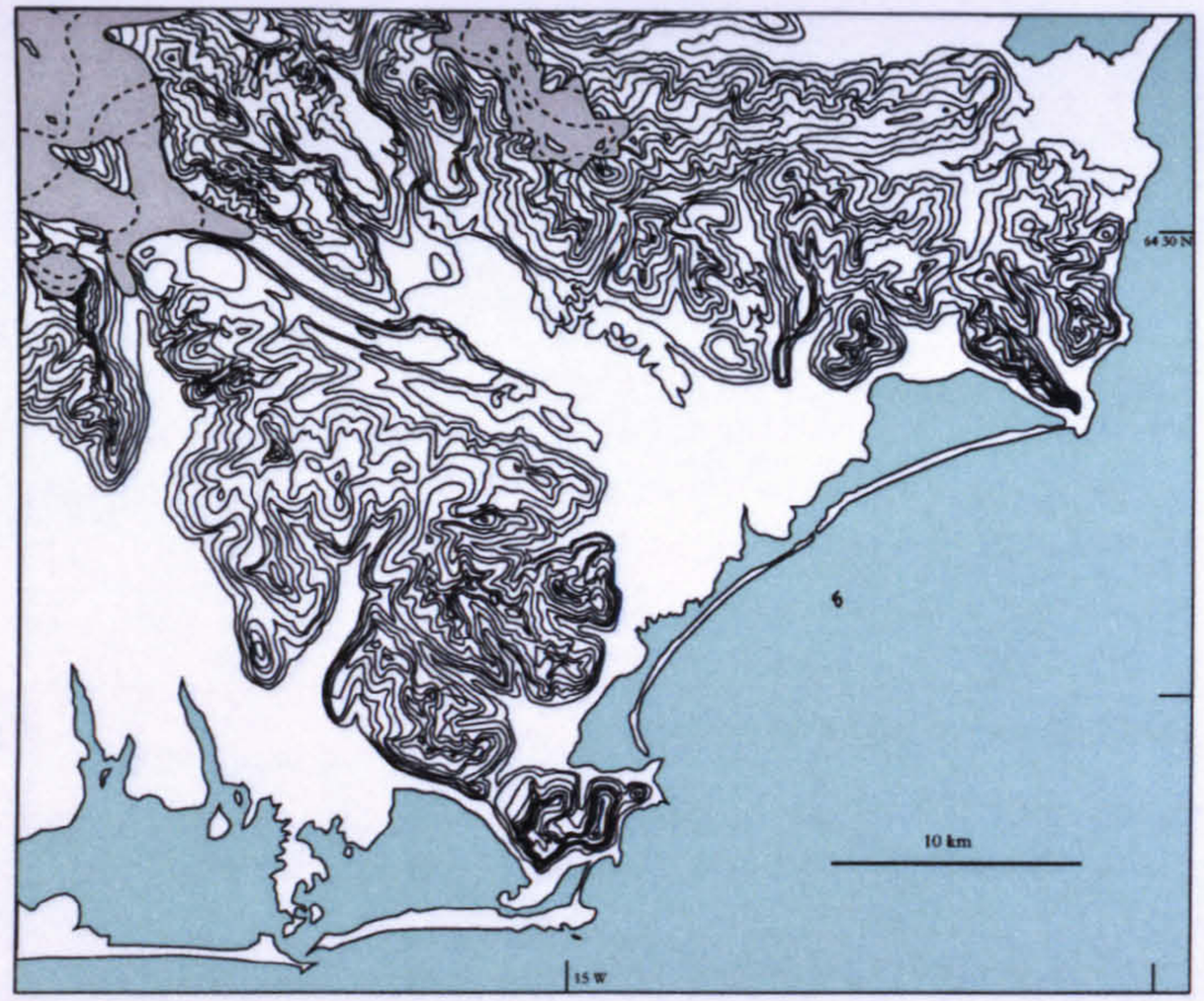


Figure 6.32 *Hoffellsdalur III* stage

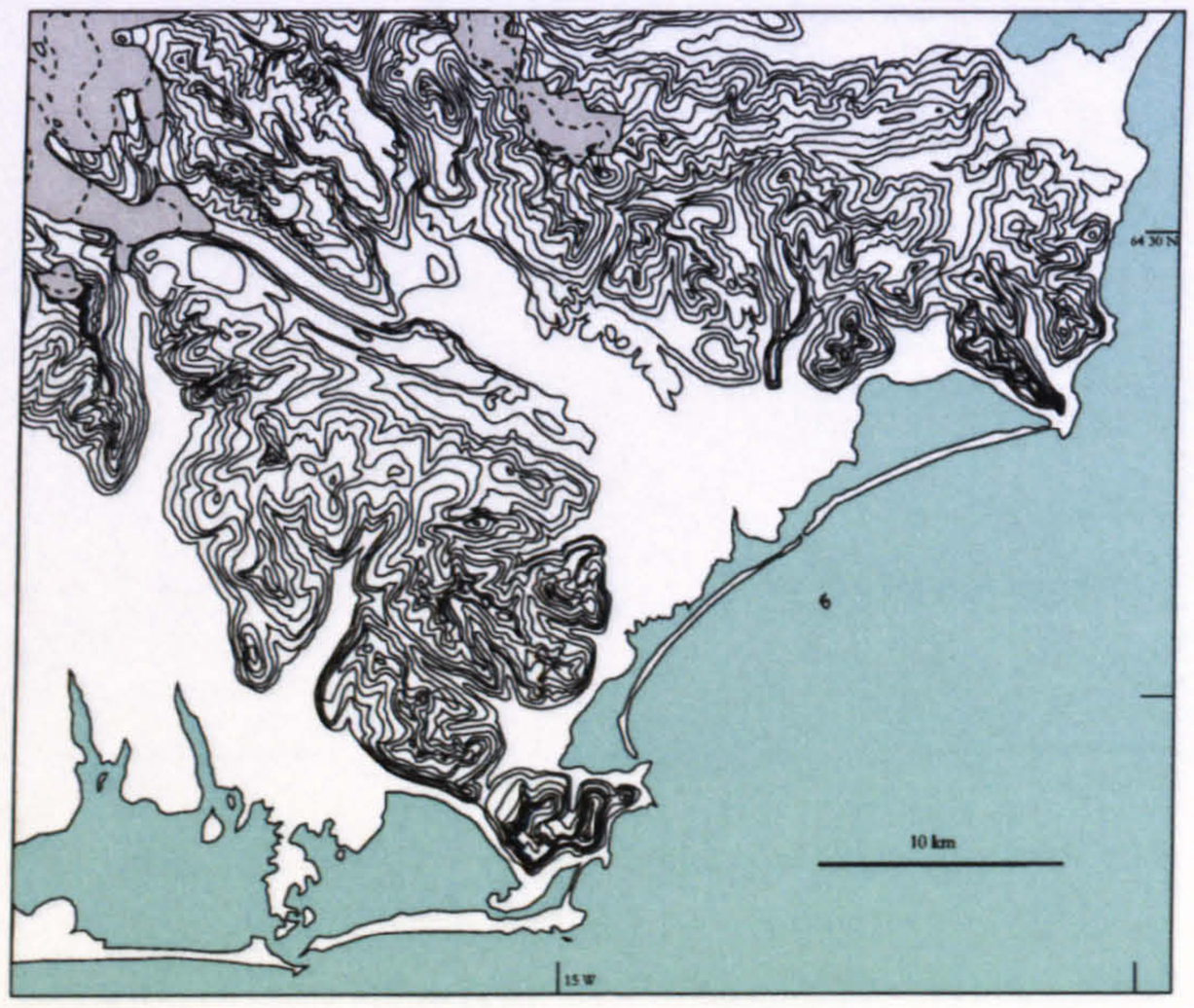


Figure 6.33 *Hoffellsdalur IV* stage

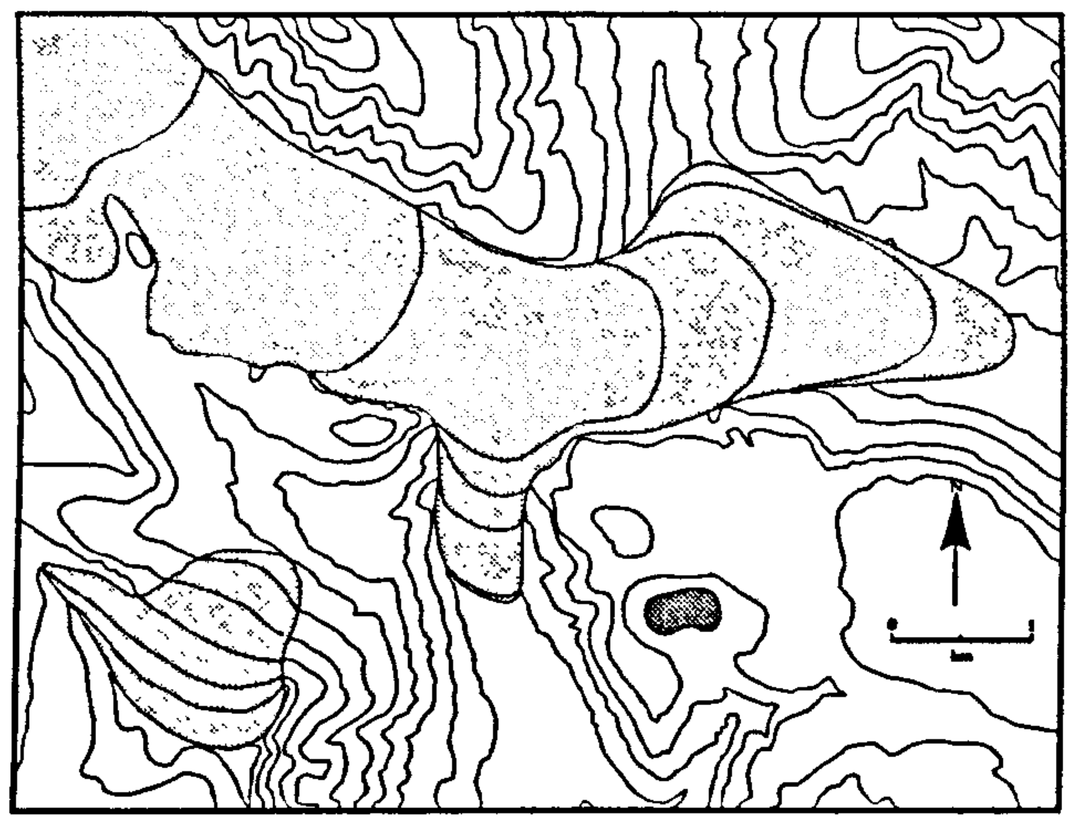


Figure 6.34 *Hoffellsdalur IV* stage

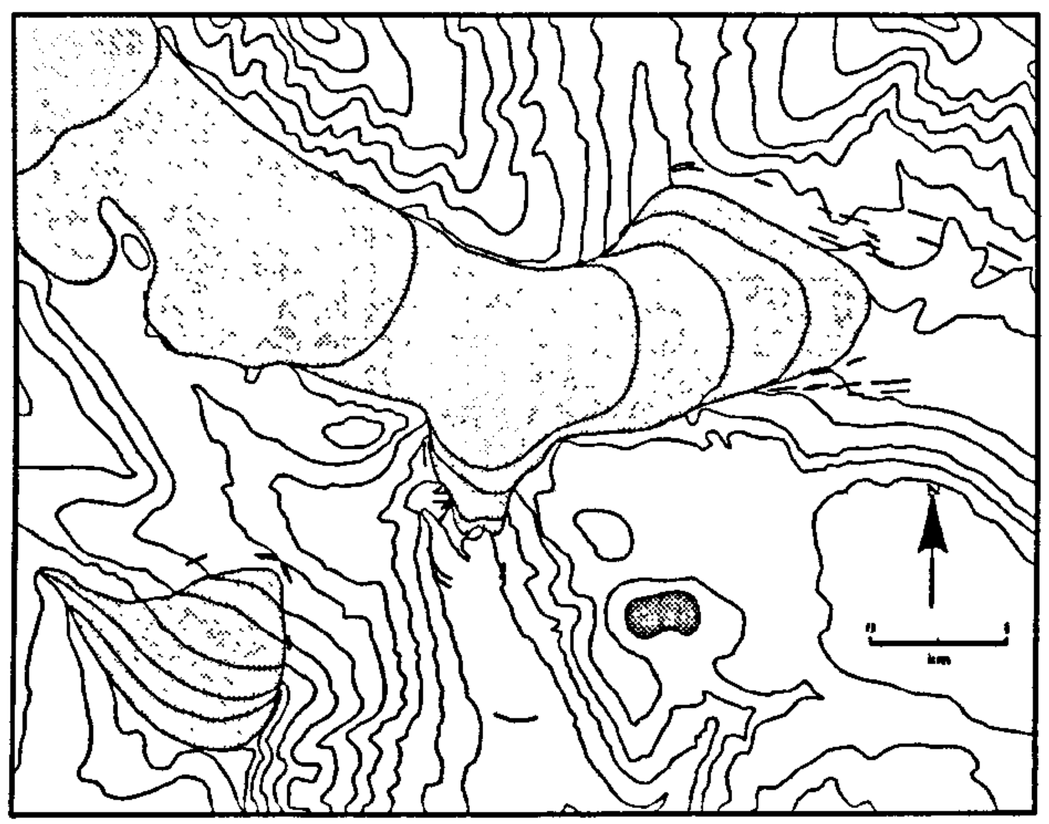


Figure 6.36 *Hoffellsdalur VI* stage

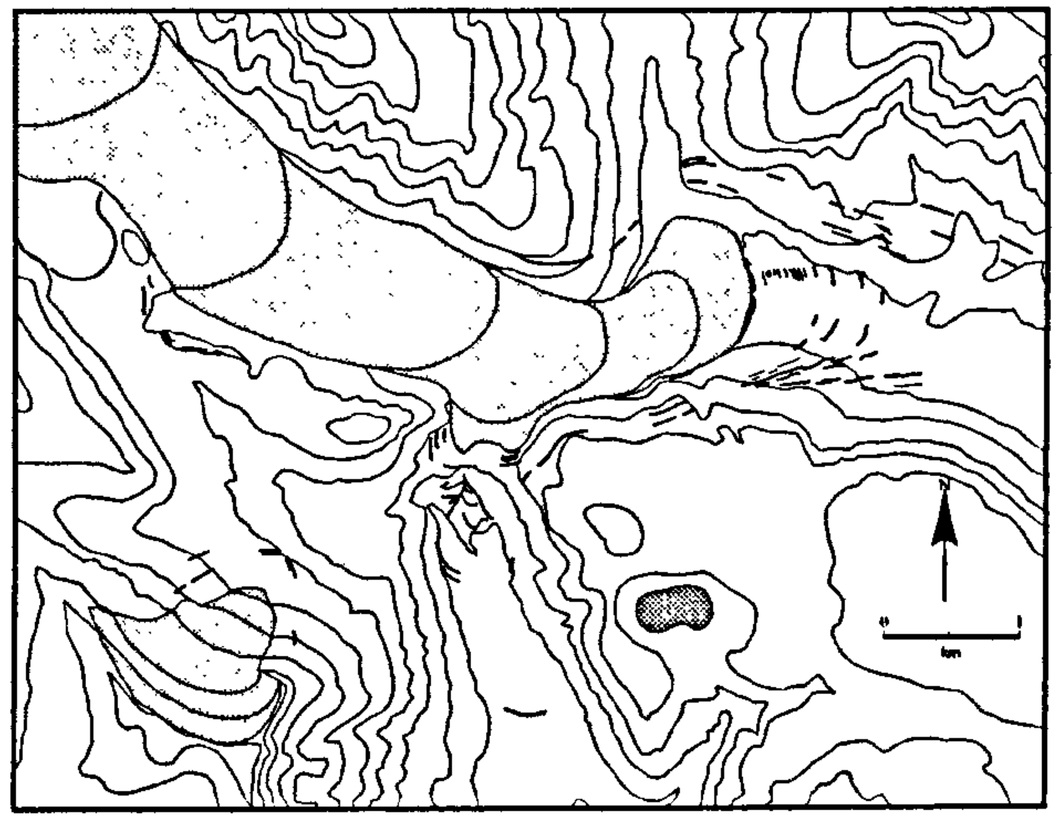


Figure 6.38 *Skyndidalur VIII* stage

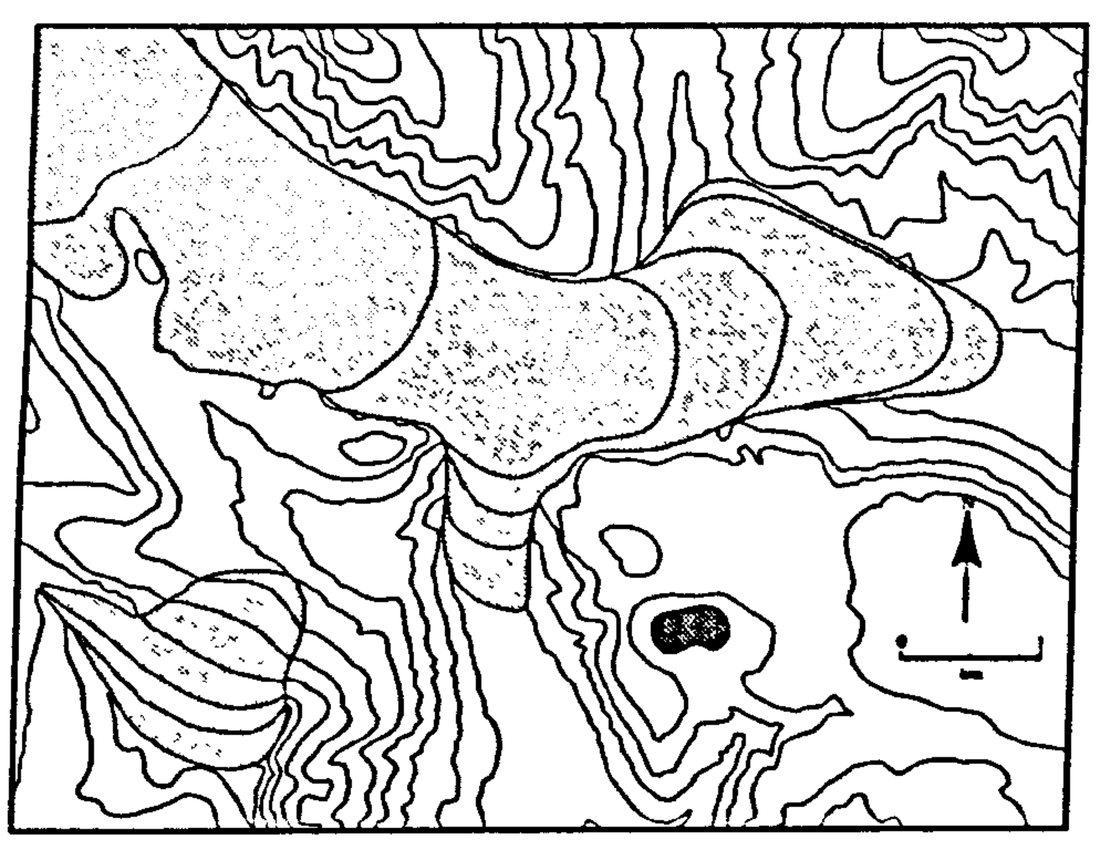


Figure 6.35 *Hoffellsdalur V* stage

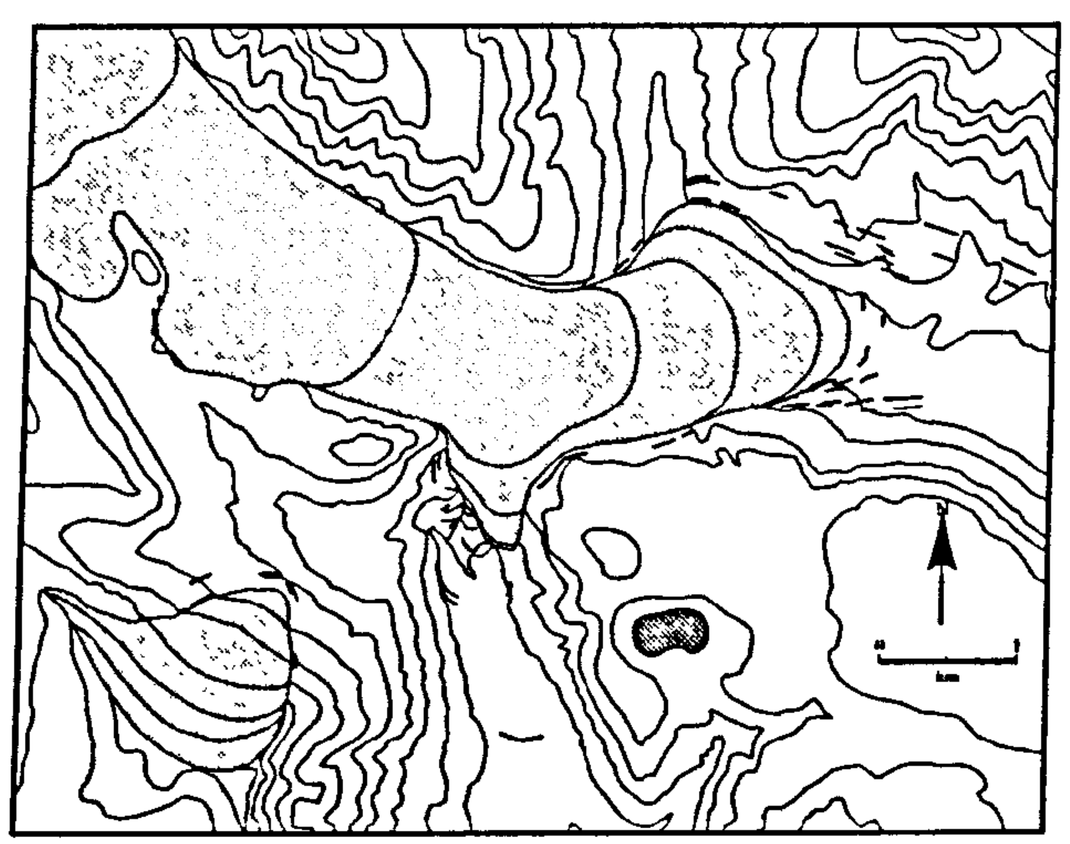


Figure 6.37 *Hoffellsdalur VIII* stage

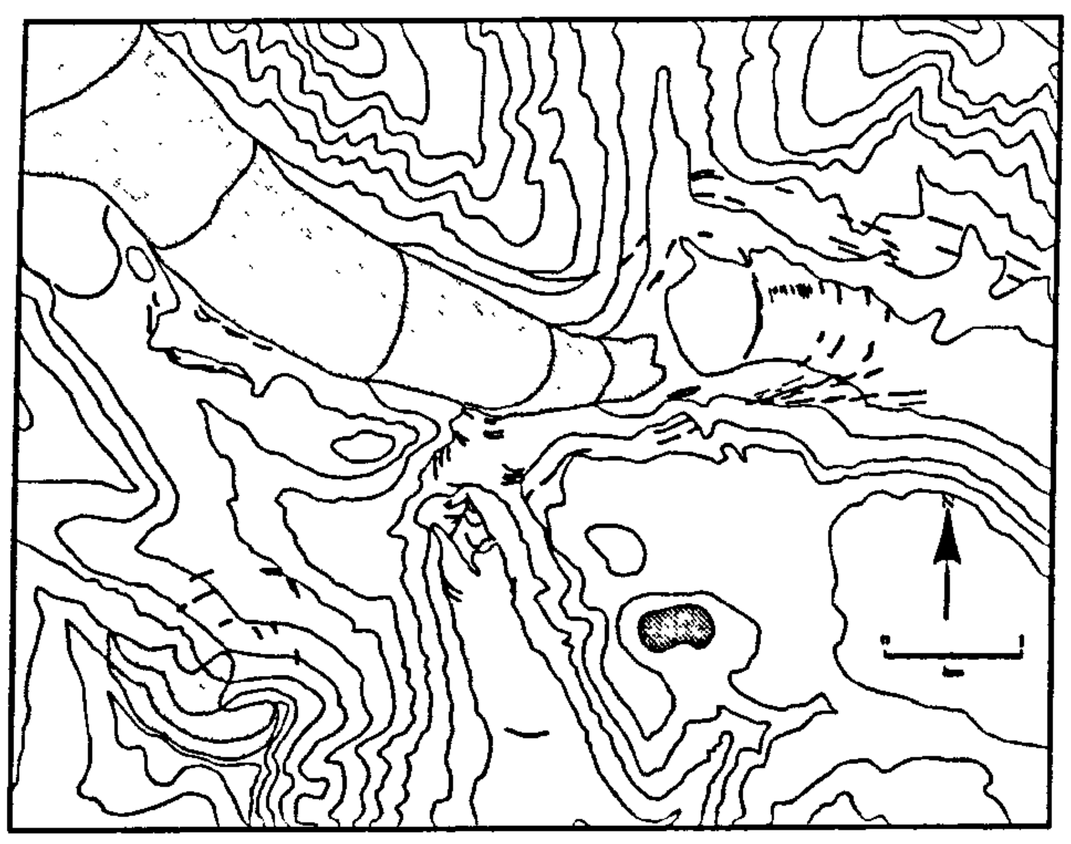


Figure 6.39 *Skyndidalur IX* stage

Chapter 7 : Results

Tephrochronological studies

Introduction

Tephrochronology was employed in order to reconstruct the deglacial chronology of Lambatungnajökull. Volcanic ash or *tephra* layers form important marker horizons in the soil. These ash layers, when linked with known eruptions, form *isochrones*, or surfaces of equal age, which can be used to date various landforms and hence help determine the timing of glacier fluctuations. A regional stratigraphic framework was developed for the area around Lambatungnajökull and enabled dating based on the presence or absence of certain tephra layers in the soil. Historically-documented ash-falls and a handful of ^{14}C-dated prehistoric eruptions, coupled with a knowledge of soil accumulation rates, have allowed the age of various features to be bracketed. Further, more detailed, tephrochronological studies should use geochemical 'fingerprinting', as first used in Iceland by Larsen (1978) and later by Dugmore *et al.* (1995) and Gudmundsson (1998a), to confirm the stratigraphic connections inferred in this study.

The tephrastratigraphy of Hornafjörður and Lón

The following discussion on the tephrastratigraphy of the Hornafjörður-Lón area has been divided into two parts. The 1362 eruption of Öræfajökull is the most easily identifiable marker horizon in the soil of southeast Iceland and for the purpose of this study is taken to represent the boundary between the historic and pre-historic (pre-settlement) soil sequences. The *Landnám*, or 'Settlement' tephra, erupted from Veiðivötn in c. AD 900 and commonly used to distinguish between the historic and pre-historic soil sequences (Thorarinsson, 1944; Dugmore, 1989) in Iceland, is often absent

in eastern Iceland. The Öræfajökull-1362 ash is a suitable isochrone for dating purposes, its stratigraphic thickness and spatial extent being relatively well known (Thorarinsson, 1956; Gudmundsson, 1998a).

The tephrastratigraphy of Hoffellsdalur and Skyndidalur has not been studied prior to this research. In all, 42 soil profiles were logged over an area of c.150 km^2 (Fig. 7.1). The tephrastratigraphy of the study area is summarised in Figure 7.2. The presence of certain tephra layers enables the **minimum** age of the soils, and hence various landforms, to be ascertained.

Eight tephra layers have been identified in the soil overlying the Ö-1362 ash (Fig. 7.2). The origin and age of these tephras is based on field observations of layer thickness, colour, particle size and stratigraphic position. These characteristics have been compared with published studies conducted elsewhere in southeast Iceland (eg. Thorarinsson, 1956; Gudmundsson, 1998a).

Based on the identification of two silicic tephra layers in selected soil profiles, typical sediment accumulation rates in Hoffellsdalur and Skyndidalur were approximately 0.68 mm a^{-1} between AD 1362 and AD 1875. This figure is comparable with the figure of 0.59 mm a^{-1} derived by Gudmundsson in Öræfi, 75 km to the southwest. However, soil accumulation rates in both these studies are considerably lower than that derived by Dugmore (1987) for Eyjafjallsveit, in southern Iceland. Such high rates of aeolian sedimentation are probably a function of the wet, windy conditions, the large numbers of grazing livestock and the unvegetated expanse of sandur surrounding Dugmore's study area. The small number of people and relative lack of farmland in southeast Iceland probably make soil accumulation rate changes over the last ~1000 years a reflection of natural environmental variability more than the pressures of human settlement.

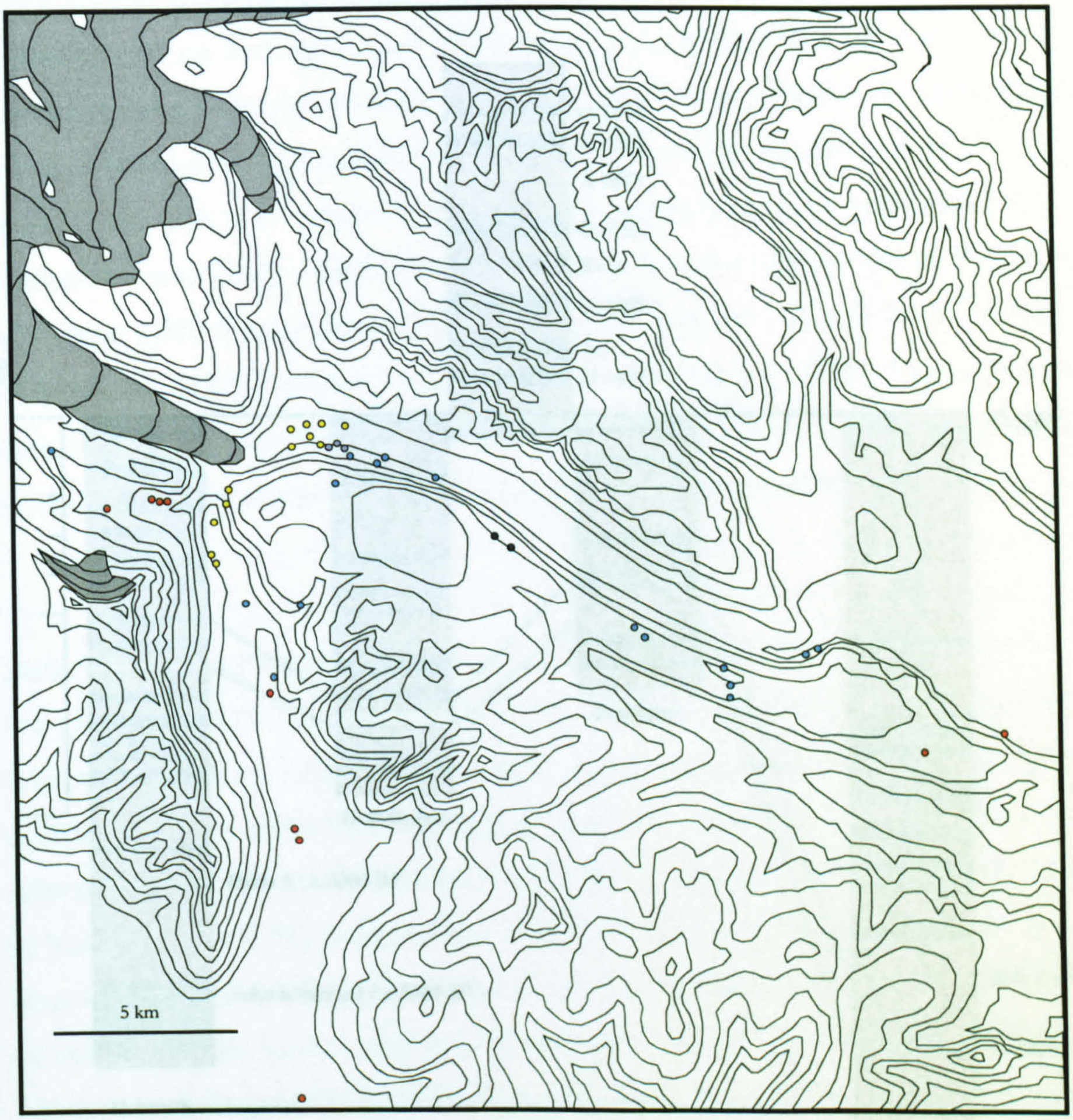


Figure 7.1 Location of 42 soil pits dug during the tephrastratigraphic investigations in 1998 and 1999. Contours are at 100 m vertical intervals. Glaciers are shaded grey.

The lowermost visible silicic tephra layer in each soil pit is shown:

- No silicic tephras visible
- Askjá 1875
- Öræfajökull 1362
- Hekla 4
- Hekla 5

Note that the area to the north of the river *Skyndidalsá* was inaccessible.

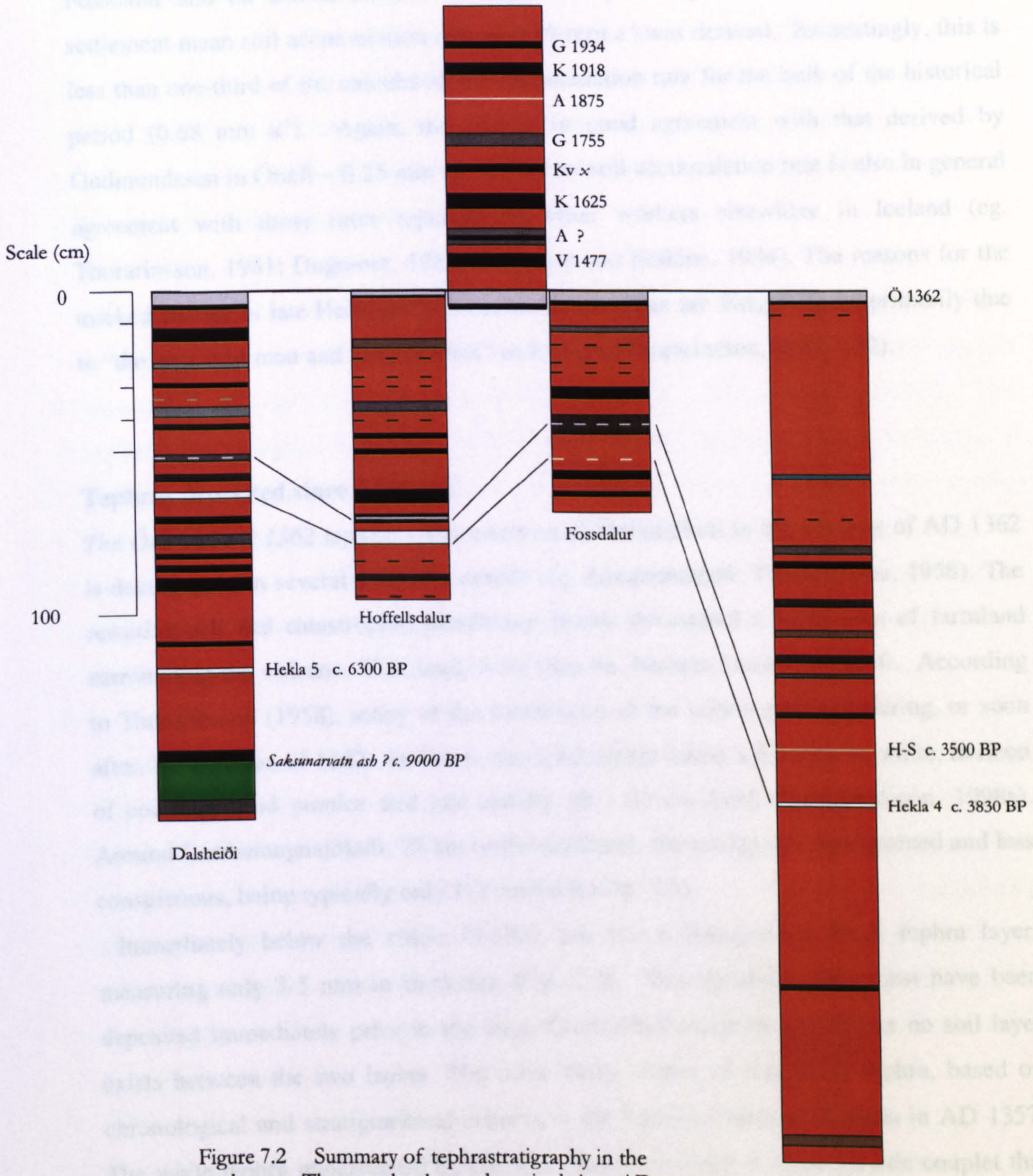


Figure 7.2 Summary of tephrastratigraphy in the study area. Those tephra layers deposited since AD 1362 can be correlated across most of the area. The prehistoric (pre-1362) tephrastratigraphy is more complicated and shows a high degree of spatial variability.

Based on the presence of silicic Hekla-4 tephra, in soil sequences in Hoffellsdalur, Fossdalur and on Dalsheiði, and Öræfajökull tephra deposited in AD 1362 a pre-settlement mean soil accumulation rate of 0.19 mm a^{-1} was derived. Interestingly, this is less than one-third of the calculated soil accumulation rate for the bulk of the historical period (0.68 mm a^{-1}). Again, the rate is in good agreement with that derived by Gudmundsson in Öræfi – 0.25 mm a^{-1}. This low soil accumulation rate is also in general agreement with those rates reported by other workers elsewhere in Iceland (eg. Thorarinsson, 1981; Dugmore, 1987; Dugmore and Erskine, 1994). The reasons for the marked change in late Holocene soil accumulation rates are thought to be primarily due to "the arrival of man and his livestock" in Iceland (Thorarinsson, 1981: 122).

Tephras deposited since AD 1362

The Öræfajökull 1362 tephra – The eruption of Öræfajökull in the summer of AD 1362 is documented in several Icelandic annals (eg. *Landnámabók*; Thorarinsson, 1958). The resulting ash and catastrophic *jökulhlaup* floods devastated a large area of farmland surrounding the volcano. This land, from then on, became known as Öræfi. According to Thorarinsson (1958), many of the inhabitants of the region perished during, or soon after, the eruption of 1362. In Öræfi, the 1362 tephra forms a distinctive, white, horizon of coarse-grained pumice and ash usually 10 - 20 cm thick (Gudmundsson, 1998a). Around Lambatungnajökull, 75 km to the northeast, the ash layer is fine-grained and less conspicuous, being typically only 1-2 cm thick (Fig. 7.3).

Immediately below the silicic Ö-1362 ash lies a fine-grained black tephra layer, measuring only 3-5 mm in thickness (Fig. 7.3). This basaltic tephra must have been deposited immediately prior to the large Öræfajökull eruption of 1362 as no soil layer exists between the two layers. The most likely source of this black tephra, based on chronological and stratigraphical criteria, is the basaltic eruption of Katla in AD 1357. The white tephra underscored by the thin black layer form a characteristic couplet that can be identified with confidence across the entire study area.

V 1477 – The first layer above the Ö-1362 marker horizon is a medium- to fine-grained black tephra, 1-2 cm in thickness. Located typically 7-10 cm above the silicic layer, extrapolation of mean soil accumulation rates (0.68 mm a^{-1}) places this basaltic tephra in the period from c. AD 1465-1510. A major eruption of Veiðivötn took place in AD 1477 (Thorarinsson, 1975; Larsen, 1984). This tephra was dispersed across much of northern and eastern Iceland and has previously been identified in other southeastern localities (Gudmundsson, 1998a: 84).

Unknown tephra (A) – Only 2-3 cm above the previous layer lies a tephra layer comprising three horizons: the lowermost fine-grained and black, the second slightly coarser and grey, and the uppermost similar to the first, fine-grained and black. This tripartite tephra layer is unusual yet distinctive and can be traced over a wide area. It probably represents three phases of a single eruption, rather than three discrete events. Soil accumulation rates place the time of eruption between c. AD 1520 and 1540. The origin of this ash is unknown, although the Katla and Grimsvötn complexes are the most likely sources based on the mineralogy and colour of the volcanic glass shards. Ash of similar age was identified by Gudmundsson (1998a: 85) in Öræfi.

K 1625 – 5 cm above the unknown tephra layer (*A*) lies another fine-grained, black tephra. The thickness of this deposit (2-3 cm) suggests a large or relatively proximal eruption. A date of c. AD 1600 – 1630 has been derived by the extrapolation of soil accumulation rates. An eruption of Grimsvötn took place in July 1619 with tephra being reported to have fallen in Norway and Faeroe (Thorarinsson, 1974). However, the tephra could also be from a large eruption of Katla in 1625 (Thorarinsson, 1975). The stratigraphic thickness and spatial distribution of this tephra layer indicate that the ash is most likely to be from Katla in 1625.

Kv x – 5-6 cm above the Katla 1625 ash lies another fine-grained ash of basaltic composition. Mean soil accumulation rates of 0.68 mm a^{-1} put the timing of deposition c. AD 1700-1720. Comparison with the documentary evidence places it between the eruptions of Katla in 1625 and Katla in 1755. The only known basaltic eruptions dating

from this period were of Kverkfjöll in 1711 and 1717. The extent of the ash-fall from these eruptions is presently unknown.

K 1755 – 3-4 cm above the previous tephra lies another fine-grained black ash, typically 1-2 cm in thickness. A soil accumulation rate of 0.68 mm a^{-1} places the time of deposition between AD 1760 and 1790. An eruption of the Lakagígar fissure system, between AD 1783 and 1786 deposited a thin and patchy ash layer in Öræfi (Gudmundsson (1998a: 79). It is possible that accumulation rates during much of the 17th, 18th and 19th centuries differed significantly from those during the historical period as a whole, primarily as a result of climatic deterioration. Hence, dating of tephra layers based on a constant soil accumulation rate may be inappropriate. If soil accumulation rates were suppressed during the cooler and less energetic climate of the 17th and 18th centuries, this tephra layer may date from the 1750s. The largest eruption of Katla in historical times took place in 1755 and deposited tephra across much of northern and eastern Iceland (Thorarinsson 1956, 1958). It would be extremely unlikely if this tephra was not present in the soil around Lambatungnajökull. The stratigraphic thickness of this tephra layer (~1-2 cm) indicates that it probably originated from the eruption of Katla in 1755.

A 1875 – In several soil sections in Skyndidalur, a fine-grained, white, discontinuous, ash layer 1-3 mm thick was identified at 0.16 m depth. Post-dating the silicic eruptions of Öræfajökull, Katla and Hekla, this ash probably originated from Askjá. A medium-sized eruption occurred in AD 1875. The main tephra plume was blown due east before reaching the coast of Norway some days later. Thorarinsson (1958) found Askjá-1875 ash near Berufjörður, 150 km SE of Askjá and only 50 km NE of Lón.

K 1918 – Increased soil accumulation over the 20th century as a result of climatic amelioration and increased storminess has almost certainly occurred in Iceland. Based on the presence of Askjá 1875 tephra at 0.16 m depth, mean soil accumulation rates are calculated to be 1.3 mm a^{-1} for the period 1875-1998. At a depth of 10-11 cm lies a

patchy, fine-grained basaltic tephra probably dating from AD 1913 – 1922. The most likely candidate for the origin of this tephra is the eruption of Katla in 1918.

G 1934? – The most recent tephra identified in the study area lies 6-8 cm below the present-day soil surface. It is a discontinuous, fine-grained, black tephra layer typically 3-7 mm thick. Dating from c. 1930 – 1950, based on a mean soil accumulation rate of 1.3 mm a^{-1} since 1875, this ash-fall probably originated from the Grimsvötn eruptions of 1934 or 1946.

The upper 6 cm or so of soil is apparently free of tephra in this part of Iceland. This reflects the relatively dormant state of Iceland's volcanoes over the last 50 years or so. High-magnitude eruptions, of the scale seen preserved in the historical soil sequence around Lambatungnajökull, have been scarce over the last 100 years.

Tephras deposited before AD 1362

The tephrastratigraphy of the pre-settlement period is much more difficult to unravel than that of the last 1000 years. Without documentary evidence of eruptions, individual tephra layers can only be placed in a chronological framework using absolute dating techniques and by association with the overall stratigraphy. Very few pre-settlement tephra layers in Iceland have been dated accurately. Problems surround the lack of dateable organic material in the soil available for radiocarbon analysis and the correlation of individual ash layers over large areas. However, Larsen (1984), Dugmore *et al.* (1995), Grönvold *et al.* (1995) and Gudmundsson (1998a) have all determined the age of various tephra layers deposited before the settlement of Iceland using ^{14}C. Most recently, Roberts (2001) has dated ancient volcanic ash in Northwest Iceland using luminescence and fission-track techniques. These marker horizons have all been used to interpret the timing of glacier fluctuations in Iceland.

Hekla-5 – The oldest tephra layer identified in any the soils in the vicinity of Lambatungnajökull is a fine-grained, grey-white, partly indurated horizon, typically 3 cm in thickness. The tephra, where observed on the upland Dalsheiði plateau, lies at a depth of 1.60 m, 60 cm below the level of the next dated 'marker' horizon and over 100 cm below the Öræfajökull 1362 ash. The geochemistry of this lowermost silicic ash suggests it is from an eruption of Hekla in 6350 ^{14}C yr BP (Dugmore *et al.*, 1995) (see Appendix I).

Hekla-4 – Two further pre-settlement tephra layers have been identified in the study area. Located at a depth of 105 cm on Dalsheiði, 100 cm in Fossdalur and normally between 125-140 cm in Hoffellsdalur is a fine-grained, pale yellow-white tephra. The ash layer is usually 2-3 cm thick and forms a distinctive marker horizon where it is found in the soils of Hornafjörður and Lón (Fig. 6.44). The tephra layer has been independently dated to c. 3850 ^{14}C yr BP elsewhere in Iceland (Dugmore *et al.*, 1995).

H-S (Selsund pumice) – Found 5-10 cm above the H-4 tephra is a third 'marker' layer dating from the pre-settlement period. Comprising a basaltic and silicic component, this fine-grained tephra layer forms a distinctive horizon, normally 1-2 cm thick. The relative quantity of silicic glass within the tephra layer varies from place to place. In lower Hoffellsdalur the silicic shards are concentrated in a band 3-5 mm thick, set within 15 mm of basaltic ash. In Fossdalur, the white shards are randomly scattered throughout the entire tephra layer making identification of the silicic component difficult. This bipartite layer is also believed to originate from Hekla and is referred to as *H-S*. The H-S tephra is synonymous with the *Selsund pumice* in Eyjafjallasveit (Thorarinsson and Larsen, 1977). Gudmundsson (1998a) dated the H-S tephra in Öræfi using soil accumulation rates to 3520 +/-30 BP. The couplet of H-S and H-4 tephras forms a key marker horizon in the valleys around Lambatungnajökull.

Figure 7.3 Pre-AD 1362 tephrastratigraphy.

(left) Tephra pit DH 03, on the Dalsheiði plateau, revealed the longest tephrastratigraphic record in the study area..

Ö-1362 tephra - 0.40 m

Hekla-4 c. 3800 BP – 1.00 m

Hekla-5 c. 6300 BP – 1.60 m

Thin, compacted peat layer containing birch fragments – 1.95 m

(left) Tephra pit HO 03, in lower Hoffellsdalur, provides a minimum age for moraine H1.

Ö 1362 – 0.70 m

Hekla-4 c. 3800 BP – 1.80 m
(not visible in photograph)

Figure 7.4 Post-AD 1362 tephrastratigraphy.

(left) Soil pit SK 06, adjacent to moraine S3. Sediment accumulation rates are high in this locality. 105 cm of soil has accumulated since AD 1362, equivalent to ~1.67 mm per year.

(left) Tephra pit FO 03, in lower Fossdalur. Compare the soil accumulation since AD 1362 in this locality with that above.

Note the pen-knife for scale (10 cm long).

Note the thin black tephra layer immediately beneath it.

Other pre-historic tephra layers, including Hekla-4 and H-S, were found at a depth of 0.90-1.00 m

Summary of tephrastratigraphy

The tephrastratigraphy of the valleys around Lambatungnajökull had not been studied prior to this research. Therefore these findings represent the first time that certain tephra layers have been identified in this part of southeast Iceland.

The white, silicic ash of the Öræfajökull eruption in AD 1362 has been identified across the study area. At least eight tephra layers have been deposited in the historical soil sequence (since AD 1362). The pre-settlement tephrastratigraphy is complicated and correlation between soil sections proved more difficult. For this reason no complete picture of the prehistoric (pre-AD 1362) tephrastratigraphy could be compiled (Fig. 7.2). Tephra layers from three pre-settlement eruptions of Hekla (H-S, H4, and H5) have been identified in several, long soil profiles. The earliest of these eruptions dates from c. 6350 BP suggesting that, in certain localities, soils have been accumulating since at least the early Holocene.

Age assessment of moraines based on tephrastratigraphy

Glacially-deposited landforms have been assigned minimum formation dates based on the presence of tephra layers in the surrounding soil. The findings of the tephrochronological investigations are summarised in Figures 7.5 - 7.9 and Tables 7.1 and 7.2.

Dalsheiði

The oldest moraines identified in the study area, ridges *S1a-d* located on the Dalsheiði plateau, were deposited more than ~9000 years BP. The presence of Hekla-5 ash overlying 50 cm of soil suggests that soil has been accumulating at this site since c. 9000 years BP (assuming a constant pre-historic soil accumulation rate of 0.19 mm a^{-1}) (Fig. 7.5). Although the assumption of a constant soil accumulation rate over much of the Holocene is somewhat unrealistic, the presence of a 20-30 mm-thick, fine-grained black tephra layer at a depth of 1.90 m may partly confirm the age of this soil sequence.

Based on its stratigraphic position, thickness and mineralogy, the tephra layer could be the *Saksunarvatn ash.* This tephra, dated to 9-9.1 ka ^{14}C BP (Mangerud *et al.*, 1986) has been found across much of southern and eastern Iceland, the Faeroe Islands, western Norway and northern Scotland. Future geochemical analysis will hopefully determine the provenance of this tephra layer with more certainty.

Hoffellsdalur

The oldest moraine identified in Hoffellsdalur, H1, formed at least c. 4800 years BP. The presence of H-4 and H-S ash in the soil immediately inside the moraine suggest that sediment has been accumulating since at least 3830 ^{14}C BP, and probably since 4800 BP based on soil accumulation rates. Soil lower down Hoffellsdalur, outside moraine H1, has been accumulating for at least c. 6000 - 7000 years based on sediment accumulation rates.

Moraines H2 and H3 are younger features. Moraine H2 formed prior to c. 2700 BP based on the depth of soil found in profile HO 01 (Fig. 6.48), and no later than c. 4800 BP. Moraine H3 was formed between c. 1200 BP and c. 2700 BP.

The youngest moraines, at the head of the valley, were all formed during the *Little Ice Age.* H4 and H5 probably did not form after c.1840 AD, according to the depth of soil between moraines H5 and H6; whilst moraines H6, H7 and H8 formed prior to c. AD 1908 but after moraine H5 (Fig. 7.6).

Skyndidalur

The absence of moraines in lower Skyndidalur does not allow a detailed picture of early- and mid-Holocene glacier fluctuations to be compiled. The deepest soils were encountered at the extreme eastern end of Skyndidalur (Fig. 7.7), here sediment has been accumulating since at least c. 3500 years BP. The height of the lateral moraine, S2, on the north side of the valley indicates that at the time of moraine formation Lambatungnajökull terminated within Skyndidalur. Hence, moraine S2 probably formed after c. 3500 years BP. Moraine S3 formed between c. 1200 and c. 1800 years BP based on the depth of soil in profiles SK 05 and SK 06 (Fig. 7.7). Moraine S4 had formed by AD 1875, but not before AD 1755, as the Katla 1755 tephra was absent from the base of

the soil profile (Fig. 7.8). The youngest moraines, S6-S9, formed after AD 1875. Moraine S9 was emplaced by AD 1934 (Fig. 7.8).

Fossdalur

Deep mature soils are indicative of several millennia of ice-free conditions in Fossdalur. The oldest soil sequences, over 1 m thick, are found in the lowest part of the valley (Fig. 7.9). The presence of H4 and H-S tephra prove that glacier ice has not occupied the main part of the valley floor for at least c. 4000 years (Fig. 7.9).

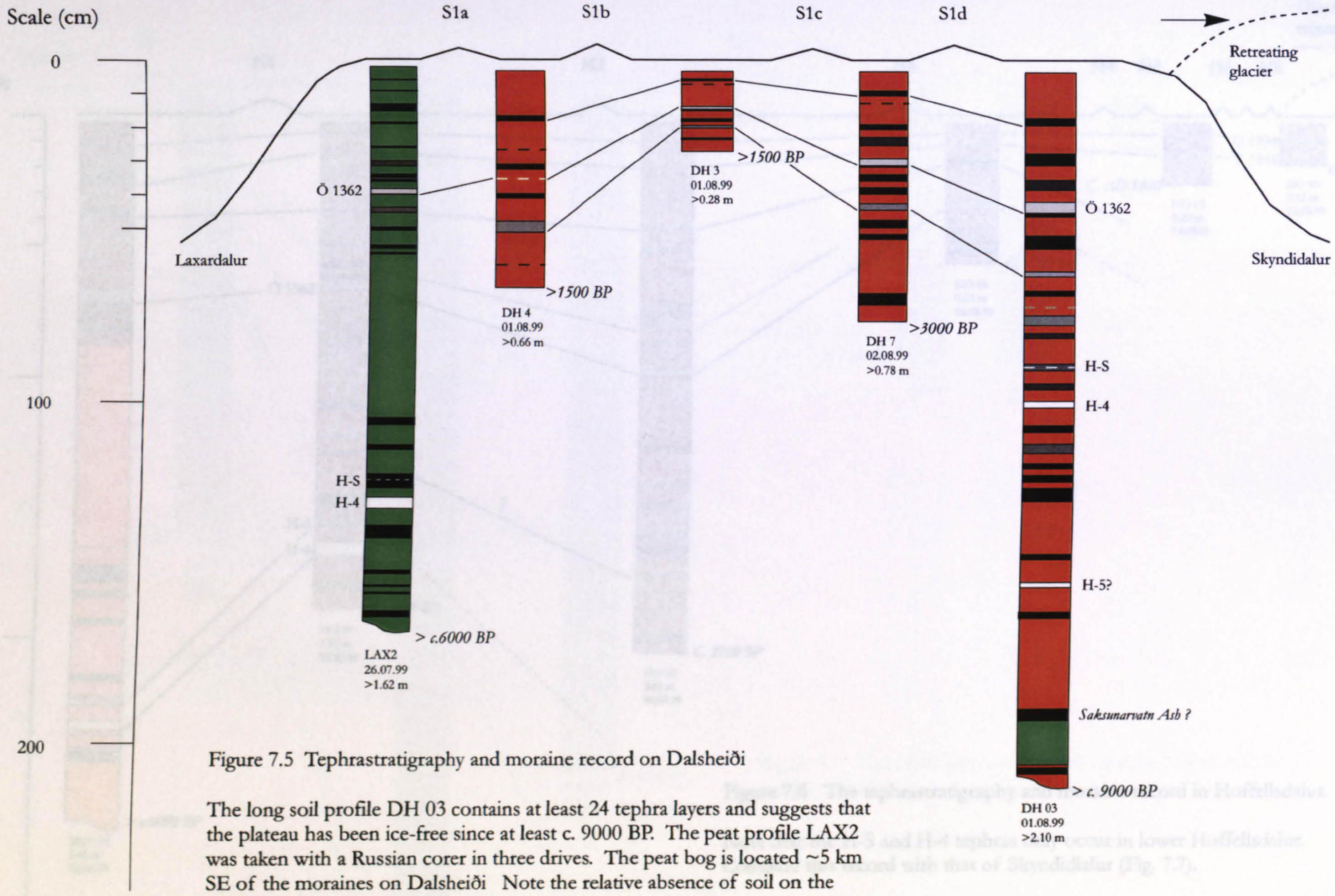


Figure 7.5 Tephrastratigraphy and moraine record on Dalsheiði

The long soil profile DH 03 contains at least 24 tephra layers and suggests that the plateau has been ice-free since at least c. 9000 BP. The peat profile LAX2 was taken with a Russian corer in three drives. The peat bog is located ~5 km SE of the moraines on Dalsheiði Note the relative absence of soil on the plateau between the moraines. Dates of moraine formation derived from soil accumulation rates are, as a result, anomalously young.

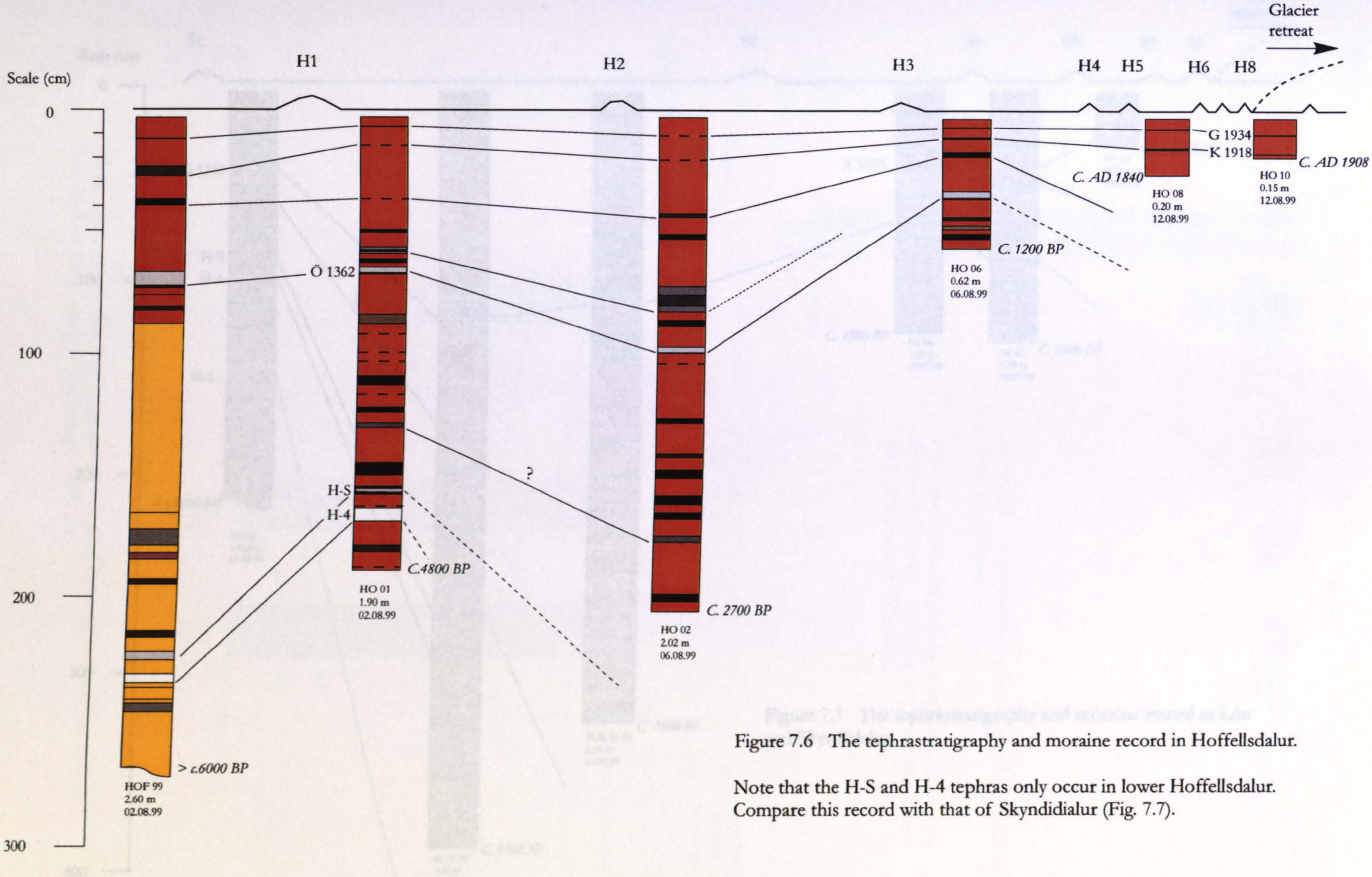


Figure 7.6 The tephrastratigraphy and moraine record in Hoffellsdalur.

Note that the H-S and H-4 tephras only occur in lower Hoffellsdalur. Compare this record with that of Skyndidialur (Fig. 7.7).

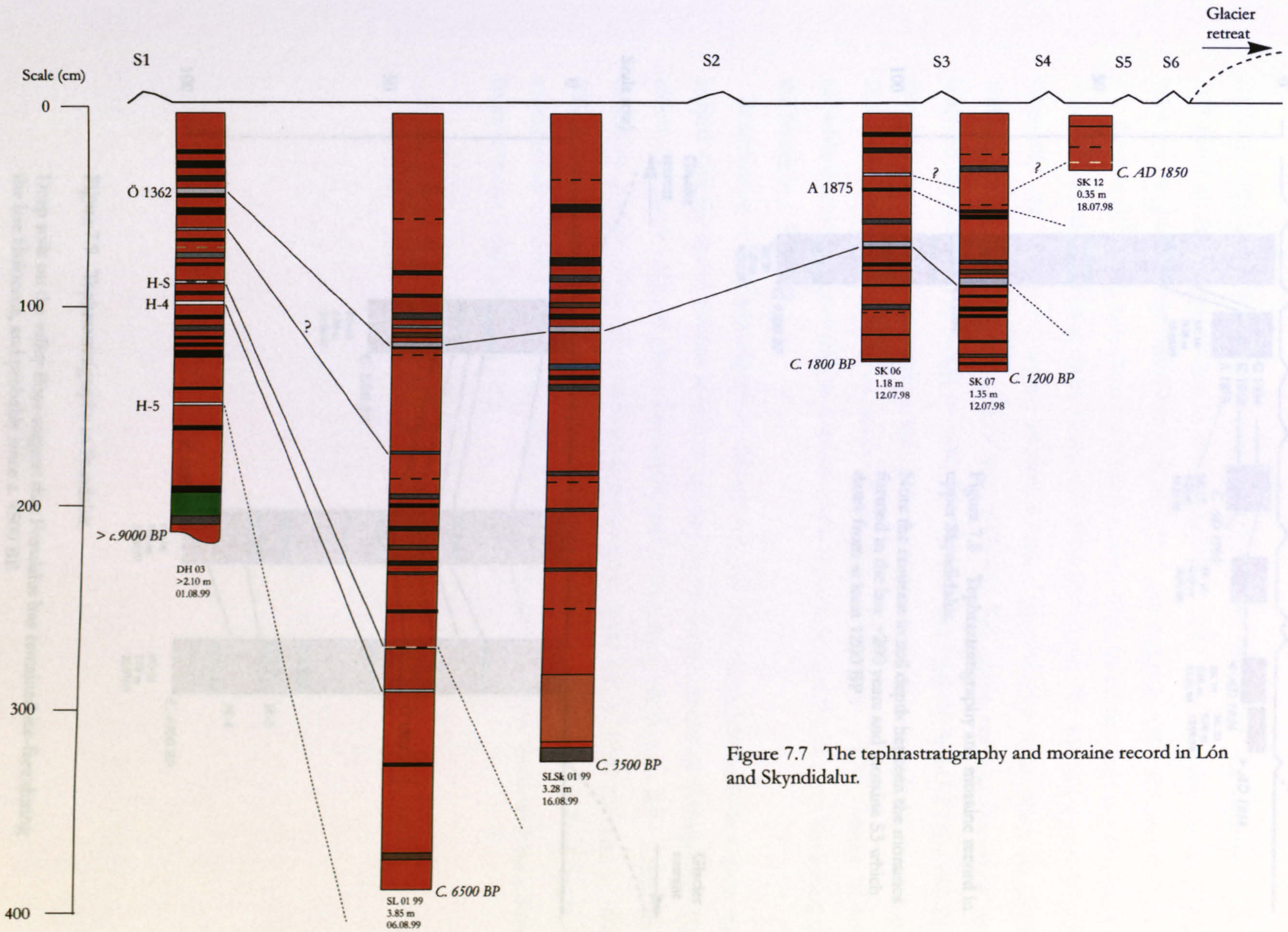


Figure 7.7 The tephrastratigraphy and moraine record in Lón and Skyndidalur.

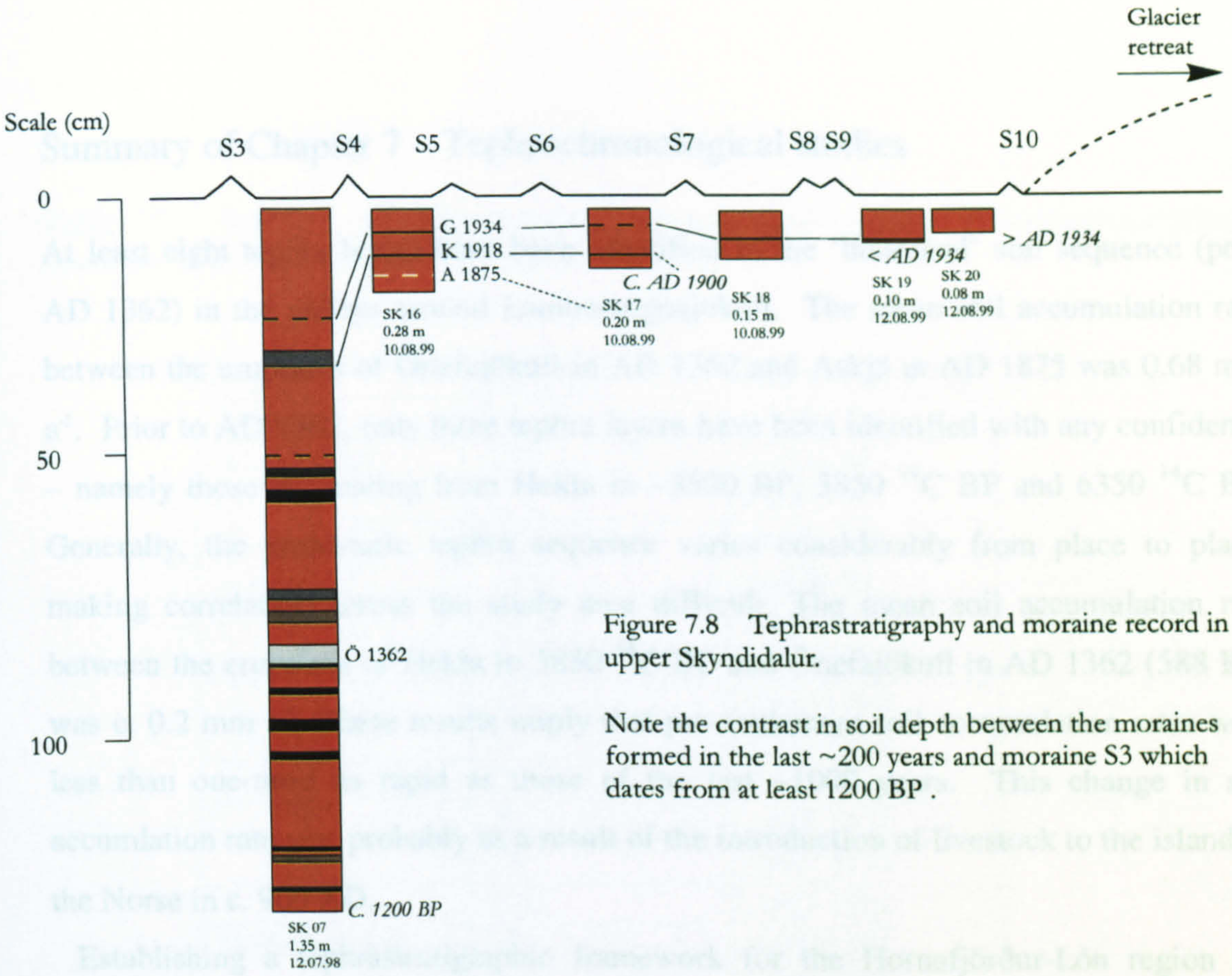


Figure 7.8 Tephrastratigraphy and moraine record in upper Skyndidalur.

Note the contrast in soil depth between the moraines formed in the last ~200 years and moraine S3 which dates from at least 1200 BP .

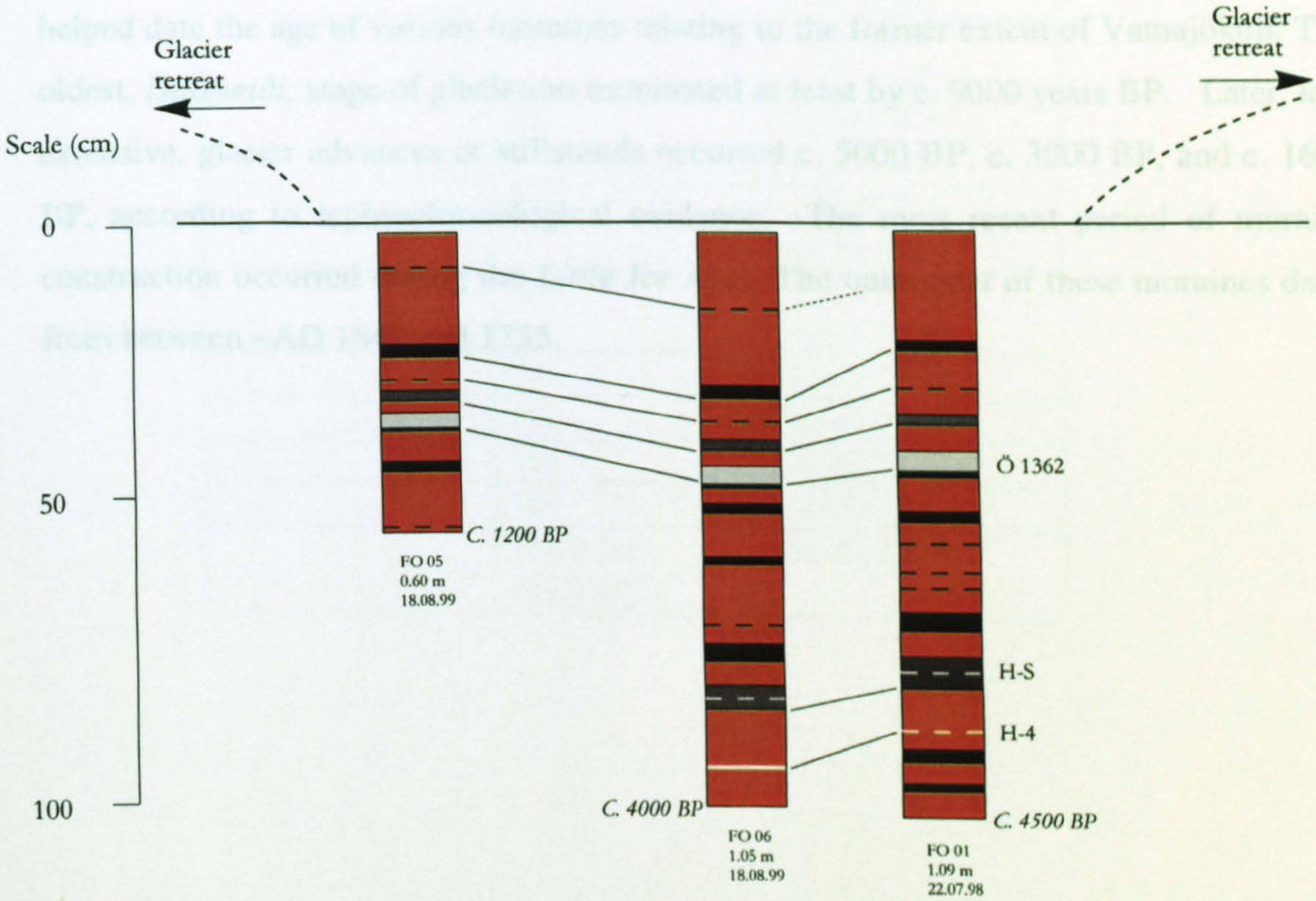


Figure 7.9 Tephrastratigraphy in Fossdalur.

Deep soils on the valley floor suggest that Fossdalur has remained ice-free during the late Holocene, and probably since c. 4500 BP.

Summary of Chapter 7 – Tephrochronological studies

At least eight tephra layers have been identified in the 'historical' soil sequence (post AD 1362) in the valleys around Lambatungnajökull. The mean soil accumulation rate between the eruptions of Öræfajökull in AD 1362 and Askjá in AD 1875 was 0.68 mm a^{-1}. Prior to AD 1362, only three tephra layers have been identified with any confidence – namely those originating from Hekla in ~3500 BP, 3850 ^{14}C BP and 6350 ^{14}C BP. Generally, the prehistoric tephra sequence varies considerably from place to place, making correlation across the study area difficult. The mean soil accumulation rate between the eruptions of Hekla in 3850 ^{14}C BP and Öræfajökull in AD 1362 (588 BP) was c. 0.2 mm a^{-1}. These results imply that pre-settlement soil accumulation rates were less than one-third as rapid as those of the last ~1000 years. This change in soil accumlation rate was probably as a result of the introduction of livestock to the island by the Norse in c. 900 AD.

Establishing a tephrastratigraphic framework for the Hornafjörður-Lón region has helped date the age of various moraines relating to the former extent of Vatnajökull. The oldest, *Dalsheiði,* stage of glaciation terminated at least by c. 9000 years BP. Later, less extensive, glacier advances or stillstands occurred c. 5000 BP, c. 3000 BP, and c. 1600 BP, according to tephrochronological evidence. The most recent period of moraine construction occurred during the *Little Ice Age*. The outermost of these moraines dates from between ~AD 1840 and 1755.

Table 7.1 Summary of moraine formation dates in the vicinity of Lambatungnajökull, based on tephrastratigraphic evidence.

Location	*Moraine*	*Bracketing dates*
Dalsheiði	S1	> 9000 BP
Skyndidalur	S2 ?	>1800 BP <3500 BP
	S3	>1200 BP <1800 BP
	S4	>AD 1755 <AD 1875
	S5	<AD 1900
	S6	<AD 1900
	S7	<AD 1920
	S8	<AD 1934
	S9	<AD 1934
	S10	>AD 1934
Hoffellsdalur	H1	>4800 BP <6000 BP
	H2	>2700 BP <4800 BP
	H3	>1200 BP <2700 BP
	H4	>AD 1755 <AD 1840
	H5	>AD 1755 <AD 1840
	H6	<AD 1908
	H7	<AD 1908
	H8	<AD 1908
	H9	>AD 1908
	H10, H11 H12, H13	>AD 1908

Table 7.2 Inferred correlation of Holocene glacier advances in Hoffellsdalur and Skyndidalur, based on tephrastratigraphic evidence.

Advance stage	*Moraines*		*Inferred date*
	Hoffellsdalur	Skyndidalur	
Dalsheiði	–	S1	> 9000 BP
Hoffellsdalur I	H1	–	c. 5000 BP
Hoffellsdalur II	H2	S2	c. 3000 BP
Hoffellsdalur III	H3	S3	c. 1600 BP
Hoffellsdalur IV	H4	S4	c. AD 1800
Hoffellsdalur V	H5	–	c. AD 1800
Hoffellsdalur VI - VIII, Skyndidalur VII - X	H6 - H13	S5 - S10	19th and 20th century AD

Chapter 8 : Results

Lichenometrical studies

Introduction

In order to reconstruct the deglacial chronology of Lambatungnajökull over the last two centuries or so, it was decided to use lichenometry. Numerous studies have shown that lichenometry can be used to provide relative and absolute surface ages in high latitude and high altitude environments. In particular, where dendrochronology and radiocarbon dating are not practical, and historical records are absent, lichenometry has proven invaluable. In this study lichens have been used to determine the surface exposure of the most recently deposited glacial landforms.

Methods

This study focuses on the crustose, yellow-green lichen *Rhizocarpon* Ram. Em. Th. Fr. Subgenus *Rhizocarpon.* Rather than merely measure a small number of 'largest' lichens, it was decided to gain an indication of the structure of the entire lichen population on a surface. The following method was adopted, which could be easily replicated at each site. Measurements were conducted within a fixed area, normally $30m^2$. Between 200 and 300 thalli of the lichen *Rhizocarpon* Section *Rhizocarpon*, were measured on each surface. An investigation found this sample size to be satisfactory for dating studies in Iceland (Appendix II). The species *Rhizocarpon geographicum* (L.) DC. predominated but other species within Section *Rhizocarpon* (eg. *R. macrosporum* Ras., *R. lecanorium* Anders, *R. saanense* Ras., *R.tinei* (Tornab.) Runem. (s.str.)) were probably also included. The longest axis of each thallus over 5 mm in diameter was recorded. Lichen measurements were made to the nearest millimetre using a transparent ruler. Elongate and irregular thalli were measured regardless of shape. Fragmented thalli, lacking a central portion, were measured when the outline could be clearly

discerned. Coalescent or multiple thalli were disregarded. For each surface a note was made regarding the inclination (horizontal, vertical or inclined) and orientation (N, S, E or W). Measurements were restricted to non-vesicular, basaltic rocks.

Establishing a lichenometric dating curve

Lichenometry is a calibrated-age dating technique. Before lichen sizes can be converted into surface ages a *dating curve* must be constructed. This stage is crucial, particularly if studies are conducted in geographical areas where no lichenometric studies have been performed previously. The construction of a lichenometric dating curve relies on the presence of independently-dated, lichen-bearing surfaces within the study region. The nature of the dating curve depends ultimately on the number and antiquity of these dated surfaces. Where dated surfaces are absent, lichenometry can be used only as a relative dating tool.

This study uses nine reliably-dated surfaces within southeast Iceland to construct a regional lichenometric dating curve (Fig. 8.1) Since climate is a crucial factor in determining the growth rate of lichens (eg. Beschel, 1950, 1957, 1961; Miller, 1973; Hale, 1974; Porter, 1981; Innes, 1985), ideally the 'control' surfaces should all be confined to a relatively small area and hence be subjected to the same environmental conditions. In this study, the nine dated surfaces are in areas representative of the climate of southeast Iceland as a whole. They are all less than 200 m above sea level and within 20 km of the Icelandic coast. All nine control sites experience broadly the same climatic conditions (Fig. 8.1), although precipitation between the sites ranges from 1500 to 2500 mm/yr and could be a potential source of lichen growth rate variation.

Table 8.1 lists the nine, dated 'control' surfaces and outlines the nature of evidence (ie. photographic, cartographic, documentary, etc.) used to derive the lichen dating curve. Unfortunately, lichenometry is problematic on surfaces less than about thirty years old due to uncertainties surrounding lichen colonisation time and the form of the initial growth phase (Beschel, 1961; Armstrong, 1974, 1983; Caseldine, 1983; 1991). The

youngest surface used for the construction of the dating curve therefore dates from AD 1967. The oldest surface used dates from AD 1727.

Graphs were generated for all lichen populations, with the logarithm of the frequency (%) being the dependent variable and lichen thallus diameter (mm) being the independent variable (Fig. 8.2). A class size interval of 3 mm was chosen for lichen diameter, as advocated by Innes (1986) and Caseldine and Baker (1998). Lichens falling below the modal class in each population have been omitted. Regression lines were then calculated using the least-squares method enabling each size-frequency distribution to be described in the form $y=mx+c$. This method also enables prediction of the '1 in 1000' thallus, namely the x-intercept (mm) at a $\log_{10}$ frequency of –1. Both these indices, regression gradient and '1 in 1000' thallus, have been used in lichenometry to describe population distributions successfully (Benedict, 1967, 1985; Locke *et al.*, 1979; Caseldine, 1991). A particular advantage of this technique is that problems regarding the incorporation of 'older', reworked debris are eradicated. Anomalously large lichen thalli that pre-date the time of surface exposure can be easily detected and disregarded.

The largest lichen in each of the nine, undisturbed, populations was plotted (y-axis) on a graph against the known age of the surface (x-axis) on which it was growing. These points connected by a smooth line yield a lichenometric dating (age-size) curve (Fig. 8.3).

Several workers have employed lichenometry in order to estimate the age of various glaciogenic features in Iceland. Most have independently constructed lichenometric dating curves (Gordon and Sharp, 1983; Caseldine, 1983; Maizels and Dugmore, 1985; Thompson and Jones, 1986; Kugelmann, 1991; Gudmundsson, 1998b; Evans *et al.*, 1999). With the exception of Thompson and Jones's study, all the published dating 'curves' have yielded a **linear** relationship between lichen age and lichen size. Furthermore, most of the previous dating curves have been produced using only the largest lichen, or lichens, growing on an entire dated surface. For this study it was important to construct a regional dating curve using a fixed sample size, as practised in this research. The resultant curve could then be applied to lichen measurements on any particular surface.

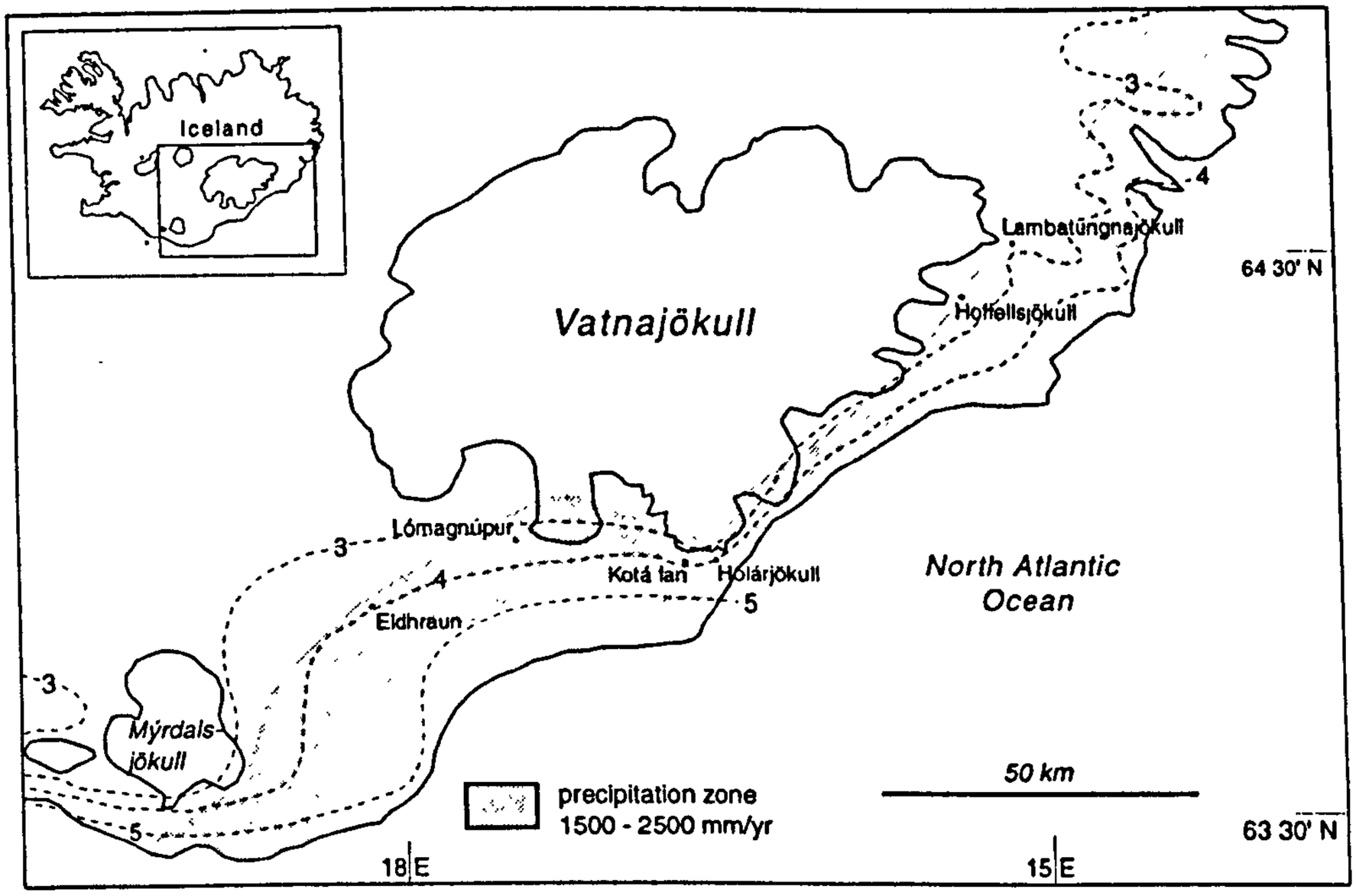
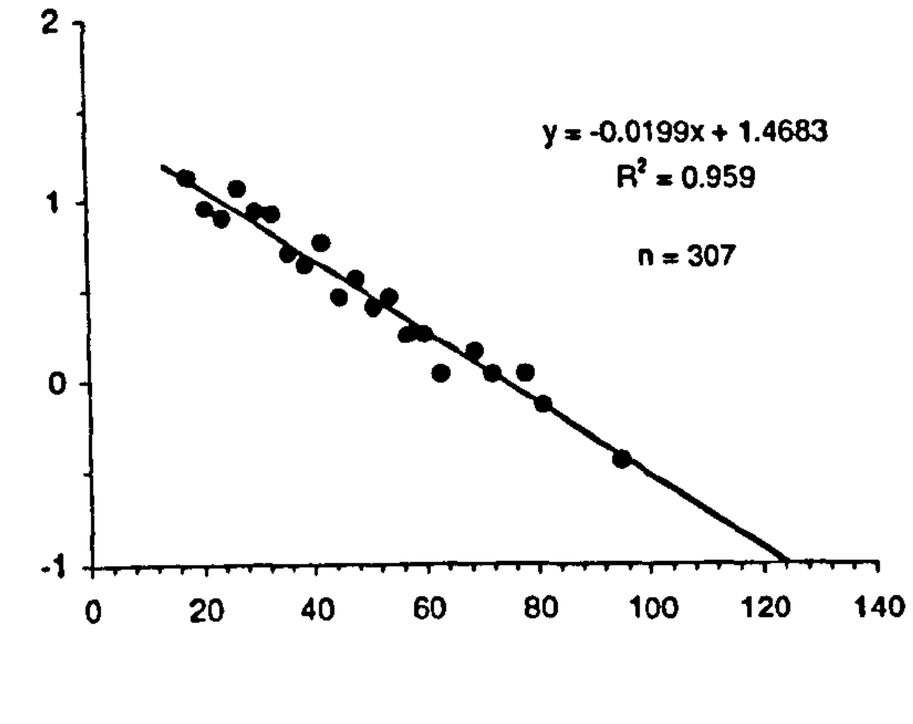


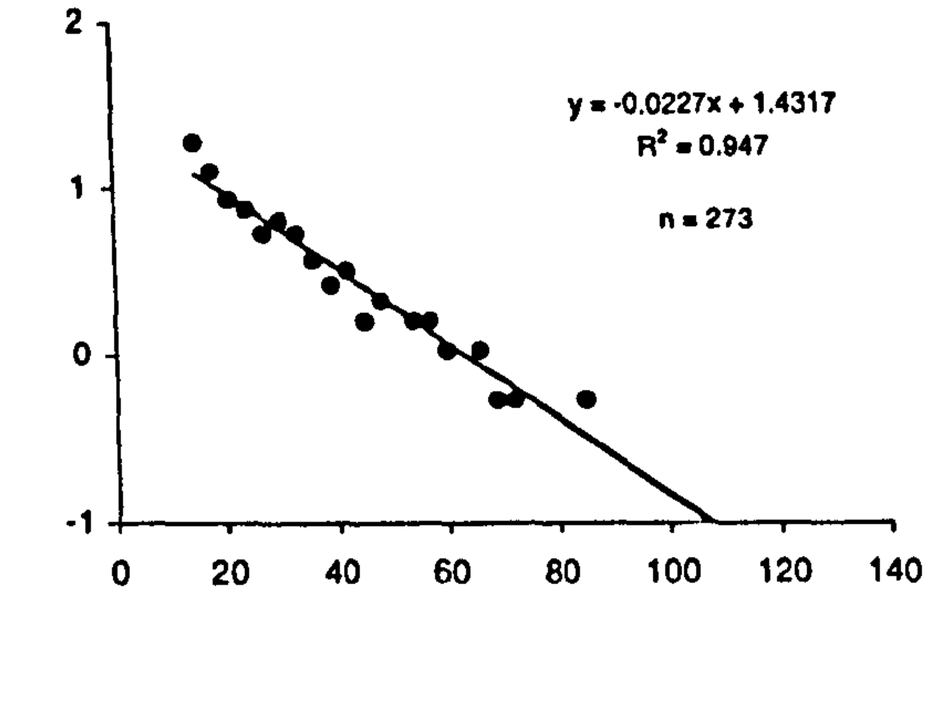
Figure 8.1 Location of dated surfaces in southeast Iceland used to establish the lichenometric dating curve.

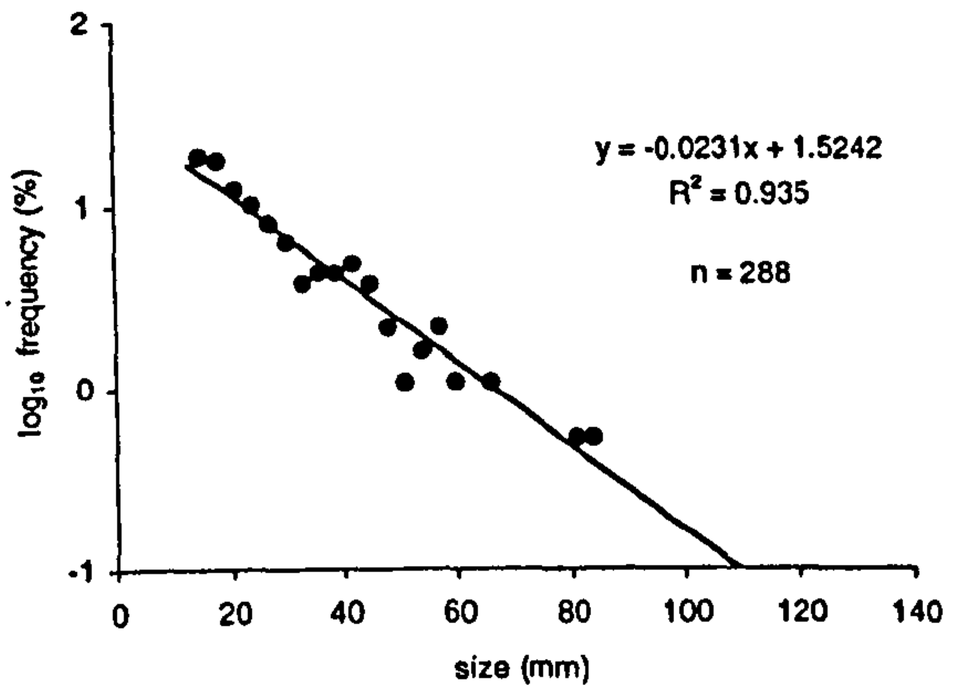
Dashed lines are isotherms of mean annual temperature (°C). Climatic data relate to the period 1961-1990 and are supplied by *Veðurstofa Islands.*

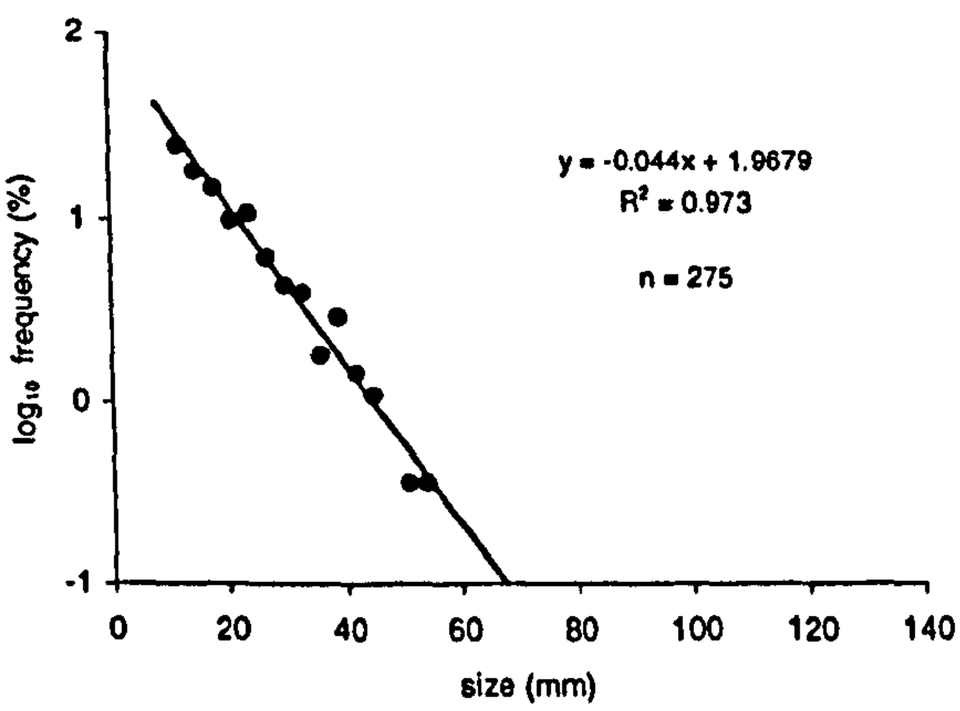
Table 8.1 *Lichenometric data from nine dated 'control' surfaces in southeast Iceland used to construct the regional dating curve in Figure 8.3. The number of the relevant aerial photograph is quoted in parentheses after the respective date.*

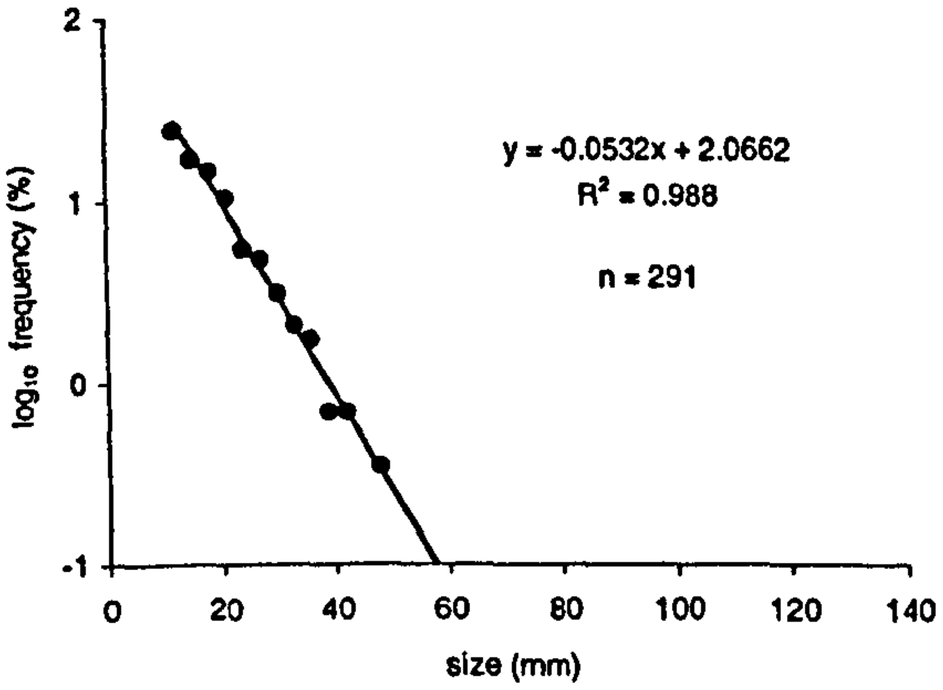
No.	Surface	Location (Lat., Long.)	Geomorphological setting	Date of surface (AD)	Source	Largest-lichen diameter (mm)
1	bedrock *roche moutonnée*	Lambatungnajokull 64°30' N, 15°18' W	glacially-striated, streamlined, basaltic, bedrock hummock	1967	aerial photograph (4577) 1.9.67	10
2	bedrock step	Lambatungnajokull 64°30' N, 15°18' W	glacially-eroded, basaltic, bedrock step	1957	aerial photograph (5168) 22.5.57	16
3	moraine ridge	Lambatungnajökull 64°30' N, 15°17' W	ice-proximal flank of low-elevation, moraine ridge	1945	aerial photograph (51/1210) 30.8.45	23
4	moraine ridge	Hoffellsjokull 64°25' N, 15°24' W	ice-proximal flank of moraine ridge	1930	surveyed by H.H. Eiríkson (1932: p22)	32
5	ground moraine	Hoffellsjokull 64°25' N, 15°26' W	Unconsolidated, morainic material deposited on striated bedrock	1903	surveyed by Danish General Staff map, (1903, Sheet no. 96, SE)	46
6	moraine ridge	Hólárjökull 63°56' N, 16°32' W	ice-proximal flank of moraine ridge	1890 or 1891	photographed by F.W.W. Howell (1893: p70)	52
7	rockfall	Lómagnúpur 63°56' N, 18°30' W	large, angular, rockfall-derived, basaltic boulders	1789	described by Henderson (1819: p218)	84
8	lava flow	Laki Eldhraun 63°47' N, 18°08' W	slightly elevated, blocky, basaltic lava (free from moss covering)	1783	described by numerous authors (eg. Ahlmann, 1938)	85
9	outwash *jokulhlaup* fan	Kotá, in Öræfi 63°55' N, 16°44' W	angular, basaltic, waterlain, cobbles and boulders (catastrophic flood deposit or saturated debris-flow)	1727	described by numerous authors (eg. Thorarinsson, 1956)	95











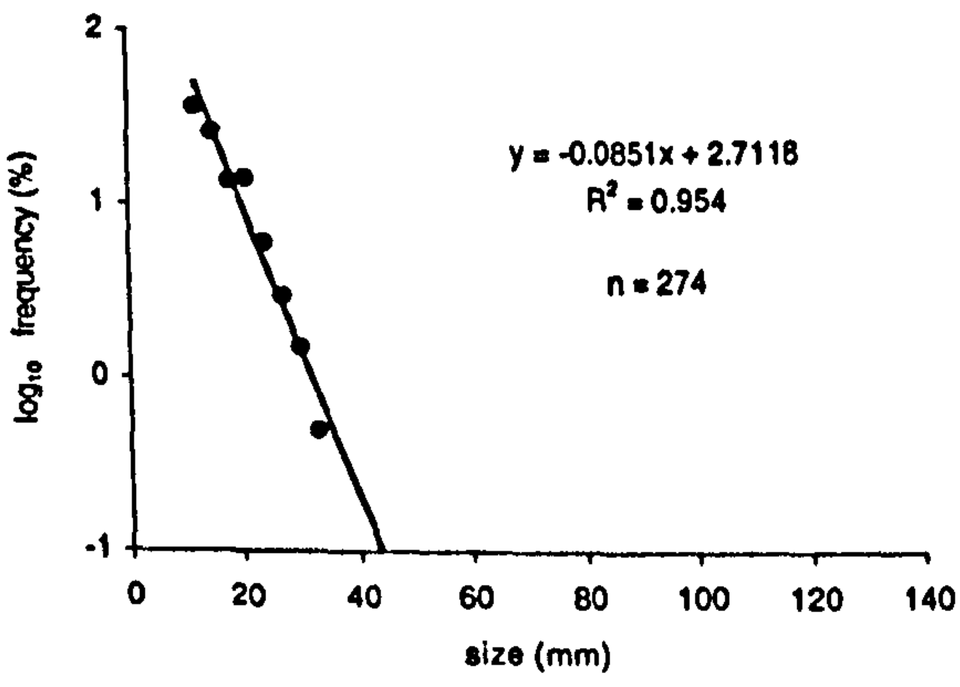
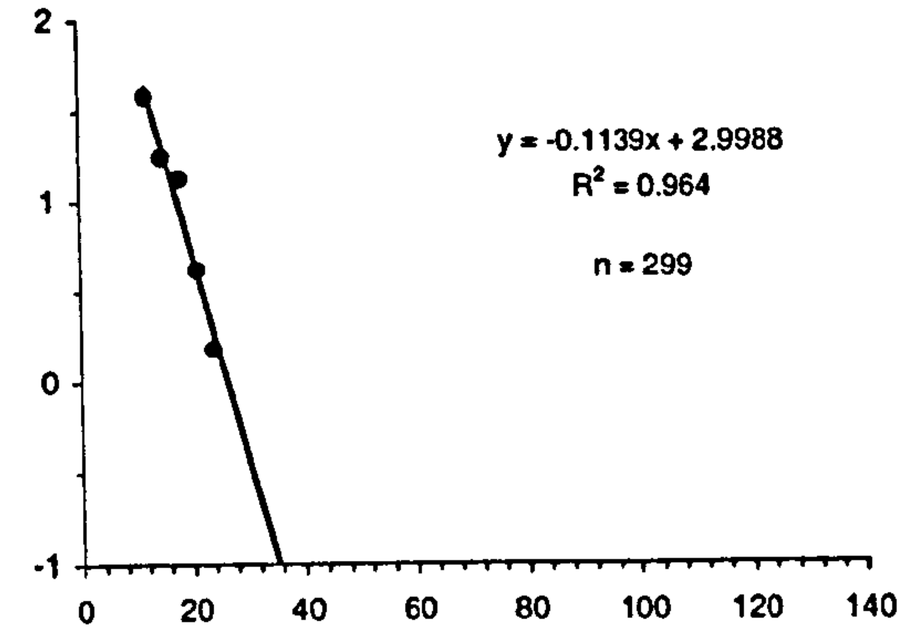
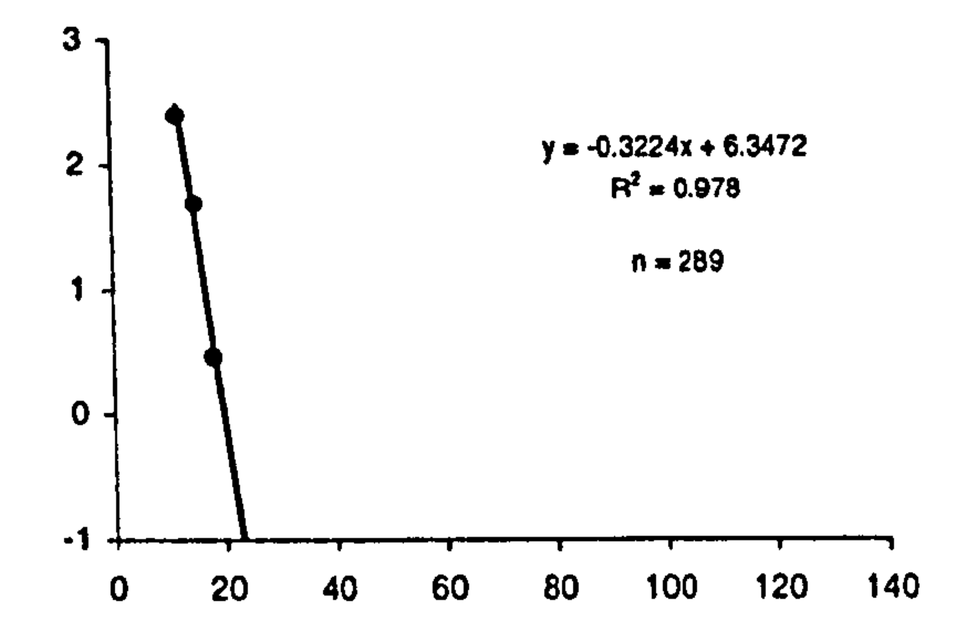


Figure 8.2 Lichen size-frequency graphs from nine dated surfaces in southeast Iceland





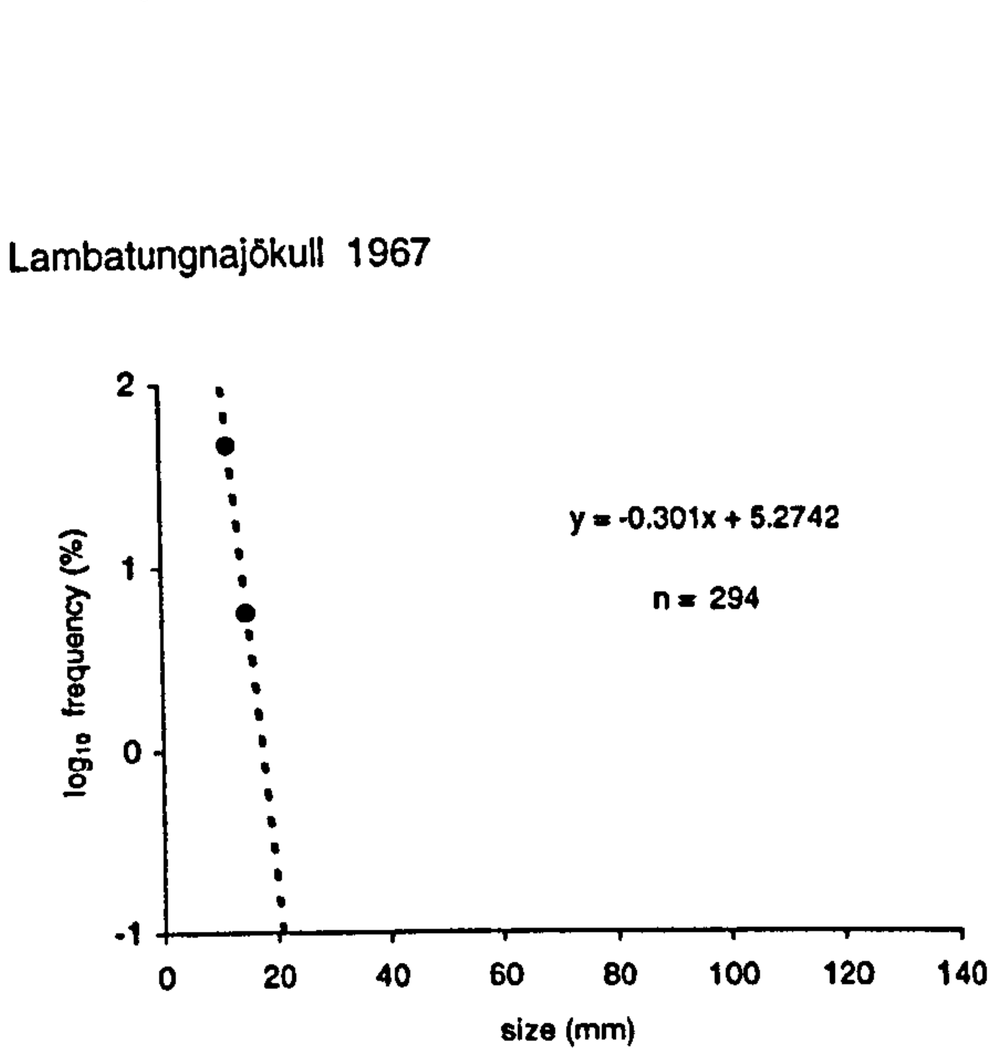
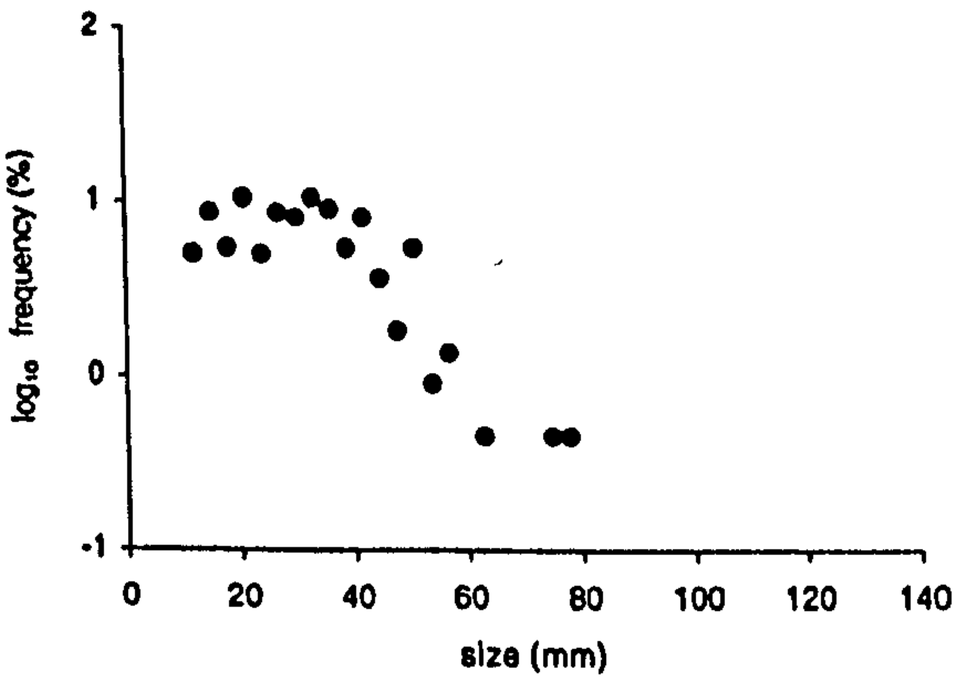


Figure 8.2 Lichen size-frequency (log_{10}) distributions on nine dated surfaces in southeast Iceland. These nine surfaces have been used to calibrate a lichenometric dating curve.

(right) Note that lichens growing on boulders deposited by a jökulhlaup in AD 1362 do not conform to the log-linear distribution model.



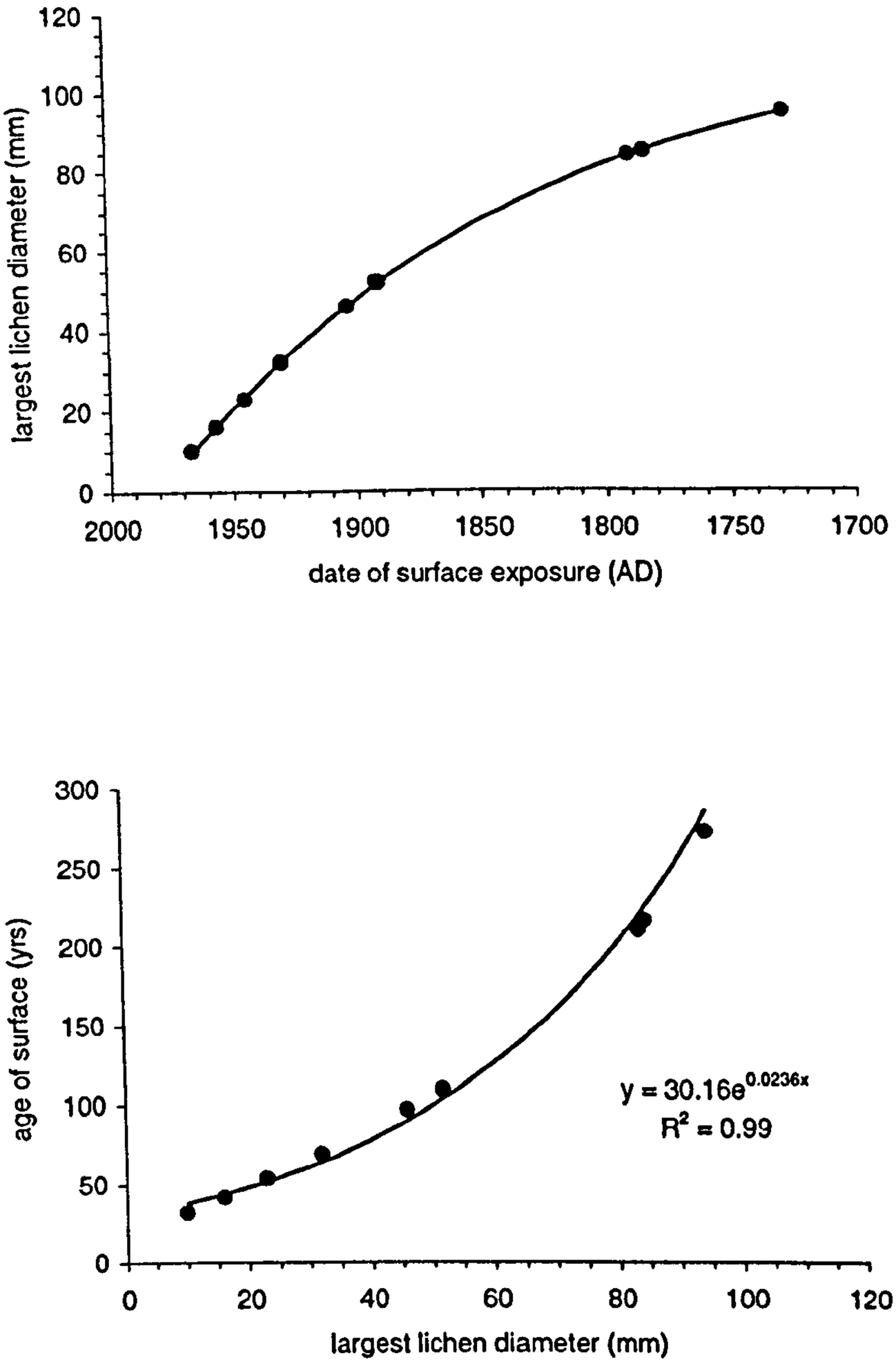


Figure 8.3 The relationship between surface age and the size of the largest lichen growing within a fixed area.

(uppermost) Calibration curve, with *age* as the independent variable. The best-fit curve is a third order polynomial.

(lower panel) Dating curve, with *lichen size* as the independent variable. The age-size relationship over the whole ~300-year period is non-linear and is best described by an exponential curve.

Figure 8.3 shows that the relationship between the size and age of the lichen *Rhizocarpon* Section *Rhizocarpon* in southeast Iceland over the last three centuries is **non-linear**. When largest-lichen diameters are plotted against surface age the relationship is optimally described by a third-order polynomial function (r^2=0.999). However, the first six points on the curve (lichens up to 52 mm in diameter) can be equally well described by a straight line ($y = 0.545x + 1.60$; $r^2 = 0.99$), suggesting that a linear growth model may be applicable over several decades. The oldest three points do not fit the linear growth rate model and therefore require a curvilinear function be employed.

When plotted with lichen size as the independent variable (x), as done by Benedict (1967, 1985) and Mottershead and White (1972), the relationship is best described by a curve of the form $y = 30.16e^{0.0236x}$ ($r^2 = 0.99$) (Fig. 8.3). This relationship suggests that the rate of *Rhizocarpon* lichen growth decreases exponentially with time. A logarithmic transformation yields a linear function.

Since the sampling method adopted in this study is based on lichen populations, it is believed to be more rigorous than strategies based on one or the mean of several largest lichen diameters (Benedict, 1967, 1985; Caseldine, 1991; Bull *et al.*, 1994; 1995). Given the stringent field procedures and the reliability of the control points involved in this study, this new lichenometric dating curve is probably the most reliable derived for southeast Iceland.

A new approach – the 'size-frequency' dating curve

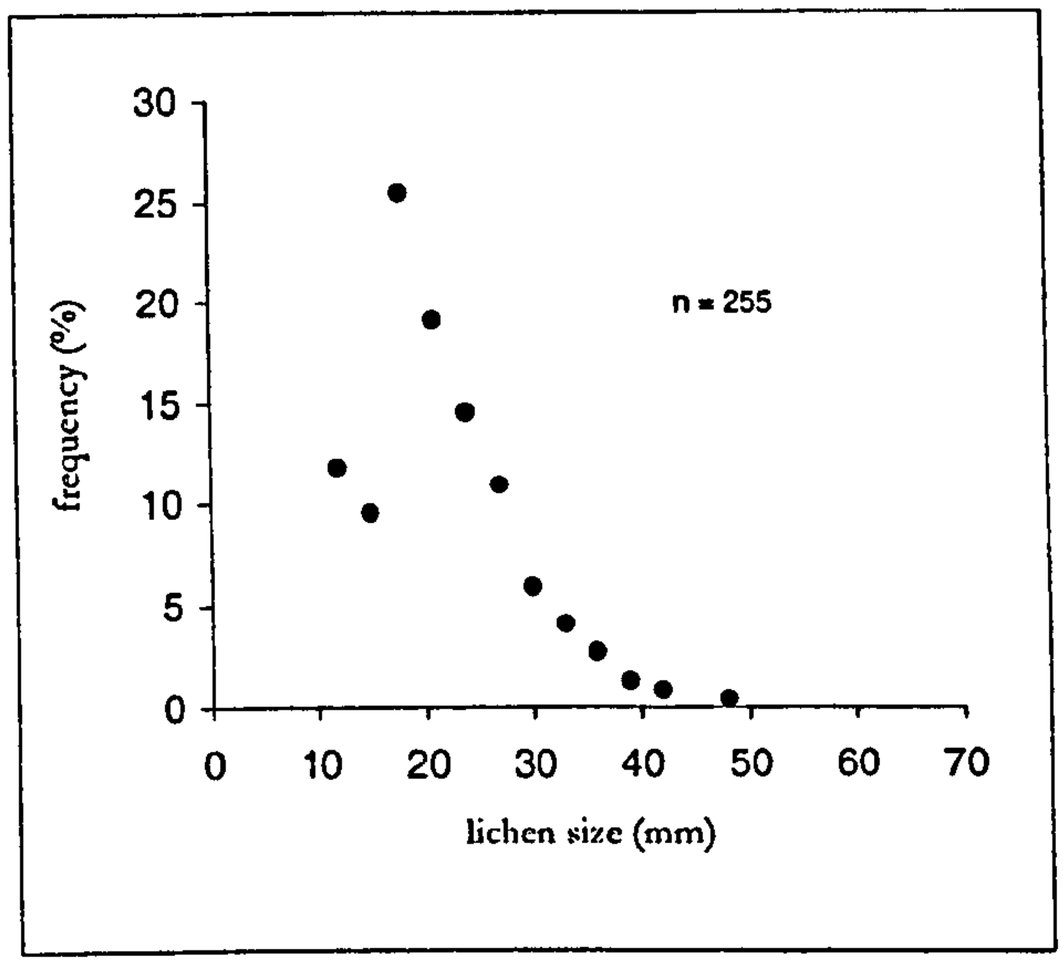
Benedict (1967, 1985), working in the Front Range mountains of Colorado, used lichenometry to determine the age of various glacial landforms and archaeological features. The adoption of the *population* or *size-frequency* approach allowed Benedict to distinguish successfully between single and disturbed, composite '*Rhizocarpon geographicum*' populations. Benedict (1967, 1985) used the best-fit slope of the size-frequency (log_{10}) distribution rather than the hypothetical '1 in 1000' thallus, as advocated by Locke *et al.* (1979). The '1 in 1000' thallus diameter is obtained by

extrapolation and is strongly influenced by the class intervals involved in the generation of size-frequency distribution graphs. The use of the gradient of the size-frequency distribution curve is preferred as it is an intrinsic characteristic of the lichen population, rather than merely an arbitrary measure (Benedict, 1985; Innes, 1986; Caseldine, 1990) (Fig. 8.4).

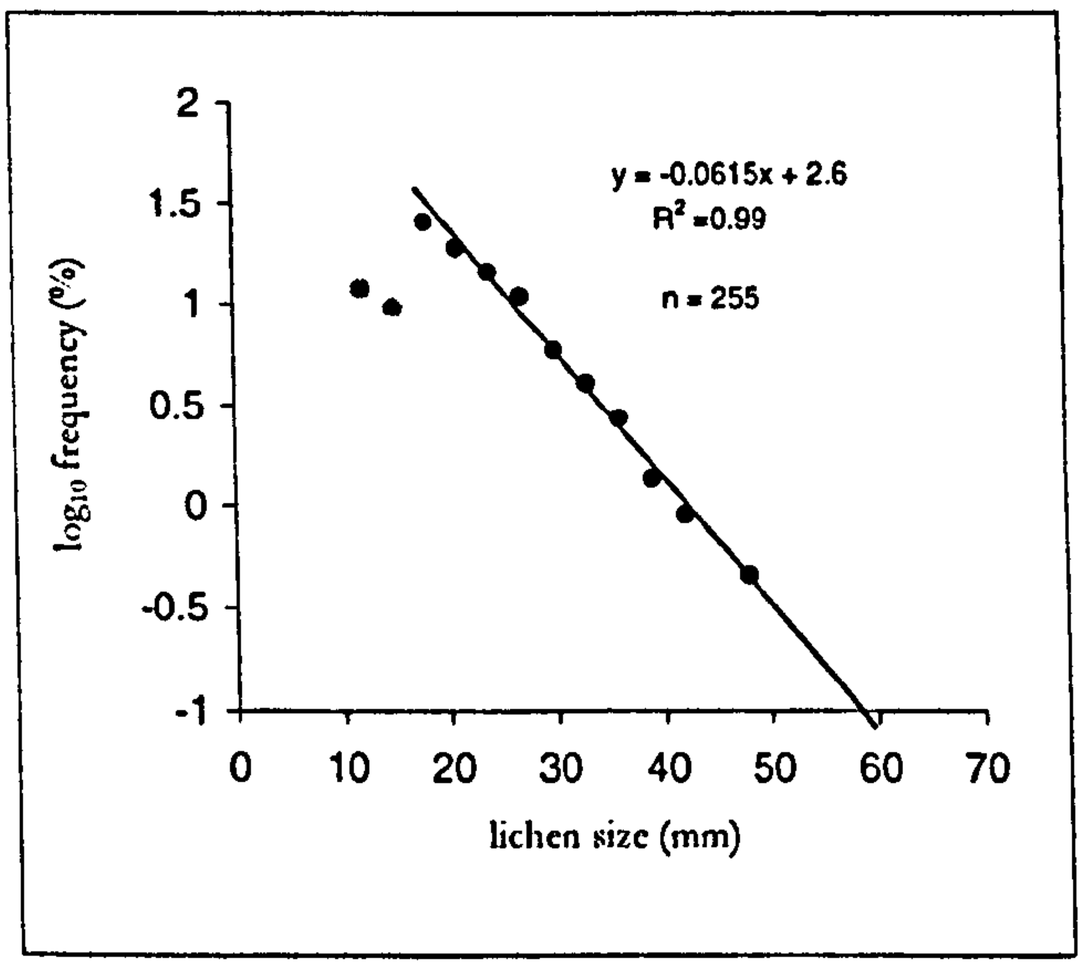
Numerous lichen size-frequency studies have been performed (eg. Benedict, 1967, 1985, 1993, 1999; Andersen and Sollid, 1971; Lindsay, 1973; Innes, 1983b, 1986; Cook-Talbot, 1991; Caseldine, 1990, 1991; Caseldine and Baker, 1998). The nature of the size-frequency distribution in an undisturbed lichen population has been the focus of some debate. Benedict (1985) proposed that the size-frequency distribution of lichens in an undisturbed community is negative log-linear and that the slope of the regression line becomes more gentle with time. Innes (1983b) argued that the form of the size-frequency distribution would range from truncated log-normal to Poisson in form, depending primarily on the age of the population being sampled. Innes (1983b: 285) pointed out that the use of the '1 in 1000' technique assumes a true log-normal distribution and may be "inappropriate as a population descriptor". However, he does not preclude the use of other indices (ie. gradient, skewness, kurtosis, etc.) as population descriptors. Caseldine and Baker (1998) stress the uncertainty surrounding the 'natural' form of lichen size-frequency distributions, and treat each population separately. The use of a best-fit regression line, as proposed by Benedict (1985), avoids the arguments surrounding the exact form of the distribution whilst describing the lichen population most appropriately and concisely.

Benedict's study of stone archaeological remains in the mountains of Colorado allowed him to construct a dating curve based on the slopes of the lichen size-frequency (log_{10}) regression lines (1985: Fig.87) (Fig. 8.5). He used six dated surfaces to construct his age-gradient curve. The range of Benedict's curve is c. 4000 radiocarbon years. The dating precision naturally decreases with time. Notwithstanding, Benedict showed how the gradients of undisturbed lichen populations could be applied to this curve to provide reliable age estimates of various undated lichen-bearing substrates.

Figure 8.4 Lichen size-frequency distributions



(a) Size-frequency distribution of lichens growing on a moraine in SE Iceland. Note the class-size interval is 3 mm.



(b) Size-frequency (log_{10}) distribution of lichen population in (a). Least-squares regression is used to calculate the gradient of the distribution (m = -0.0615). Note those lichens to the left of the modal class have been omitted from the regression calculation. The high r-sqared value suggests that this is a single, undisturbed population.

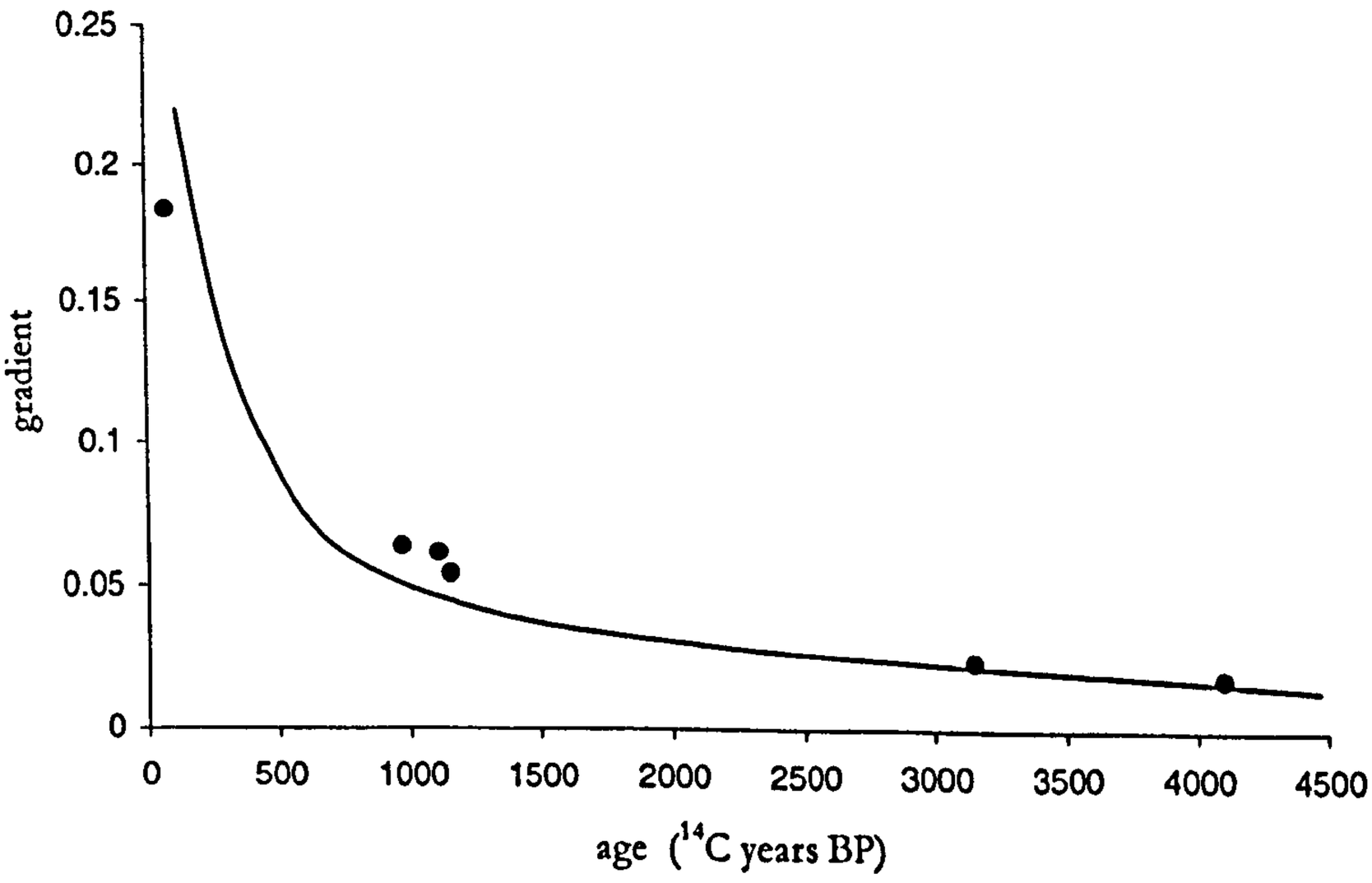


Figure 8.5 (above) The relationship between the age of a lichen population and the gradient of its size-frequency (log_{10}) distribution – Benedict's *rate curve* (redrawn from Fig. 87. 1985). Gradients have been multiplied by -1 for convenience. The curve is based on *Rhizocarpon geographicum* lichens growing on dated surfaces in the Front Range, Colorado.

There are various disadvantages of the size-frequency technique, which Benedict highlights. Firstly, the technique is dependent on a large number of lichens being present (Benedict recommends a sample size of 1000 thalli), and consequently is restricted to surfaces over a certain size. Secondly, dating precision will always be lower on older surfaces because of the exponential shape of the 'gradient' dating-curve. Finally, there are lithological and ecological uncertainties surrounding lichenometry, such as the weathering history of rock surfaces and the lichen colonisation rate, which may affect the slope of the size-frequency distribution. Benedict adds that the size-frequency dating method is not likely to prove useful in temperate or maritime environments where the growth of crustose lichens is relatively rapid. Subsequent studies have shown that this is not necessarily the case (Innes, 1986; Cook-Talbot, 1991; Caseldine, 1991; Caseldine and Baker, 1998).

There are several advantages of the 'size-frequency' technique over other lichenometric methods. Firstly, age estimates are based on a large population of lichens and, therefore, are not dependent on the identification of the single largest-lichen growing on an entire surface. Sampling several hundred lichens, as opposed to one or five, will ultimately produce results of greater statistical significance. Secondly, it allows the easy identification of disturbed or multiple populations, difficult to detect using standard lichenometrical procedures, that could yield erroneous surface ages. Thirdly, studies have shown that size-frequency results are largely reproducible, particularly if a standardised method is adopted (Locke *et al.*, 1979; Benedict, 1985, 1993; Caseldine, 1990, 1991). Finally, Benedict (1985) points out that, as so many steps are involved in size-frequency data analysis, it is very unlikely that any bias could be incorporated into the results. In short, the technique can provide a non-destructive, reliable age-estimate for substrates deposited or exposed during the last few centuries in temperate, polar and alpine environments, and even longer in more arid continental climates.

It was decided to construct an age-gradient curve using the lichen size-frequency data shown in Figure 8.2, collected from nine dated surfaces in southeast Iceland. The slopes of the regression lines of the size-frequency graphs have been plotted against the known

age of the various surfaces. The resulting age-gradient plot represents a calibrated dating curve for lichen-bearing surfaces in southeast Iceland (Fig. 8.6).

The relationship between the age of a surface and the gradient of the lichen size-frequency (log_{10}) distribution growing on the surface is an interesting one. Young surfaces are characterised by steep gradients whilst shallow gradients indicate older lichen populations, as observed by other workers (Benedict, 1967, 1985; Innes, 1983b; Caseldine, 1990, 1991). The rate at which the gradient decreases with time is not linear and approximates to a negative power function. When the slope of the size-frequency distribution (multiplied by –1, for convenience) is plotted as the independent variable (x-axis), with surface age on the y-axis, the resulting relationship is best described by a curve of the form $y = 7.307x^{-0.90}$ ($r^2 = 0.989$). A lower threshold is placed at around m = 0.13 – this equates to a lichen population with a maximum age of c. 45 years. This is comparable to the youngest surface on which a meaningful age estimate can be made using traditional 'largest-lichen' techniques in southeast Iceland (c. 35 years). On older surfaces, the minimum gradient likely to be useful for dating purposes is less well constrained, as the precision of the dating curve decreases with increasing age. Distributions where m = 0.01 are probably at the upper limit of the dating curve (ie. c. 400 years old). In the oldest lichen populations the assumption of a log-normal size-frequency distribution may not be appropriate (Innes, 1983b, 1986).

Testing the lichenometric dating curves

The form of both the 'age-size' and 'age-gradient' lichen dating curves and the strength of the statistical correlations ($r^2 = 0.999$ and 0.989) are very promising. It was decided to test the validity of the 'age-gradient' relationship using lichen populations from several other surfaces in southeast Iceland. Age predictions have been made for ten moraines of known age and two of unknown age at Fjallsjökull and Hólárjökull, based on the **gradient** of the lichen size-frequency distribution. By way of comparison, age predictions based on the **largest lichen** have also been made (Tables 8.2 and 8.3).

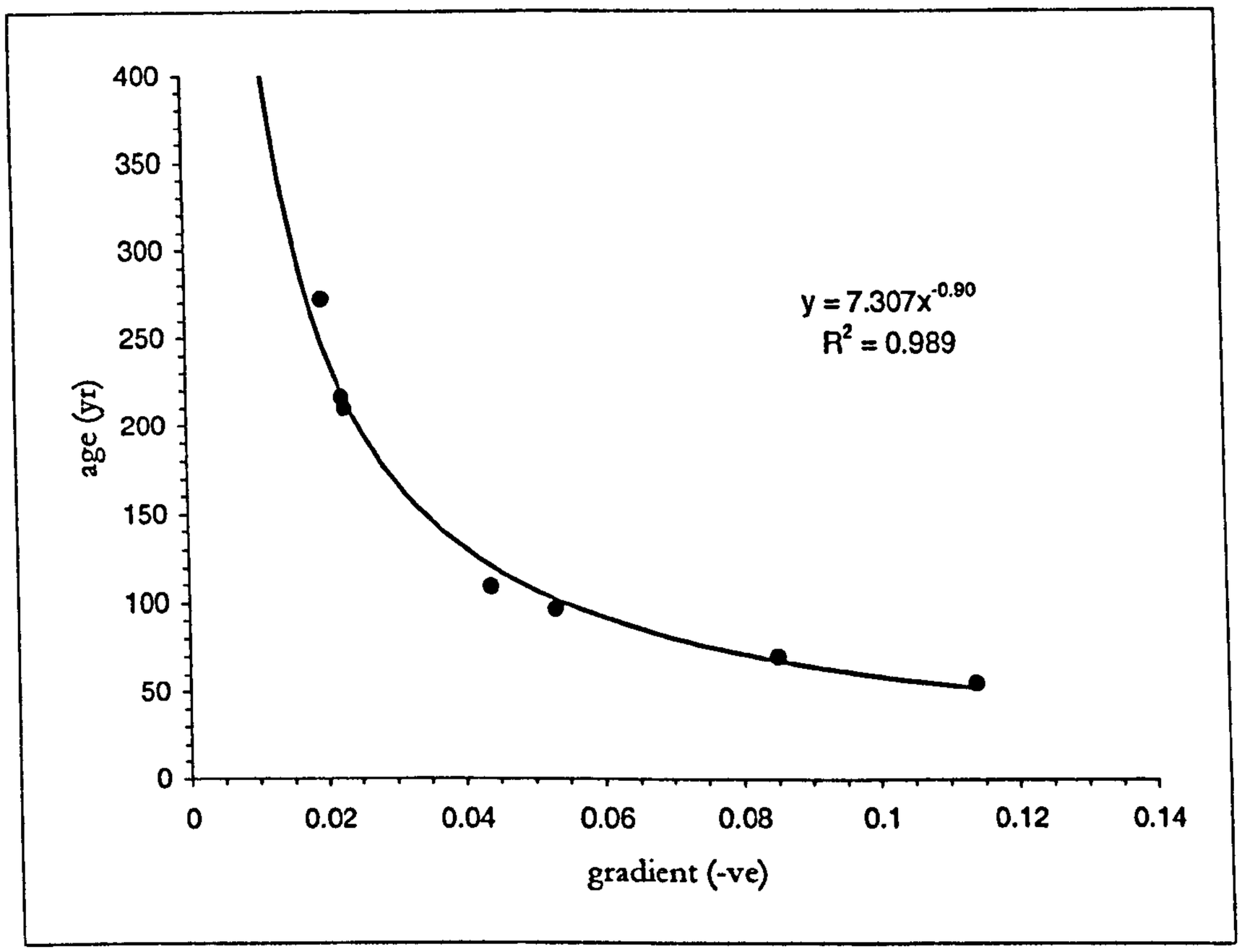


Figure 8.6 The relationship between the gradient of the lichen size-frequency (log_{10}) distribution growing on a surface and the age of the surface. The curve has been calibrated using *Rhizocarpon* Section *Rhizocarpon* lichens growing on surfaces of known age in SE Iceland.
Note that gradients have been multiplied by -1.

This *age-gradient* relationship could be used as a dating curve.

date of surface (AD)	age (yrs)	no. of lichens	gradient
1727	272	307	-0.0199
1783	216	273	-0.0227
1789	210	288	-0.0231
1890	109	275	-0.0440
1903	96	291	-0.0532
1930	69	274	-0.0851
1946	53	299	-0.1139

(above) Data used in the construction of the *age-gradient* dating curve.

Fjallsjökull

The broad, lobate glacier of Fjallsjökull descends steeply from the eastern flanks of Öræfajökull. Formerly coalescent with Breiðamerkurjökull, Fjallsjökull has retreated considerably over the course of the 20th century.

Thorarinsson (1943: 30) summarised the information pertaining to fluctuations of Fjallsjökull (formerly known as Hrútarjökull). In about AD 1850 the glacier was thought to be at the position of the outermost moraines, this being the "maximum extension so far in historical times". Glacial recession since the 1850s was interrupted by various re-advance events, notably around 1870 and between 1890-94 when the glacier was again close to the 1850 terminal moraine (Thorarinsson, 1943: 30). The margin of the glacier was accurately surveyed in 1904. Between this time and 1932 the glacier margin receded a further 300 m. A diligent father and his two sons from the nearby farm have made careful measurements of ice-marginal retreat at Fjallsjökull since the early 1930s. A good record of ice-front positions at Fjallsjökull therefore exists for the period AD c. 1850-1998.

Hólárjökull

A small, steep, southern outlet of Öræfajökull, Hólárjökull once reached the surrounding farmland forming a large, arcuate terminal moraine complex. Presently, the glacier terminates on the lower flanks of the Öræfajökull massif. Thorarinsson (1956) studied the moraines left by Hólárjökull and concluded, based on the presence of certain tephra layers in the soil, that the largest, outermost ridge was deposited in pre-historic times (ie. before 900 AD). Details of ice-front fluctuations at Hólárjökull are sporadic. Thorarinsson (1956: Fig. 7) states that the ice probably reached the large terminal moraine in c. AD 1870. A photograph taken in 1890 or 1891 (Howell, 1893) shows the glacier to be less extensive than in c. 1870. By 1904, when the glacier was surveyed by the Danish General Staff, the ice-margin had receded almost 500 m since 1870. The glacier was monitored between 1933 and 1939 by a local farmer. Further evidence regarding the position of the terminus is available in the form of aerial photographs taken regularly since 1945. A record of ice-front positions at Hólárjökull exists for the period c. AD 1870-1998.

Yellow-green *Rhizocarpon* Section *Rhizocarpon* lichens were measured on seven moraine surfaces at Fjallsjökull (Fig. 8.7) and five at Hólárjökull (Fig. 8.8). All but the oldest moraine, at each glacier, were of approximately known age before the experiment was conducted. Only one surface dates from the last 75 years. The same method used to construct the dating curve was adopted in all the lichenometric surveys. Size-frequency (log_{10}) distribution graphs were generated for each of the twelve populations (Figs. 8.9 and 8.10). The gradient of the population distribution (greater than the modal frequency) was calculated using least-squares regression analysis. The results are presented below.

Table. 8.2 – Lichenometric data from Fjallsjökull

Moraine	Time of surface exposure	Age (yrs)	Gradient (-ve)	Largest-lichen (mm)	Date (based on 'gradient' dating curve)	Date (based on largest-lichen dating curve)
1	unknown	> 150	0.0232	85	1784	1785
2	c. 1850s	c. 150	0.0347	66	1850	1855
3	c. 1870s	c. 130	0.0422	61	1872	1871
4	1890-95	104-109	0.0551	49	1900	1895
5	1904	95	0.0577	47	1904	1903
6	1908-12	87-91	0.0602	44	1907	1910
7	1915-19	84-80	0.0688	42	1918	1915

Table. 8.3 – Lichenometric data from Hólárjökull

Moraine	Time of surface exposure	Age (yrs)	Gradient (-ve)	Largest-lichen (mm)	Date (based on 'gradient' dating curve)	Date (based on largest-lichen dating curve)
1a	unknown	> 130	0.0325	74	1849	1840
1b	c. 1880	c. 120	0.0378	70	1860	1850
3	1892-1904	95-107	0.0418	47	1893	1903
5	c. 1915-25	79-84	0.0504	38	1919	1926
7	1935	64	0.1048	30	1943	1937

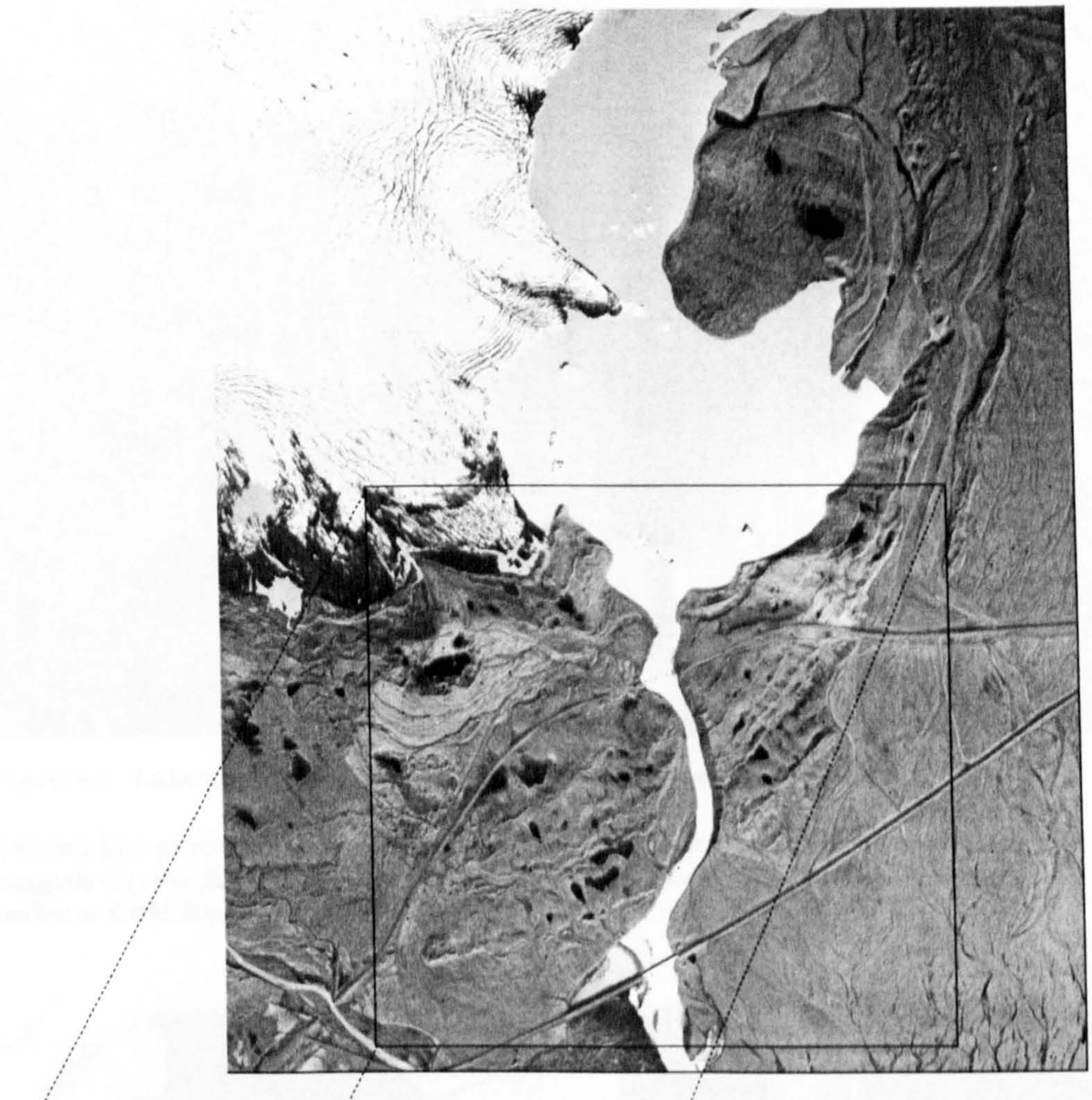


Figure 8.7 Lichenometry on the foreland of Fjallsjökull, southeast Iceland.

(Above) Part of an aerial photograph of Fjallsjökull, taken in 1998.

(Left) Map of moraines representing the former extent of Fjallsjökull. Lichenometric size-frequency studies were conducted on moraines 1-7. The present-day glacier margin (1998) is located c. 700 m from moraine 7.

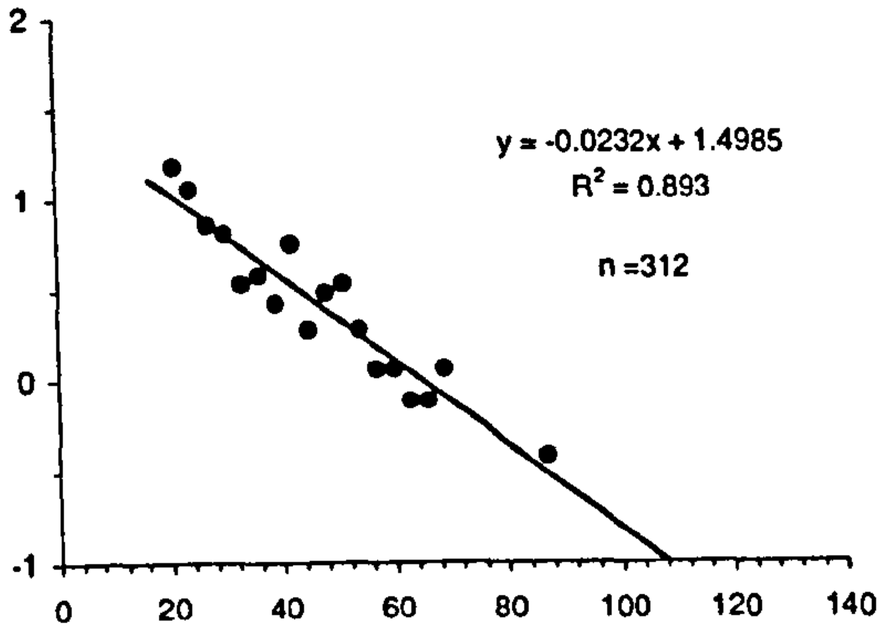
Figure 8.8 Lichenometry on the foreland of Hólárjökull, southeast Iceland.

(Above) This photograph was taken from the crest of moraine 1 in 1999. Thorarinsson (1956) suggests that the glacier was up to this point in c. 1870, however the moraine itself dates from earlier in the Holocene, possibly c. 2500 BP.

(Above) Map of Hólárjökull compiled by Thorarinsson (1956: Fig.7). The position of the ice margin c. AD 1870 and 1904 are shown. The glacier retreated considerable in the 1930s and 1940s. In 1955 the snout lay at 260 m asl. In 1999 the glacier terminus was at 150 m asl (see photo, above).

(Above) Part of an aerial photograph of Hólárjökull taken in July 1998. The moraines sampled in this study are numbered 1-7.

Fjallsjökull 1



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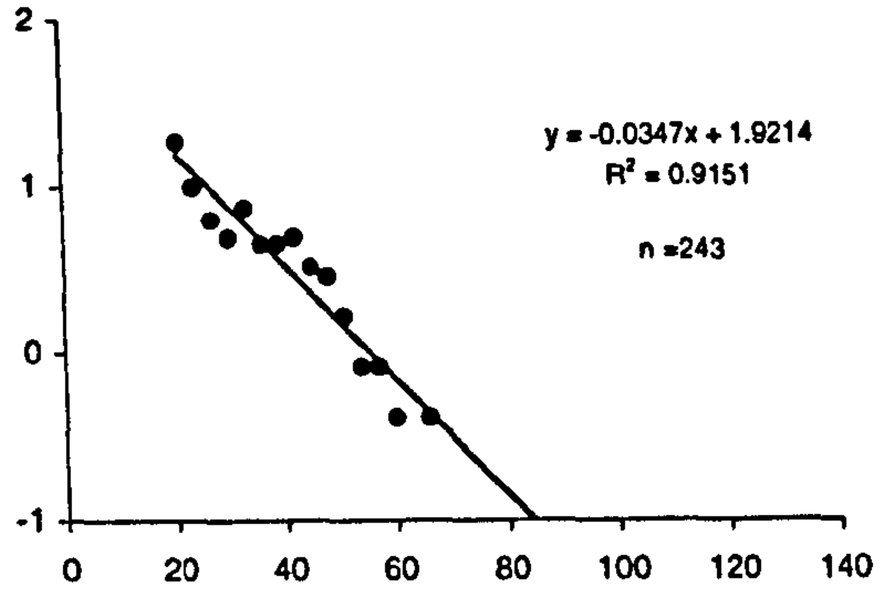

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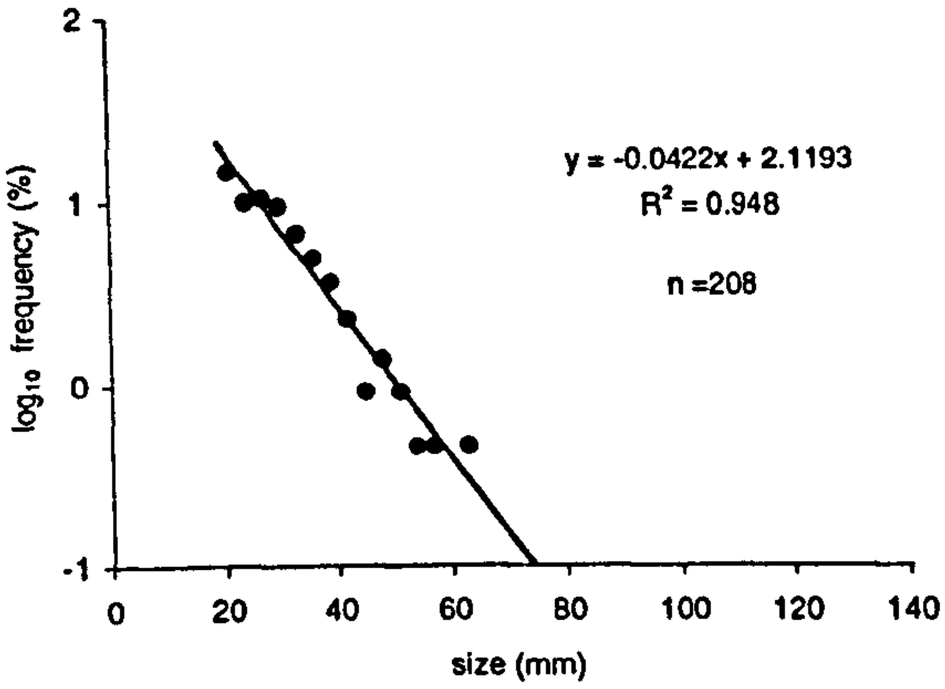

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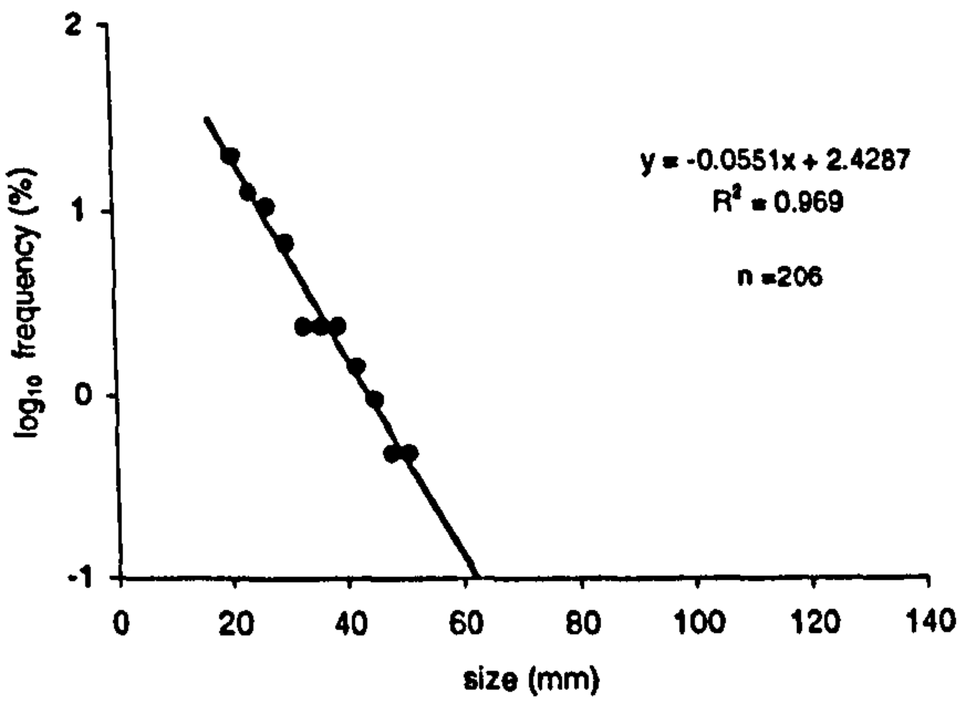

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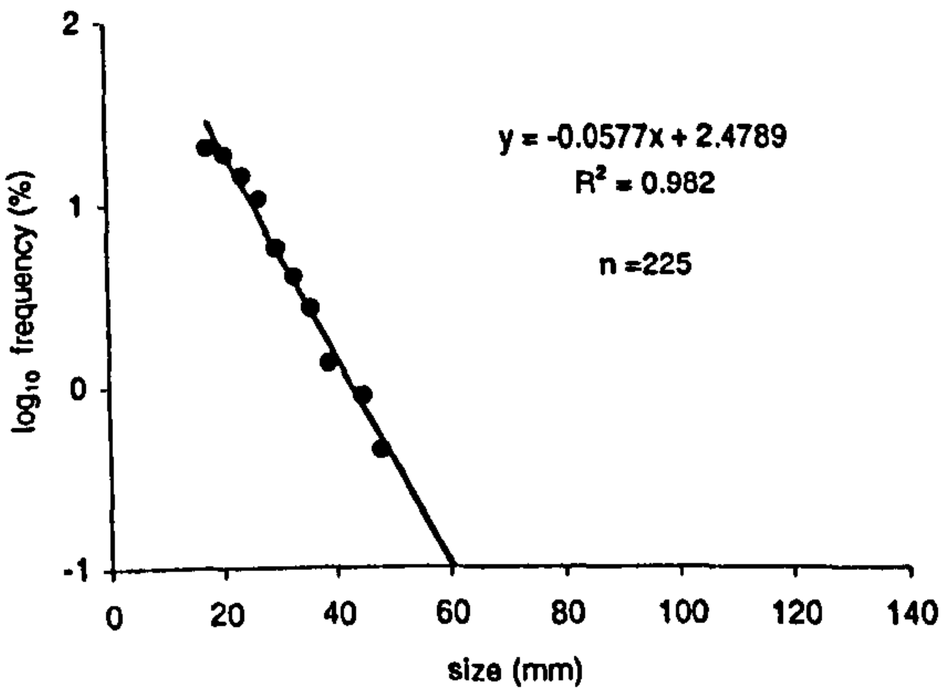

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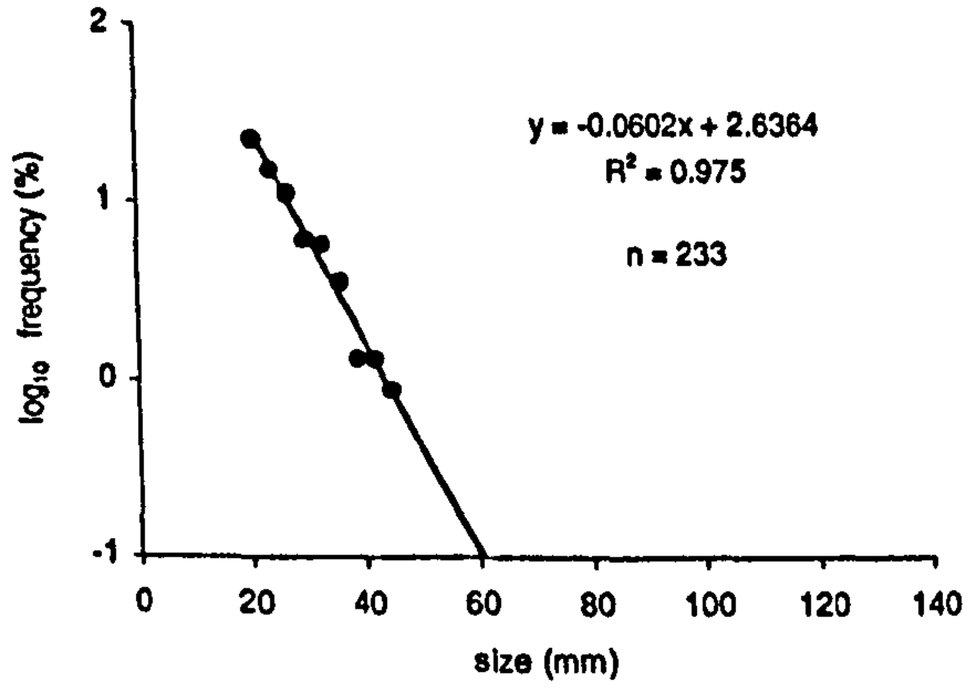
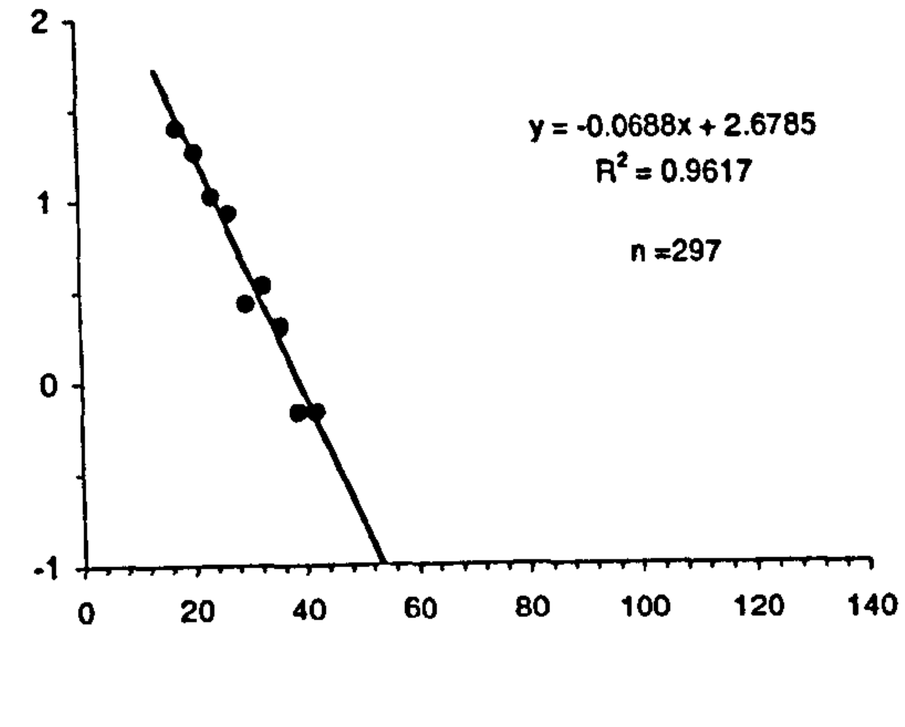
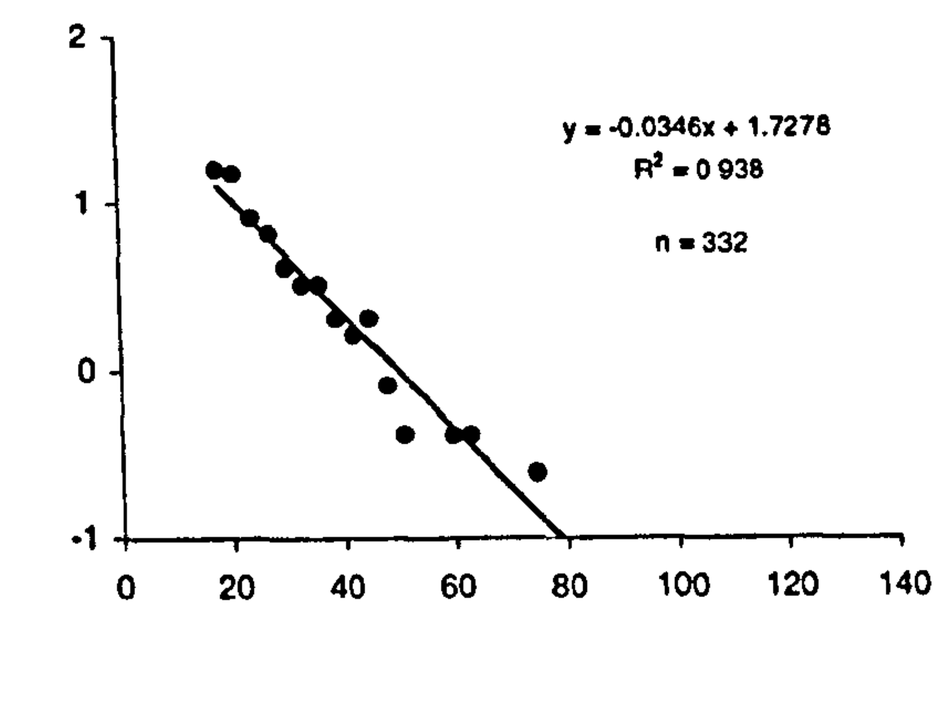
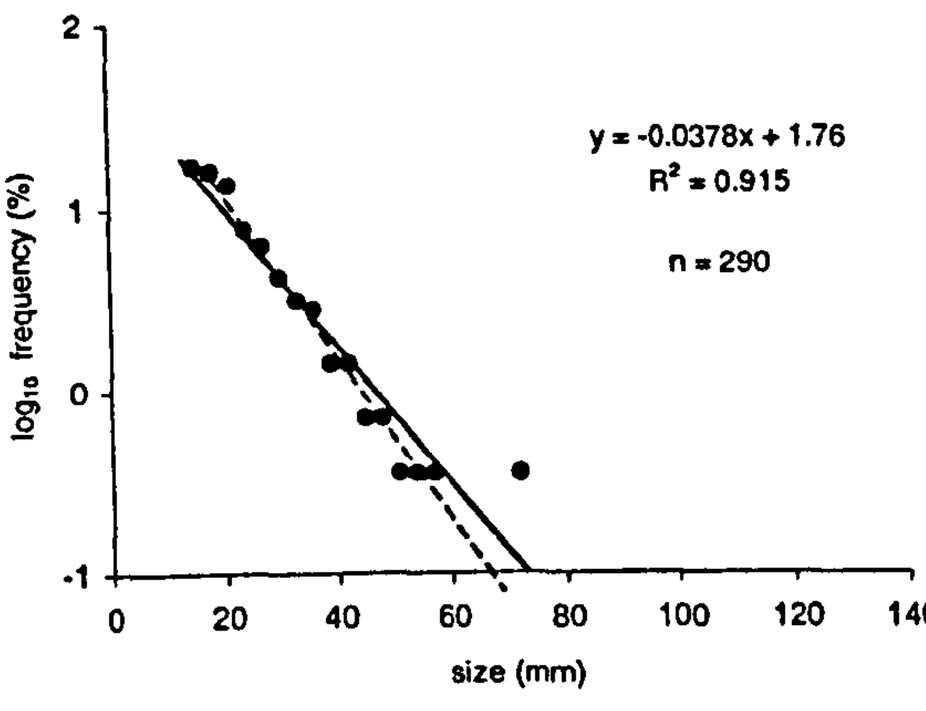
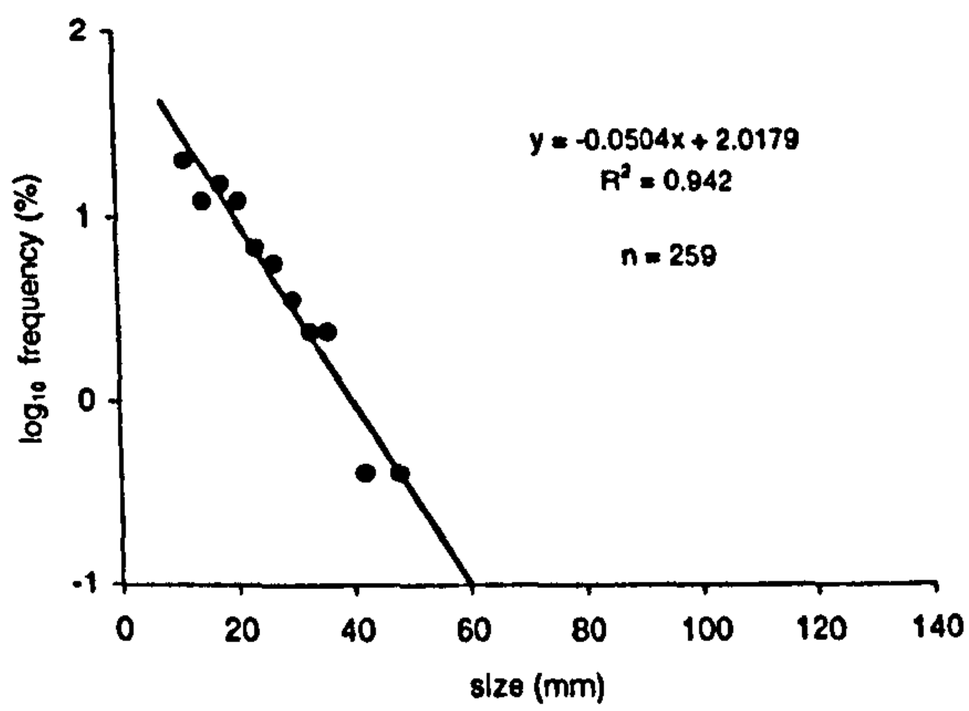


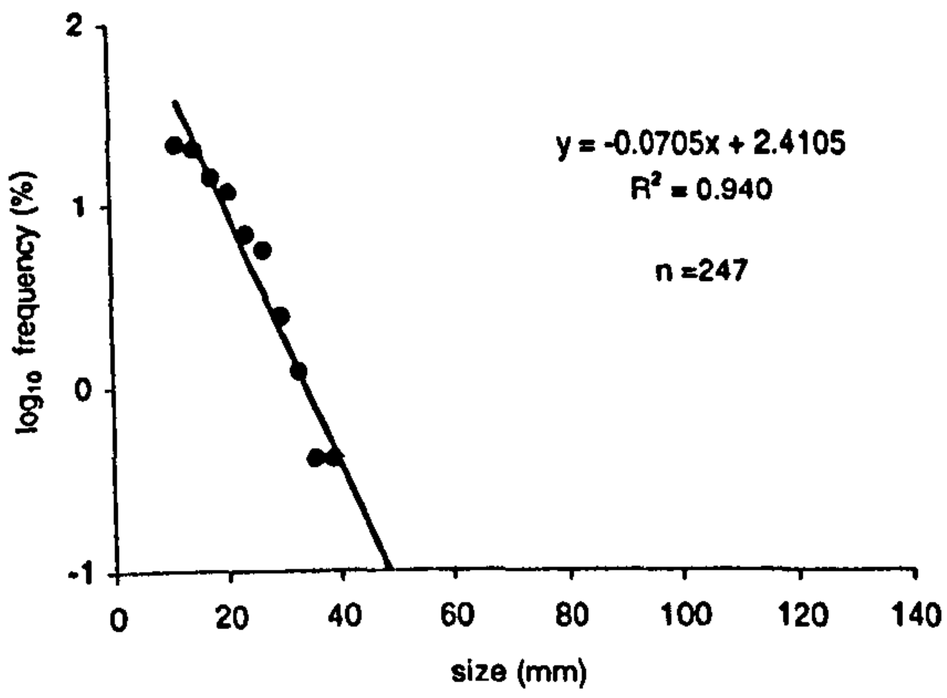
Figure 8.9 Lichen size-frequency graphs from the foreland of Fjallsjökull











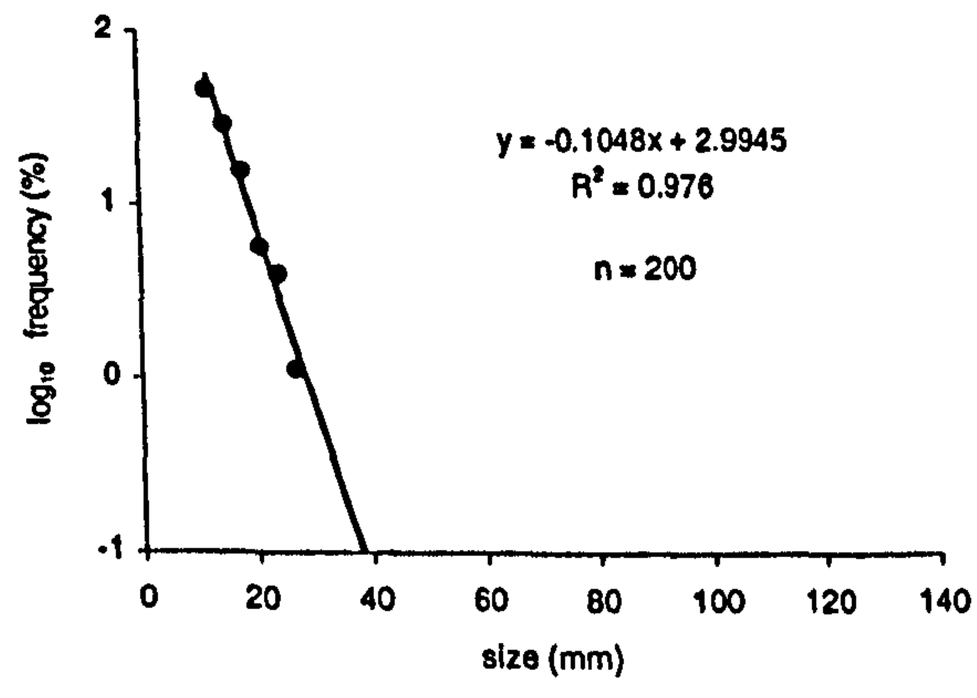


Figure 8.10 Lichen size-frequency graphs from the foreland of Hölárjökull

The results of the lichen-dating experiment are interesting. Both the *largest lichen within a fixed area* and the *gradient of the size-frequency distribution* yield very close age estimates for all ten recently-deposited moraines. Both techniques estimate the age of the moraines generally to within five years, and to within one or two years in certain cases. The only inaccurate age estimation is for moraine *1b* at Hólárjökull – a distinct trimline on the proximal flank of the large terminal moraine. A 'largest lichen' of 70 mm yields a surface age of AD 1850 according to the age-size dating curve (Fig. 8.3). However, the lower part of the proximal slope of Moraine 1b was ice-covered in c. 1870 and was probably not ice-free until the late 1870s, at the earliest. The gradient of the population distribution indicates that the surface was exposed in AD 1865. It is possible that the large, 70 mm lichen belongs to an older population and is anomalous, and thus produces an inaccurate age estimate. If the 70-mm thallus is omitted from the analysis, *largest-lichen* and *gradient* estimates produce dates of 1880 and 1888, respectively. It is hoped that the use of both techniques in tandem will give a more reliable, more considered, age estimate than lichenometric dating based on a single, or the mean of several, large lichen.

These results suggest that Benedict's (1985: 106) assertion that the size-frequency analysis method is "not likely to prove useful in temperate or maritime environments" is incorrect. The technique can successfully distinguish between lichen populations of different ages in southeast Iceland, at least on surfaces formed during the last 150 years or so.

The oldest moraines at Fjallsjökull and Hólárjökull are presently undated. The small, low-elevation, moraine fragment, identified by the author, located 60 m outside moraine 2, dates from the late 18th century according to both the size of the largest lichen and the gradient of the lichen size-frequency distribution. No other moraines dating from this period (AD 1770-1800) have so far been identified in southeast Iceland.

Moraine 1a at Hólárjökull is a large, complex feature, possibly dating from the Sub-Atlantic time (c. 2500 BP) (Thorarinsson, 1956). Öræfajökull-1362 ash was found in soil sections on its distal flank. Lichenometry was conducted on the uppermost part of the proximal flank where the soil cover was thin and patchy. The *gradient* technique

yielded a date of 1836, and the *largest-lichen* technique placed the time of exposure around 1840. These results suggest that the glacier re-occupied the prehistoric terminal moraine during the first half of the 19th century. Lichen size-frequency distributions from the distal flank of the terminal moraine ridge could not be interpreted using the gradient technique, indicating a mature, composite, lichen population on an old surface.

A further test of the size-frequency *gradient* dating-technique on another old surface also yielded rather complex results. Boulders close to the western margin of Virkisjökull deposited by a *jökulhlaup*, originating from the Öræfajökull volcano in AD 1362, were examined for *Rhizocarpon* lichens. The resulting size-frequency distribution is difficult to interpret, as it does not conform to the negative log-linear model (Fig. 8.2). The lichen cohort distribution is bimodal and better resembles a log-normal or Poisson model. This may be due to the rocks possessing an inherited lichen population. However this hypothesis is deemed unlikely as the transportation of rocks by high-energy flood waters – as witnessed in 1996 on Skeiðarársandur – will usually remove any saxicolous lichens by scouring and abrasion. It is more likely that the Poisson lichen distribution identified at Virkisjökull is in agreement with the findings of Innes (1983b, 1986), who argues that mature lichen populations do not conform to the log-linear model. Use of regression analysis and the *gradient* approach in such situations would be inappropriate.

Summary – A new lichen-dating approach

Several workers have advocated the use of a population approach in lichenometry. However, relatively few have stressed the advantages of the **gradient** of the size-frequency distribution as the most representative age descriptor (Benedict 1967, 1985; Innes 1983b, 1986; Caseldine 1990). Critical testing of a dating curve, calibrated using lichen population distribution gradients, shows that the technique is generally applicable in maritime climates, at least over the last two or three centuries. However, the technique failed to estimate correctly the age of a 600-year-old surface, due to the nature of mature lichen population distributions, and therefore may not be appropriate for dating older surfaces in southeast Iceland.

The age of moraines in Hoffellsdalur, Skyndidalur and Fossdalur

In view of the success in constructing and testing the dating curves, two different lichenometric approaches were used to date moraines around Lambatungnajökull. Surface exposure dates for moraines have been derived using the *largest lichen (in a fixed area)* and the *gradient of the size-frequency (log_{10}) distribution.* The results of the lichenometric surveys on 38 moraines are shown in graphical and table form (Figures 8.15 – 8.14 and Tables 8.4, 8.5, 8.6 and 8.7). The findings are summarised in Figure 8.15.

The results are revealing. The *largest lichen* technique produces younger exposure ages than the *gradient* technique in all but a few cases. According to the *gradient* technique, the outermost moraine in Hoffellsdalur on which lichenometry was conducted yields a surface exposure date of AD 1796. A similar age of 1805 was derived for the outermost lateral moraine in Skyndidalur (S4). Later recessional moraines in both valleys were also exposed coincidentally around 1857, 1880, 1898 and 1930, according to the *gradient* technique. A closely-spaced sequence of minor moraines formed in Skyndidalur between 1898 and 1948. Those formed since 1931 are sawtooth in plan. 13 ridges formed in the years between 1931 and 1945, indicating that they are *annual moraines*. Moraines relating to the small cirque glacier on Grasgiljatindur were exposed in 1864, 1898 and 1926. The pattern and rate of deglaciation in the valleys adjacent to Lambatungnajökull, since the *Little Ice Age* maximum, has been illuminated by the application of lichenometry. Over the last 200 years or so, the technique appears to be a highly valuable dating tool.

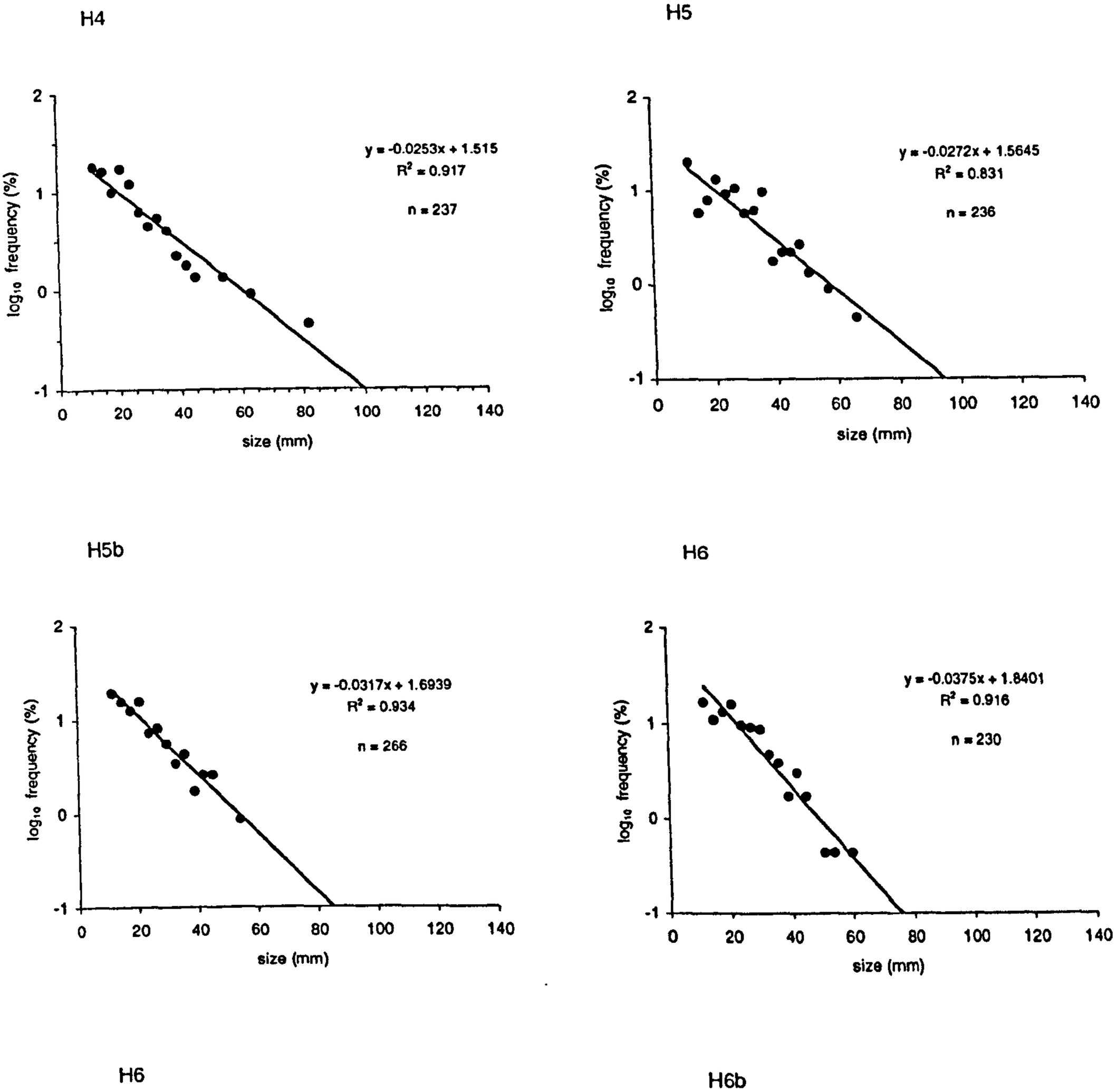
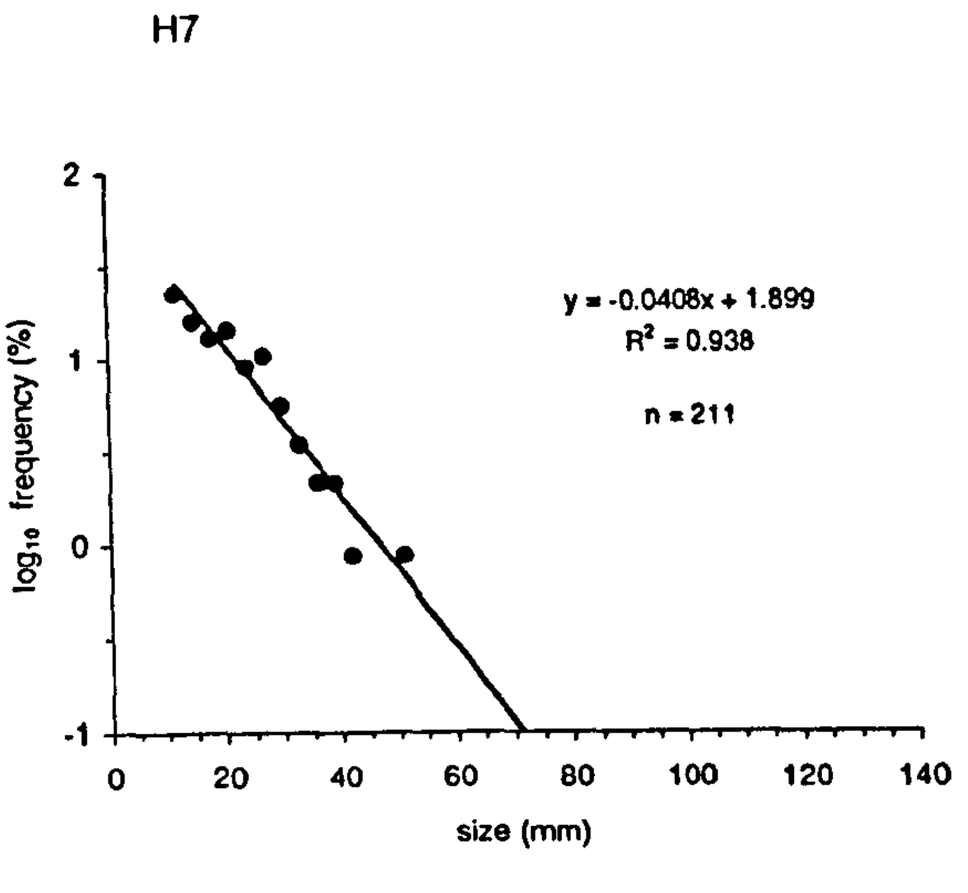
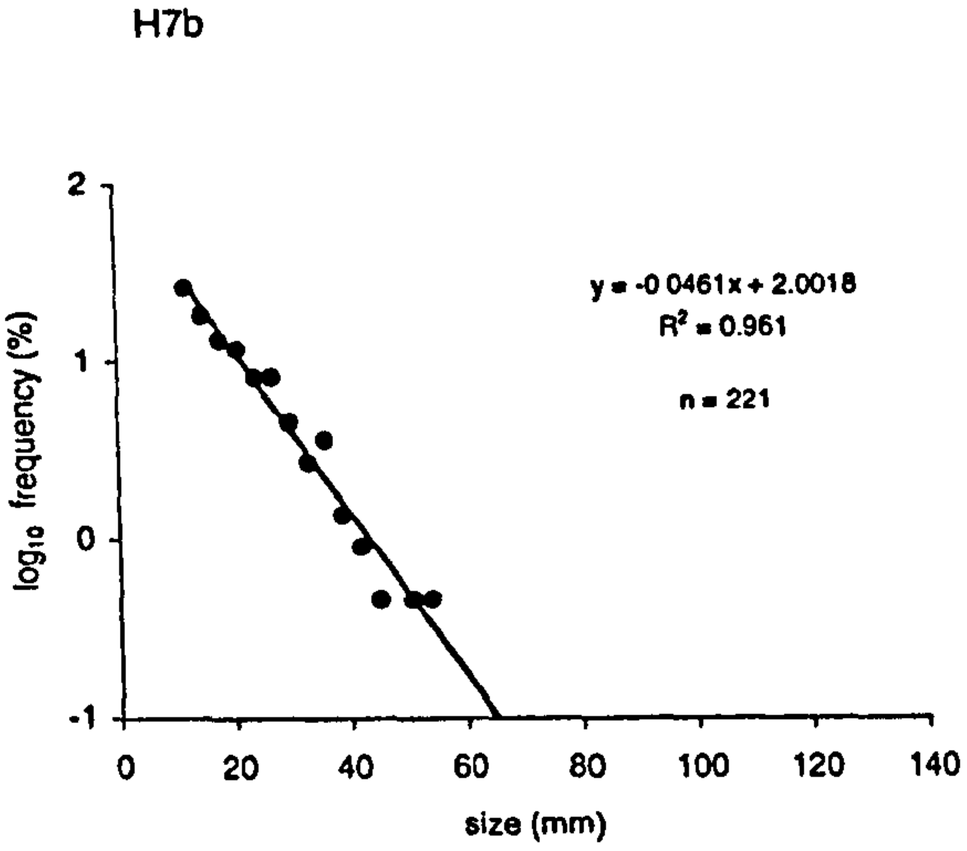
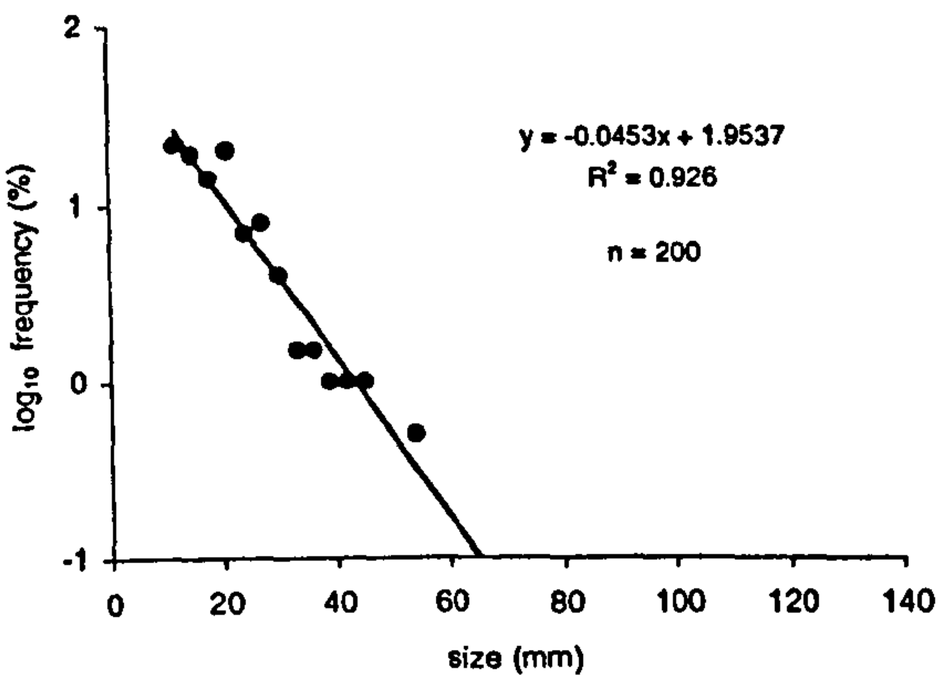
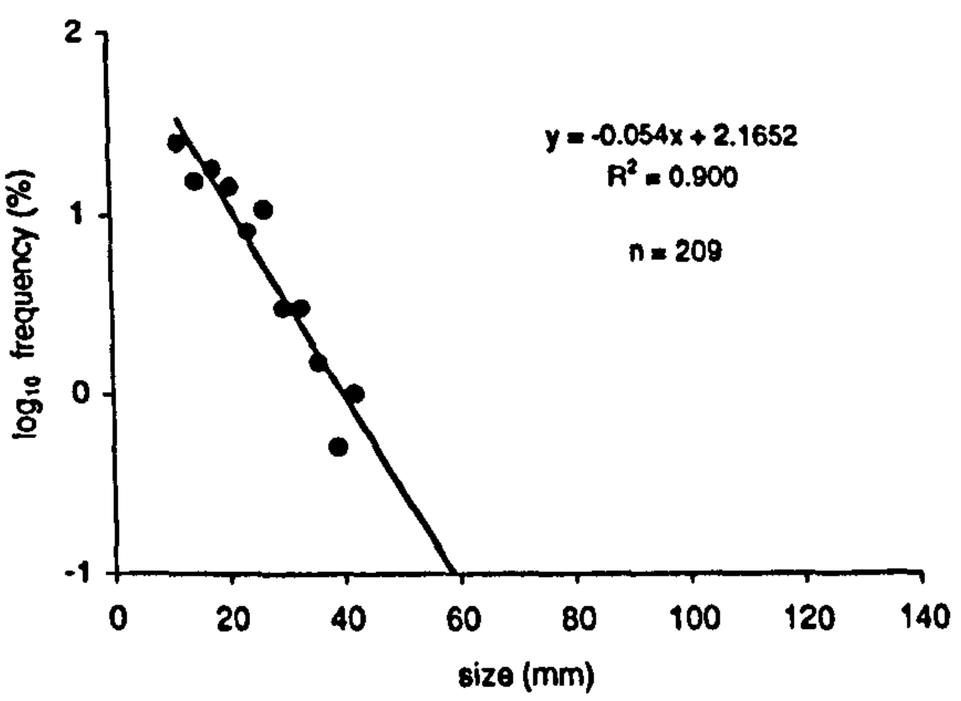


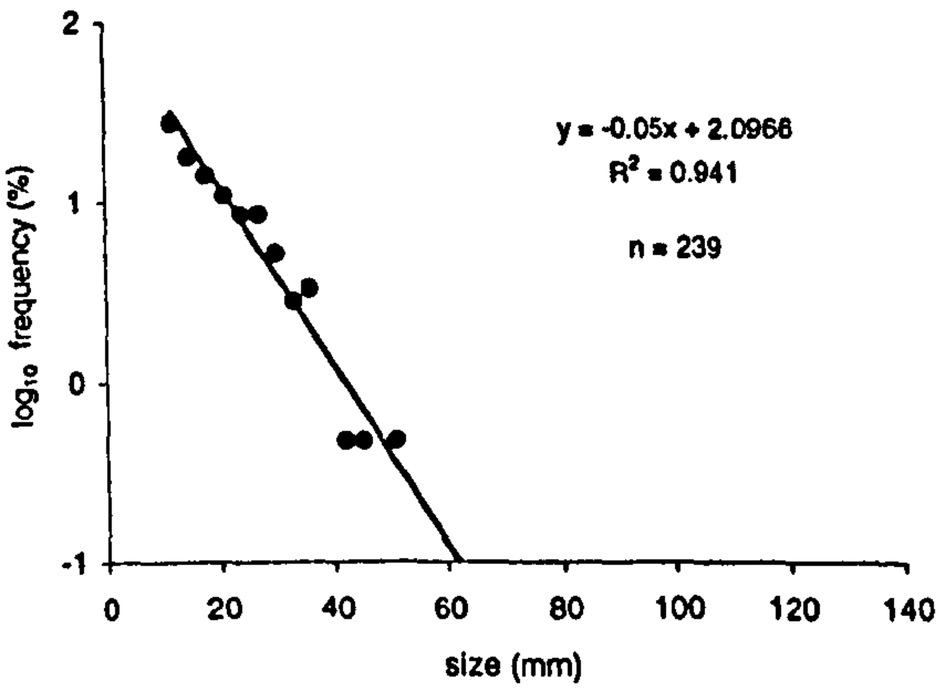
Fig. 8.11 Lichen size-frequency distributions from moraines in Hoffellsdalur











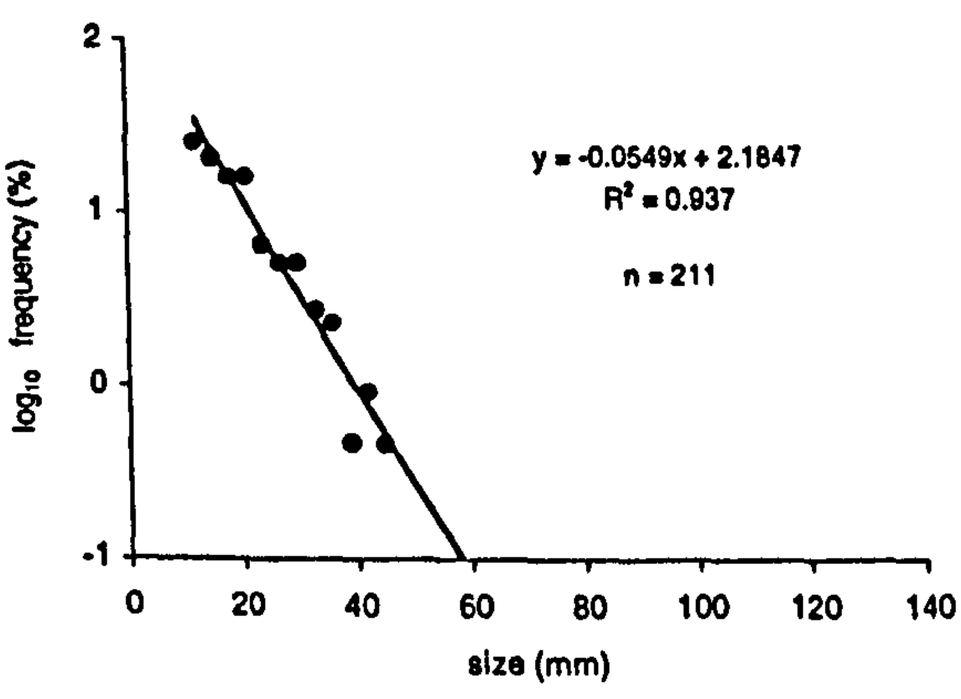
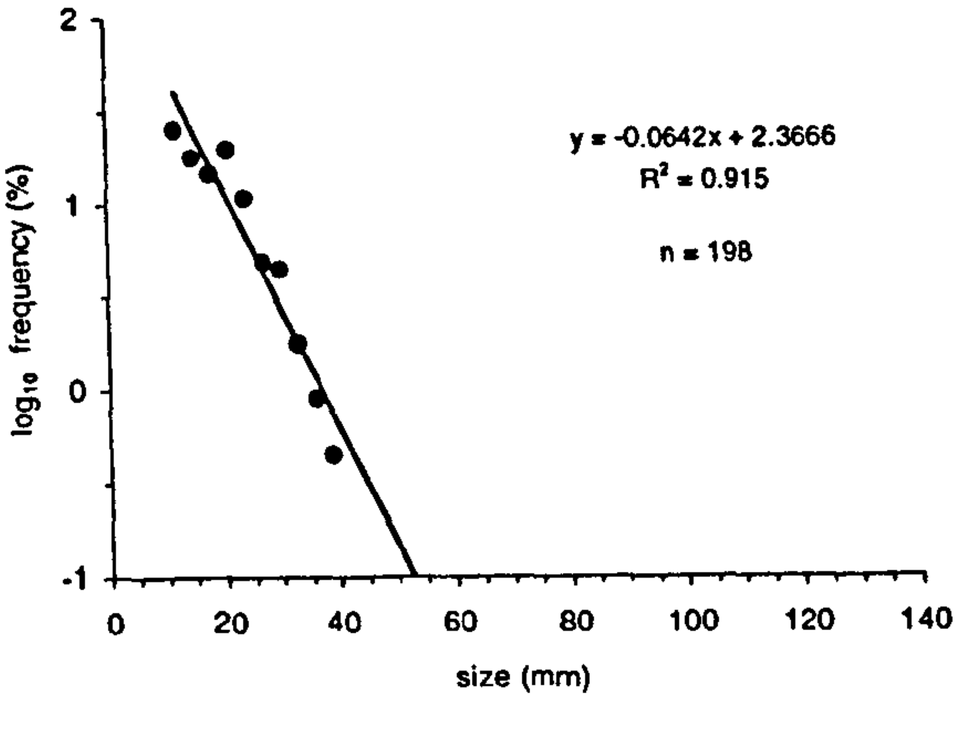
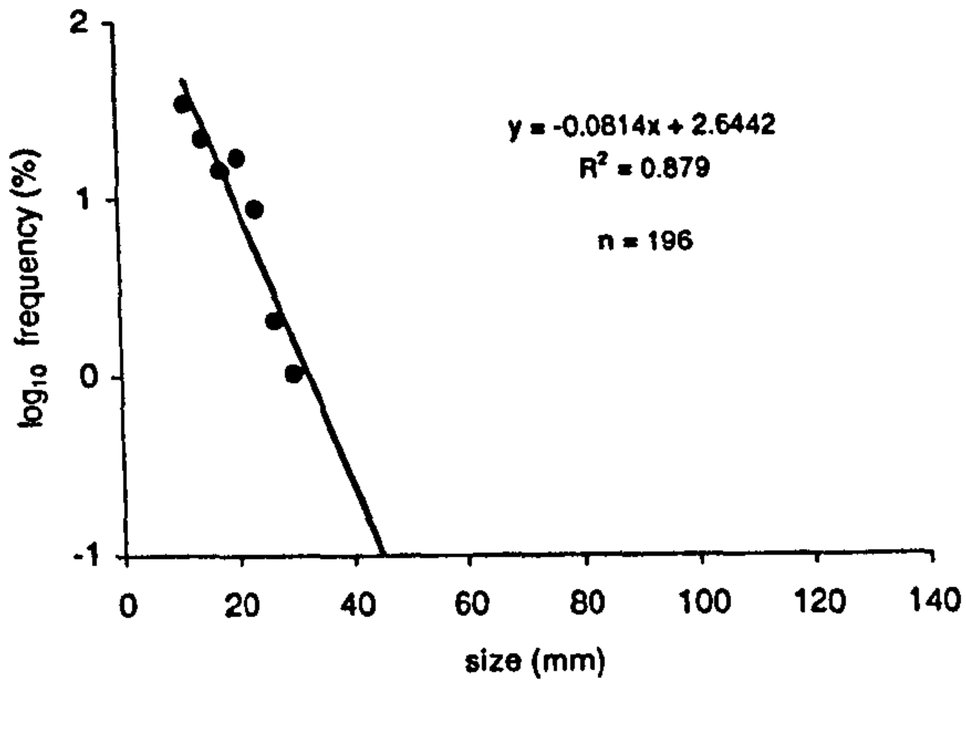
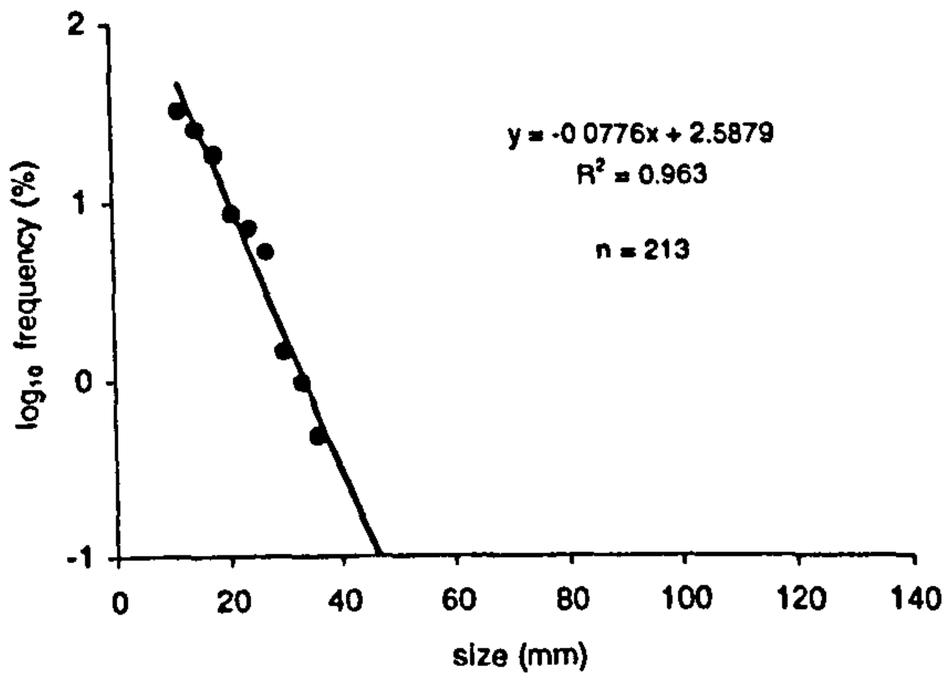
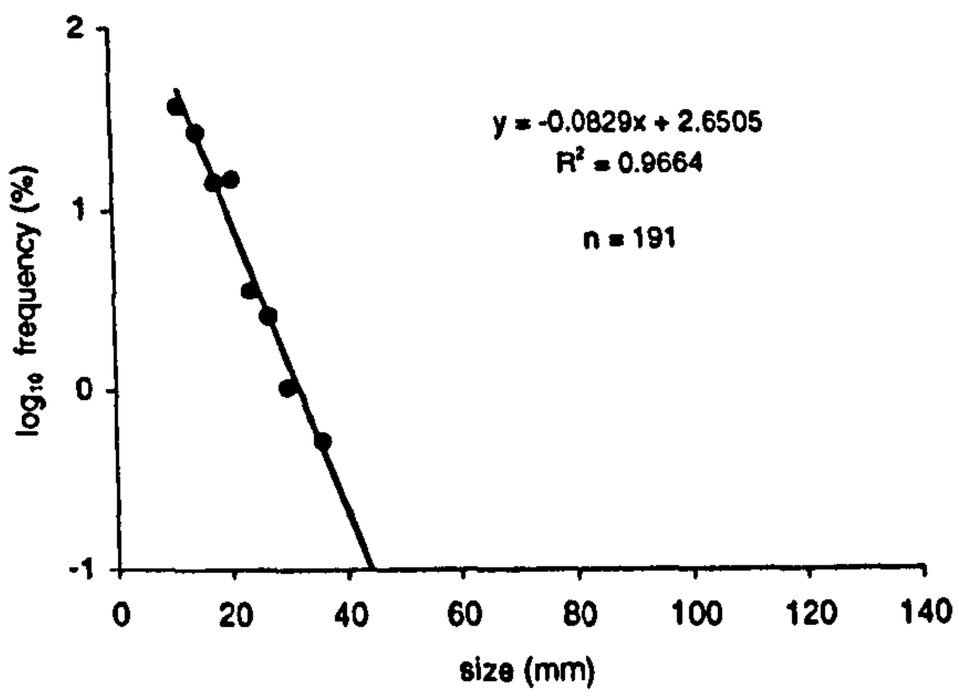


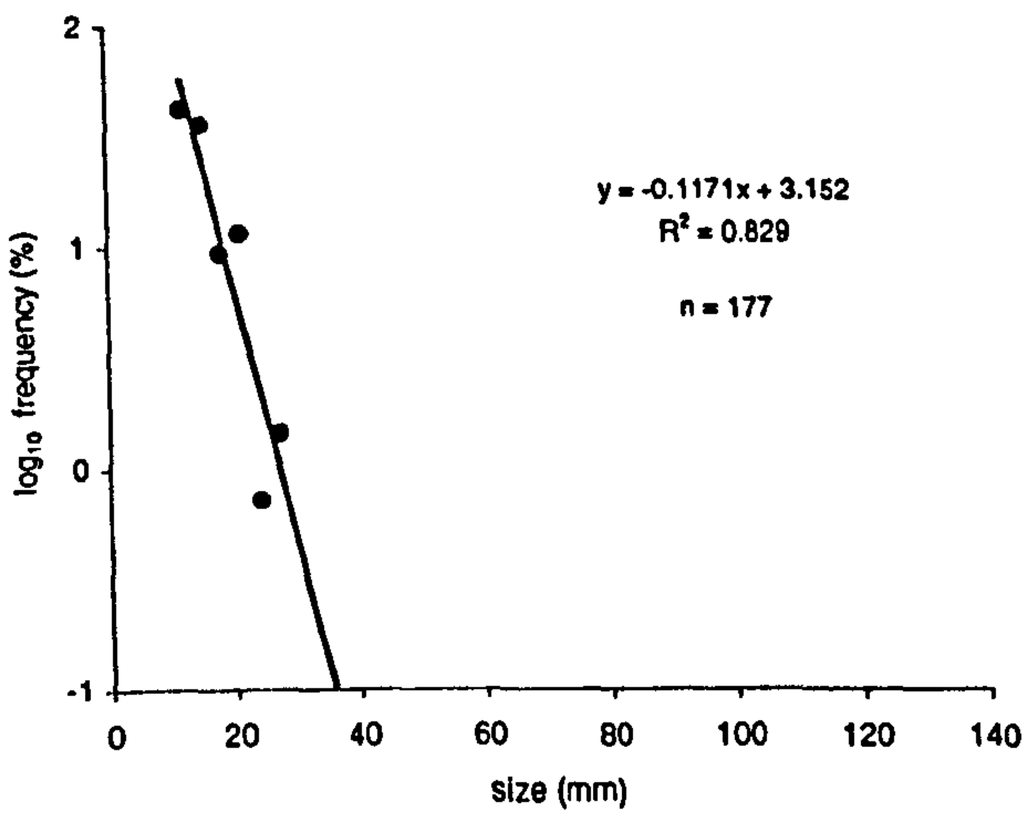
Fig. 8.11 Lichen size-frequency distributions from moraines in Hoffellsdalur











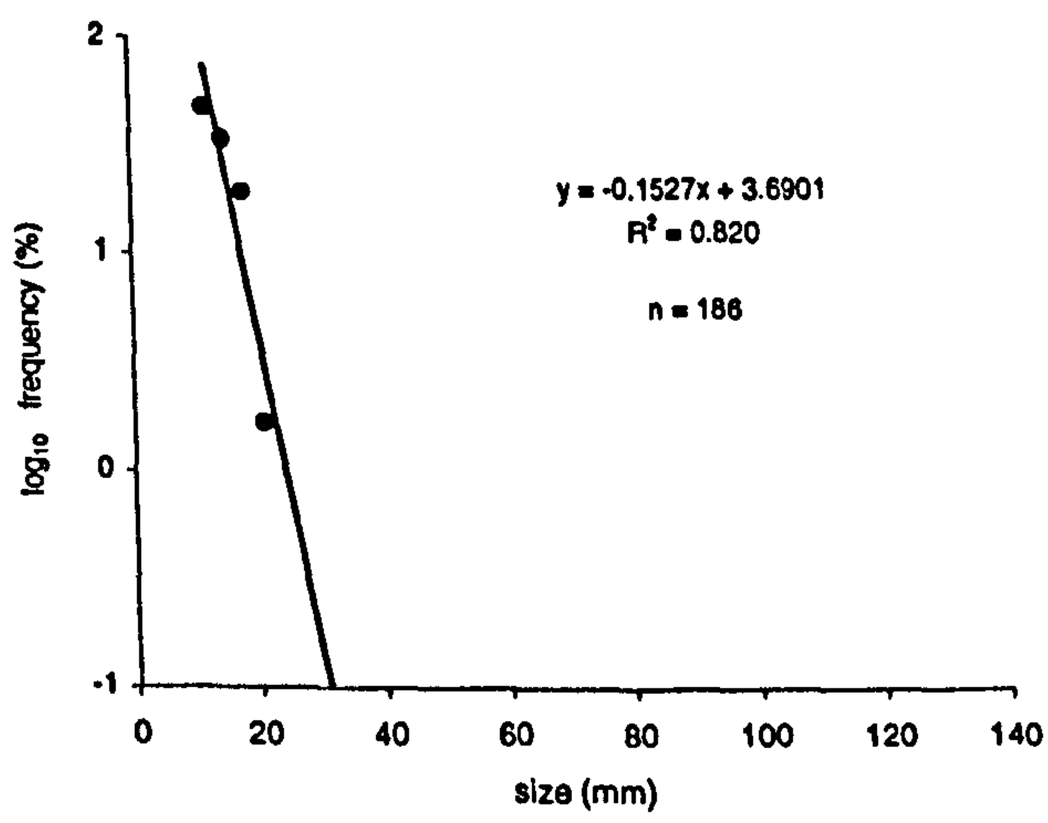
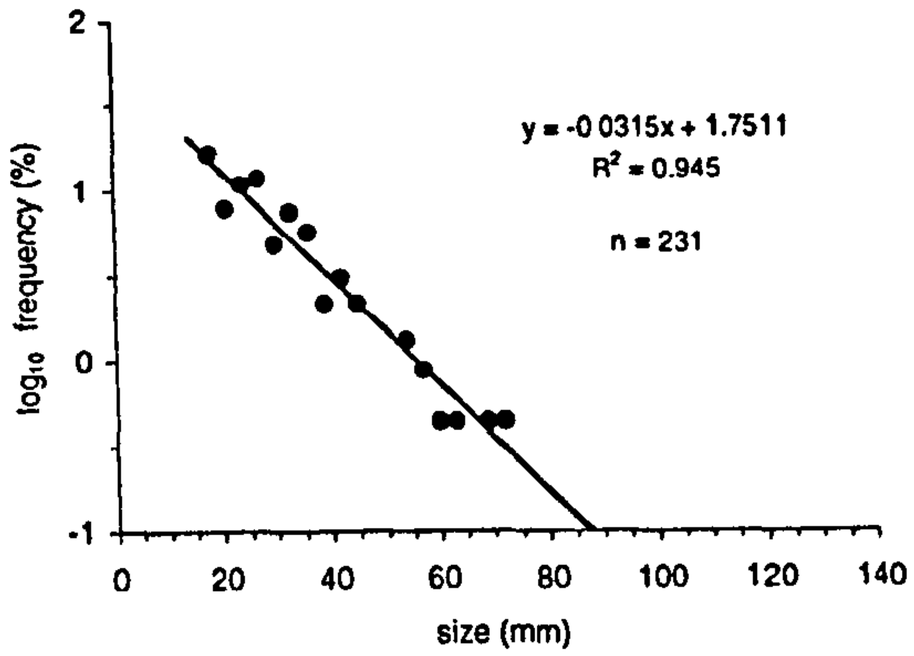
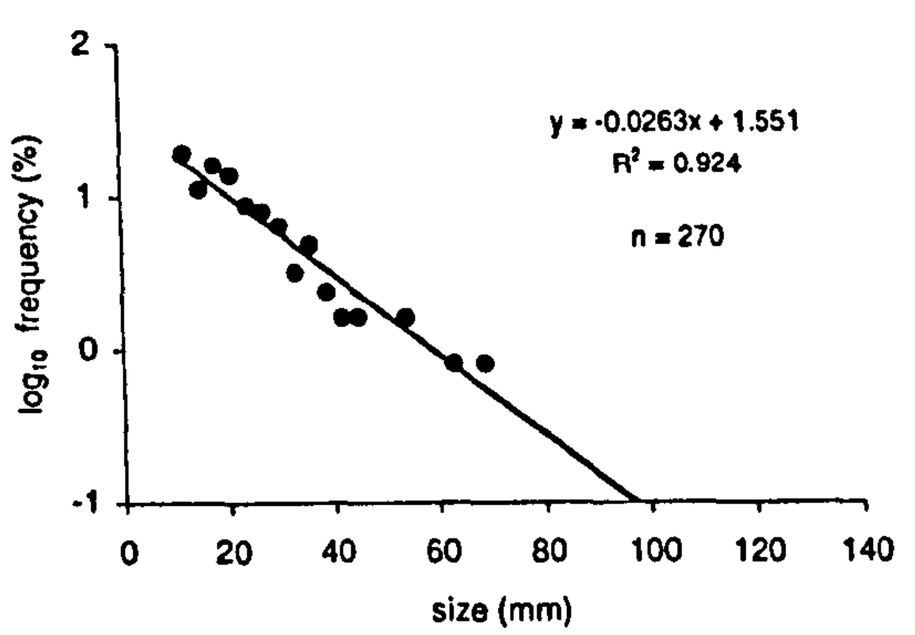
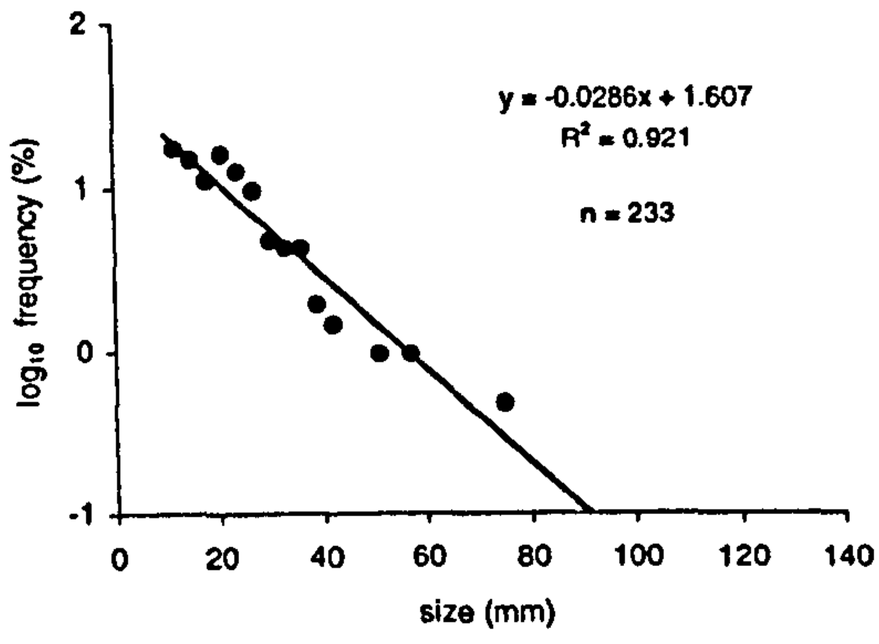
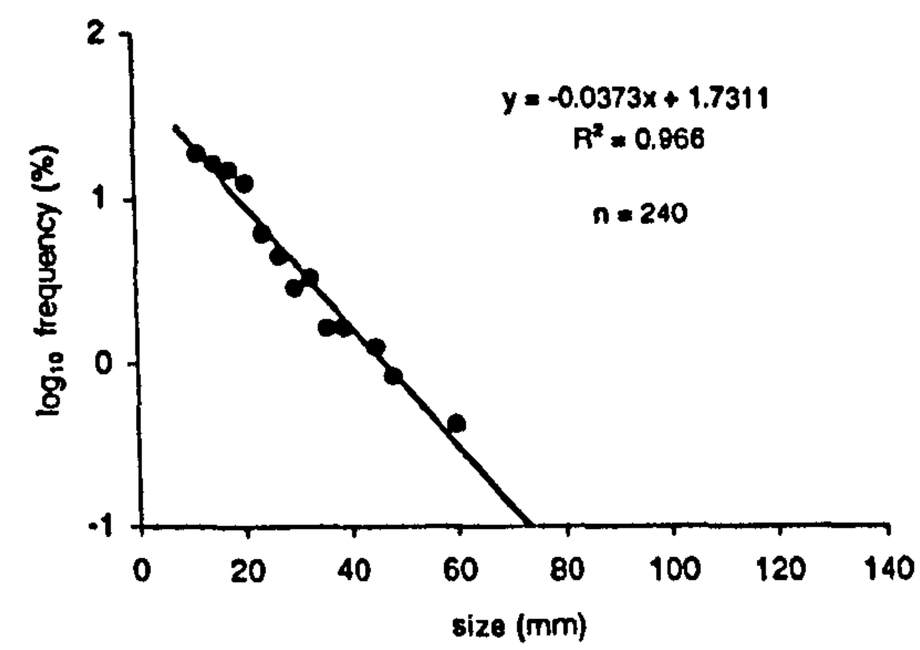


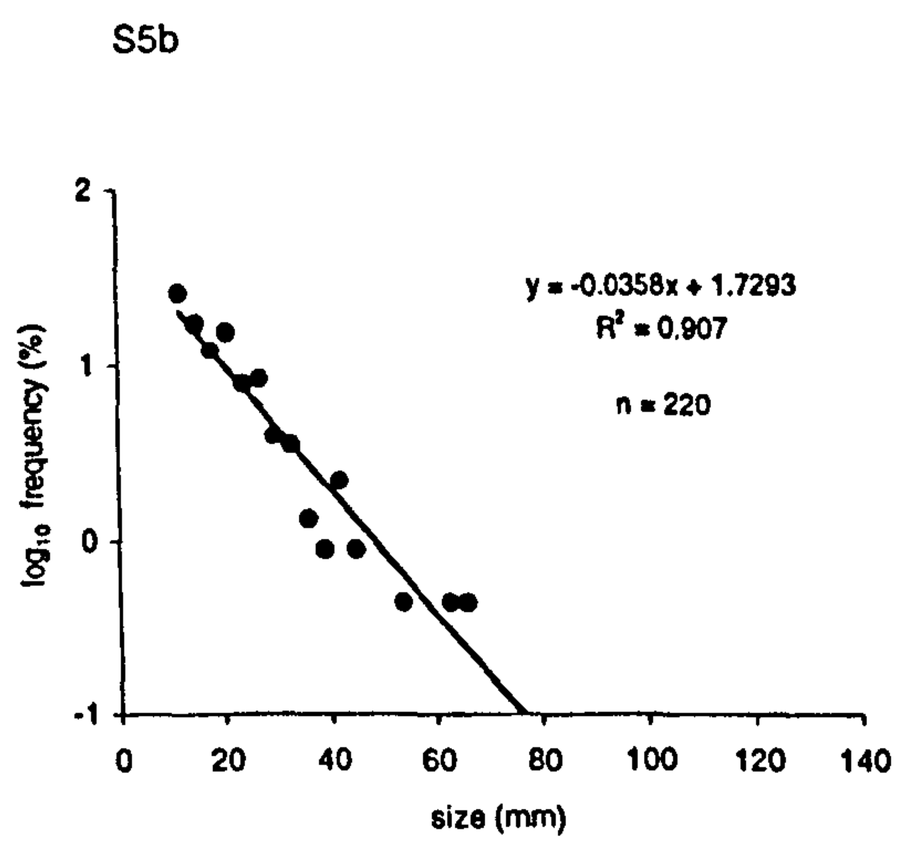
Fig. 8.11 Lichen size-frequency distributions from moraines in Hoffellsdalur











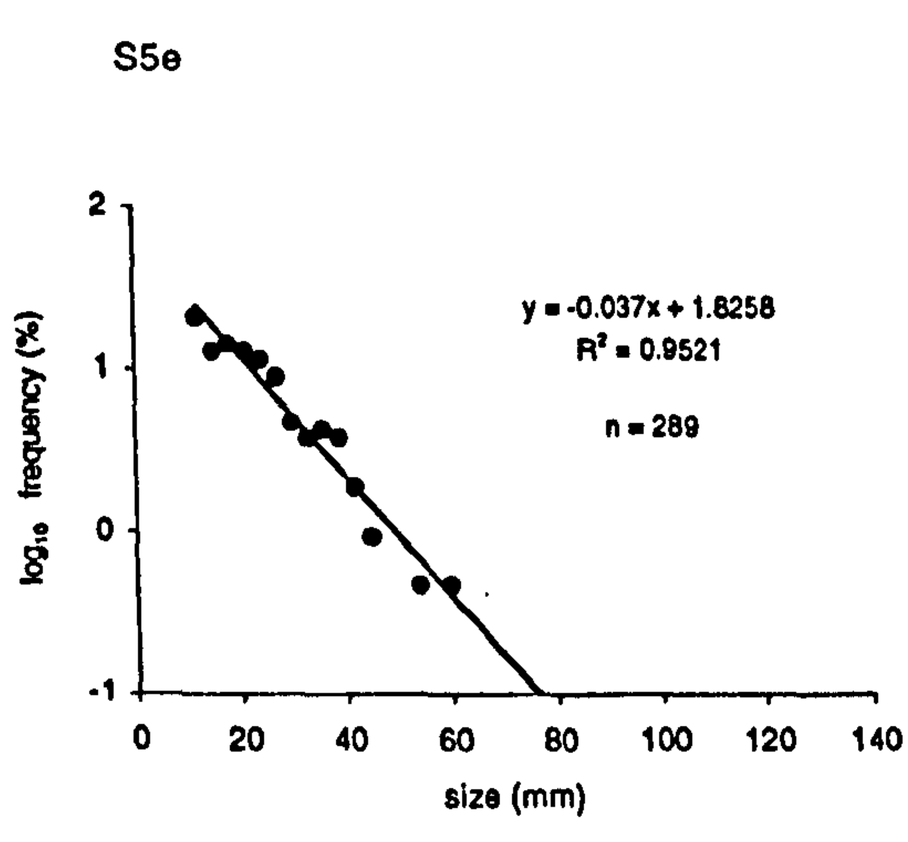


Figure 8.12 Lichen size-frequency graphs from moraines in Skyndidalur

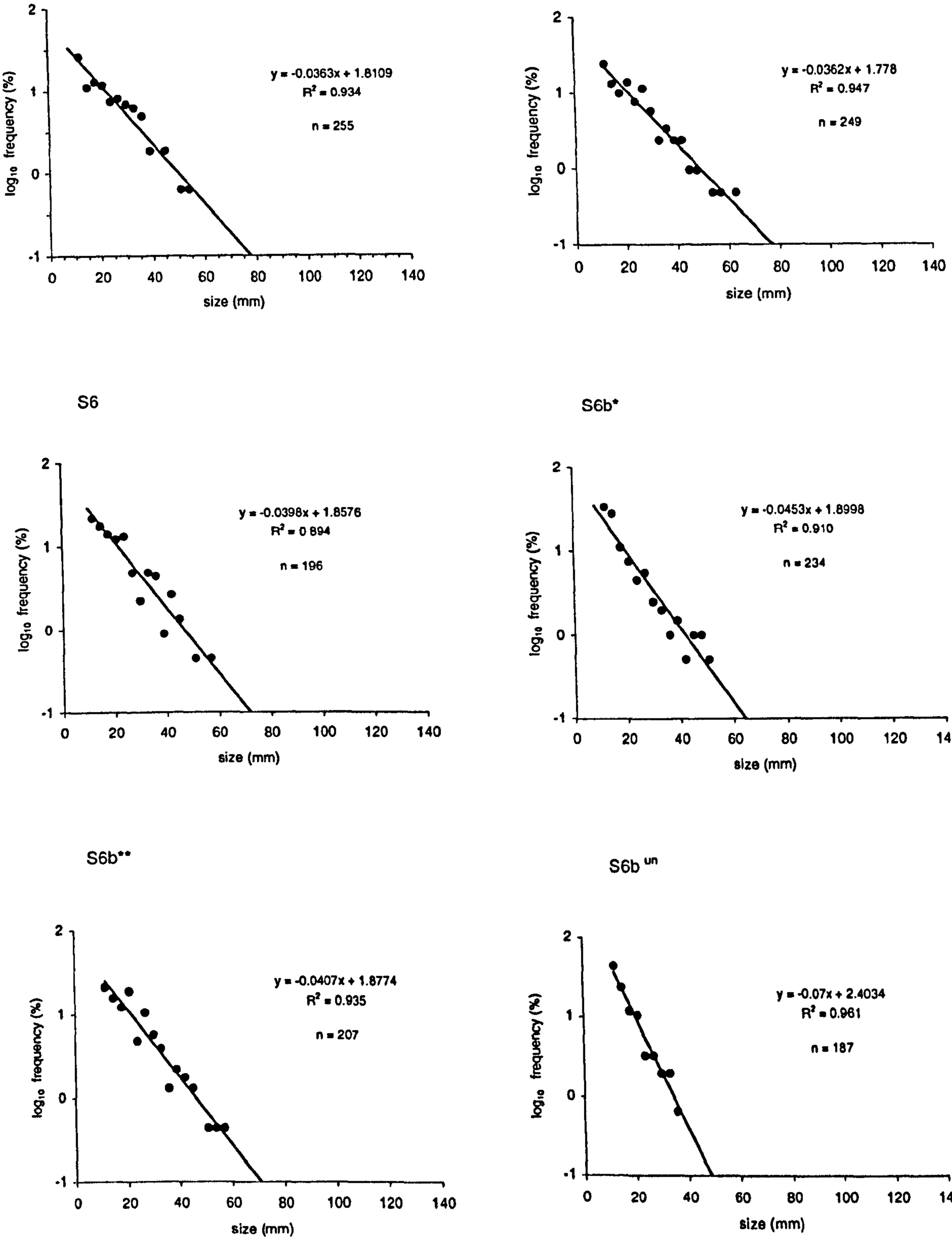


Figure 8.12 Lichen size-frequency graphs from moraines in Skyndidalur

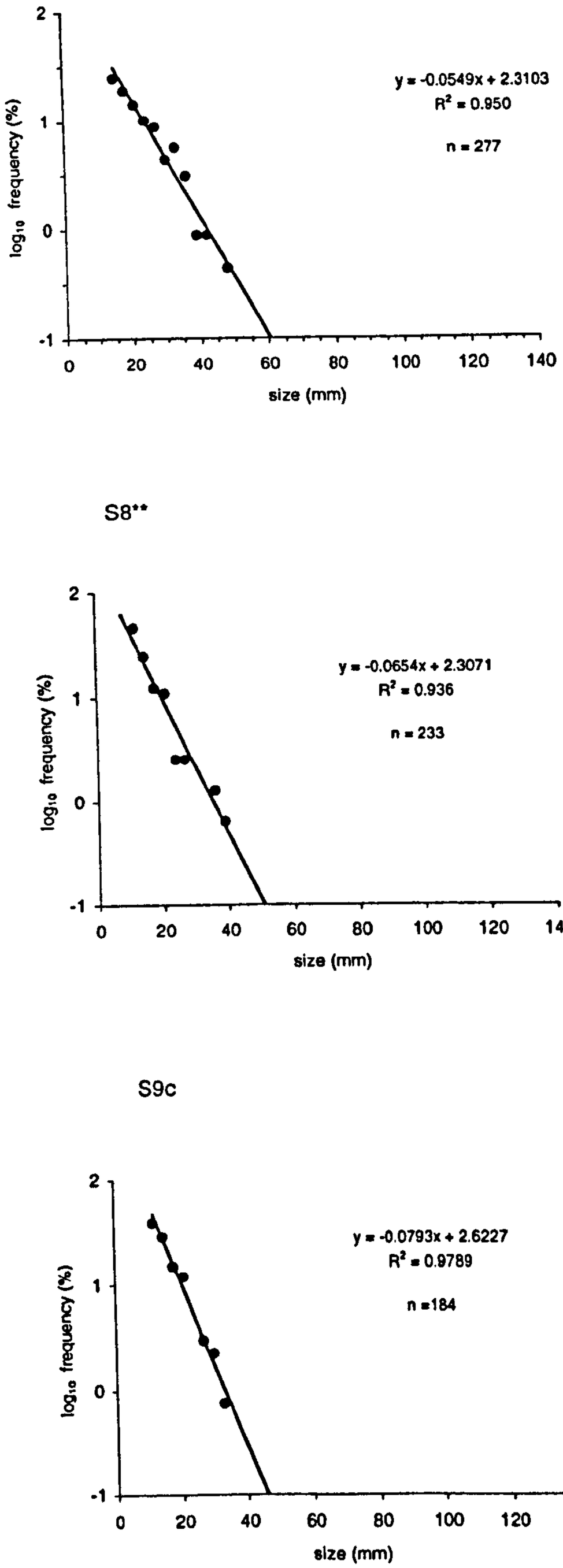
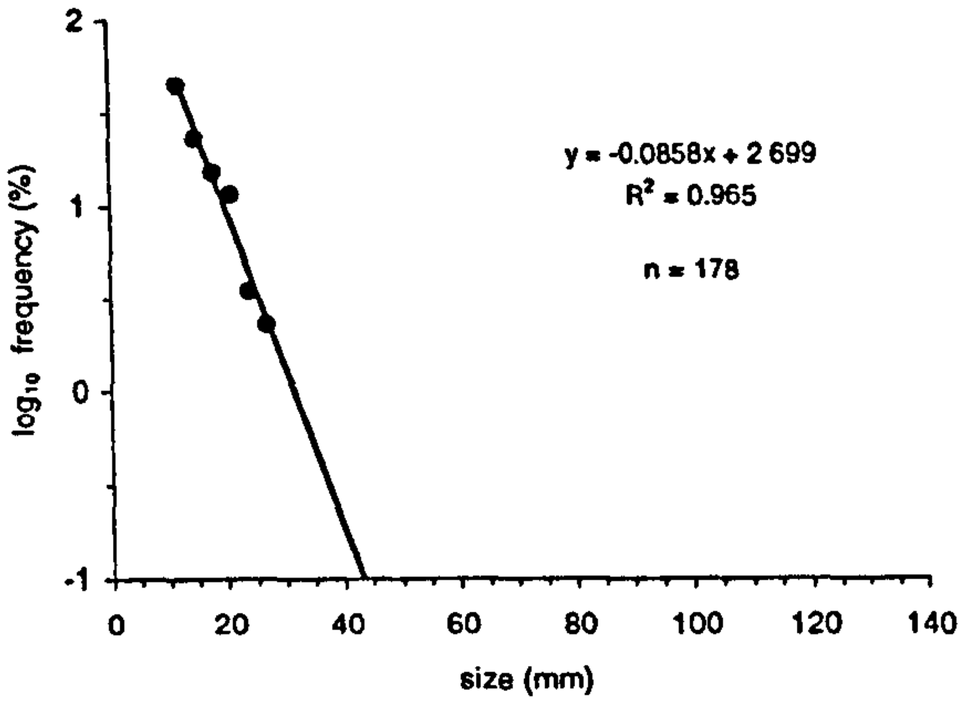
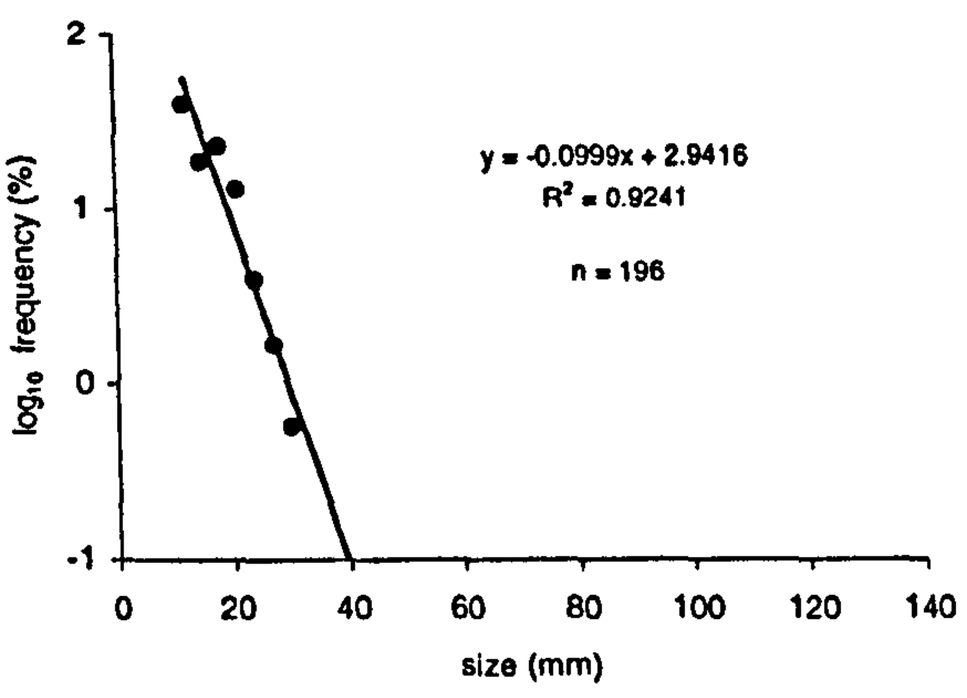
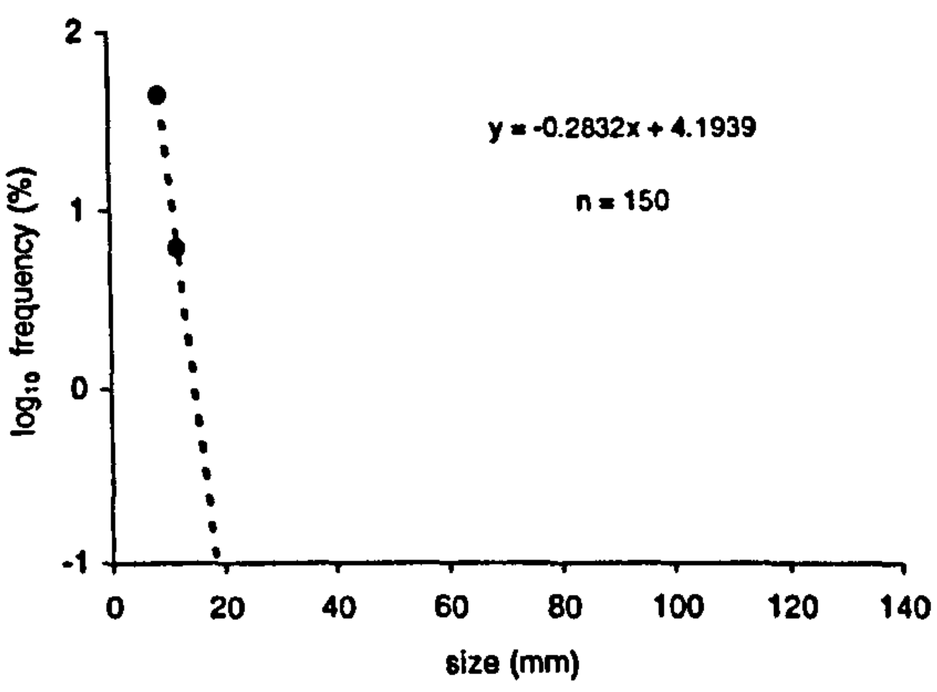


Figure 8.12 Lichens size-frequncy graphs from moraines in Skyndidalur







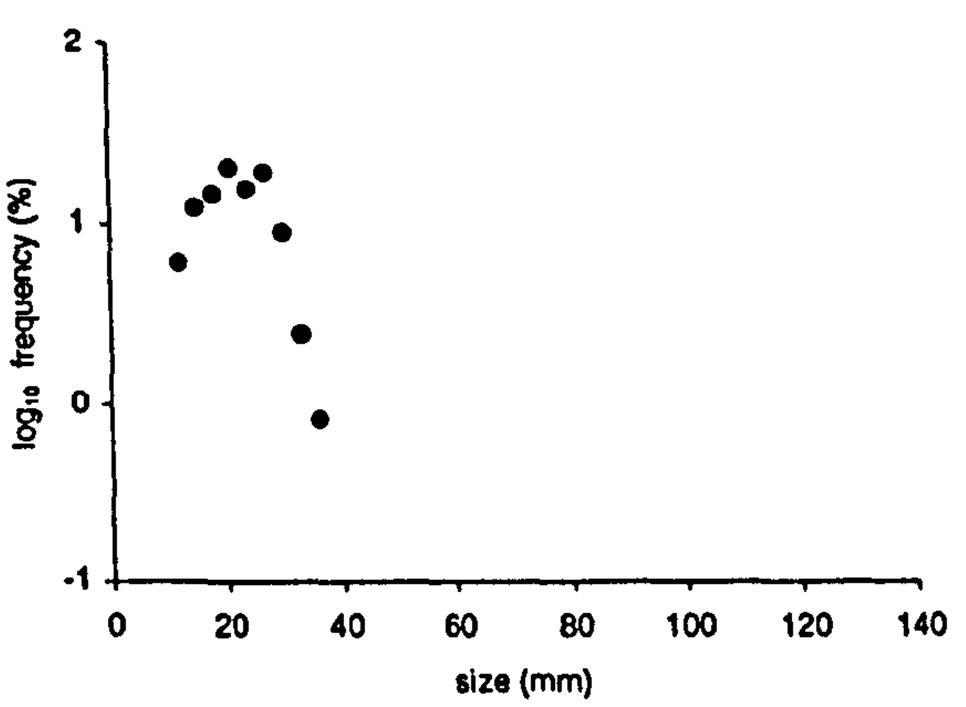


Figure 8.12 Lichens size-frequncy graphs from moraines in Skyndidalur

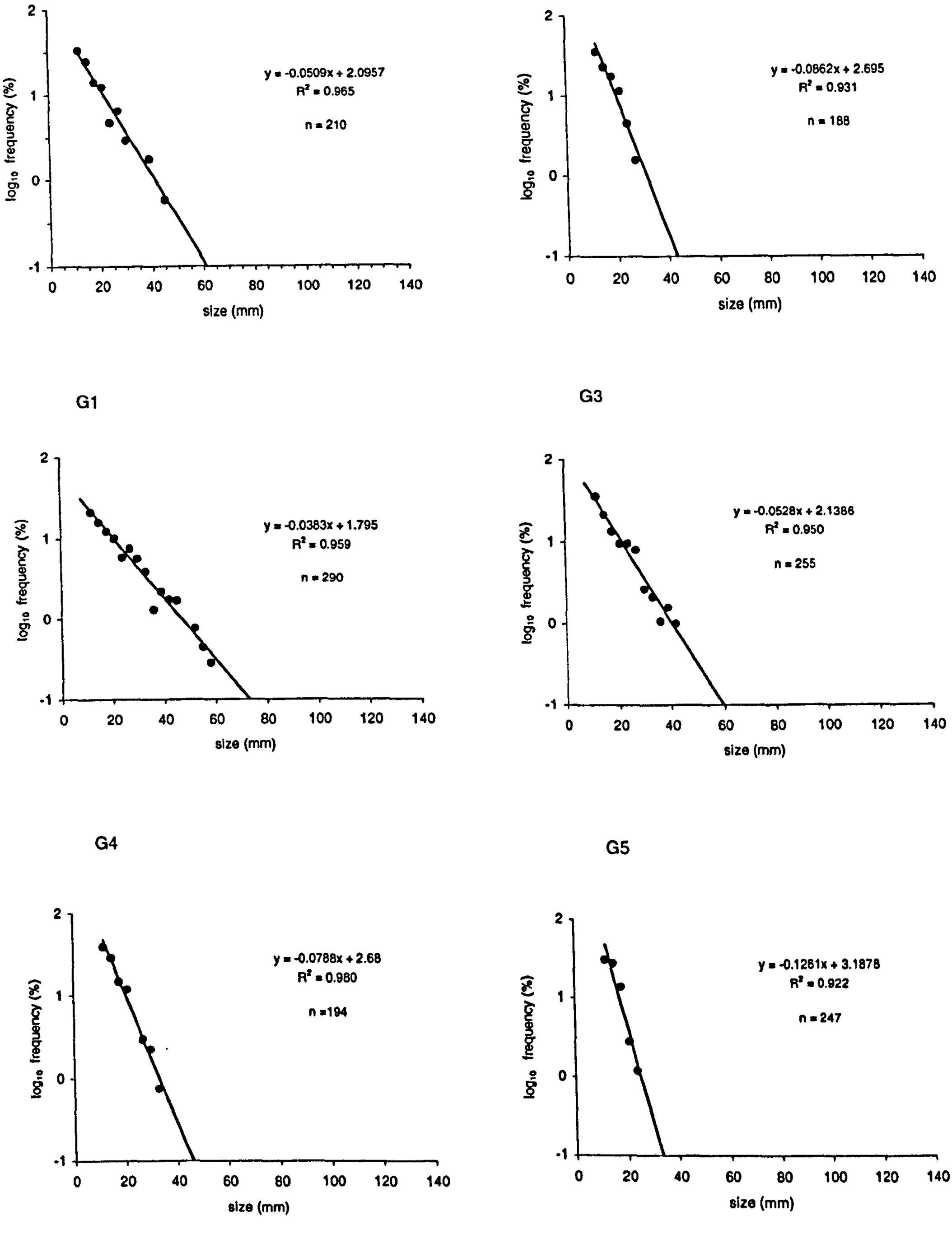


Figure 8.13 Lichen size-frequency graphs from moraines in Fossdalur

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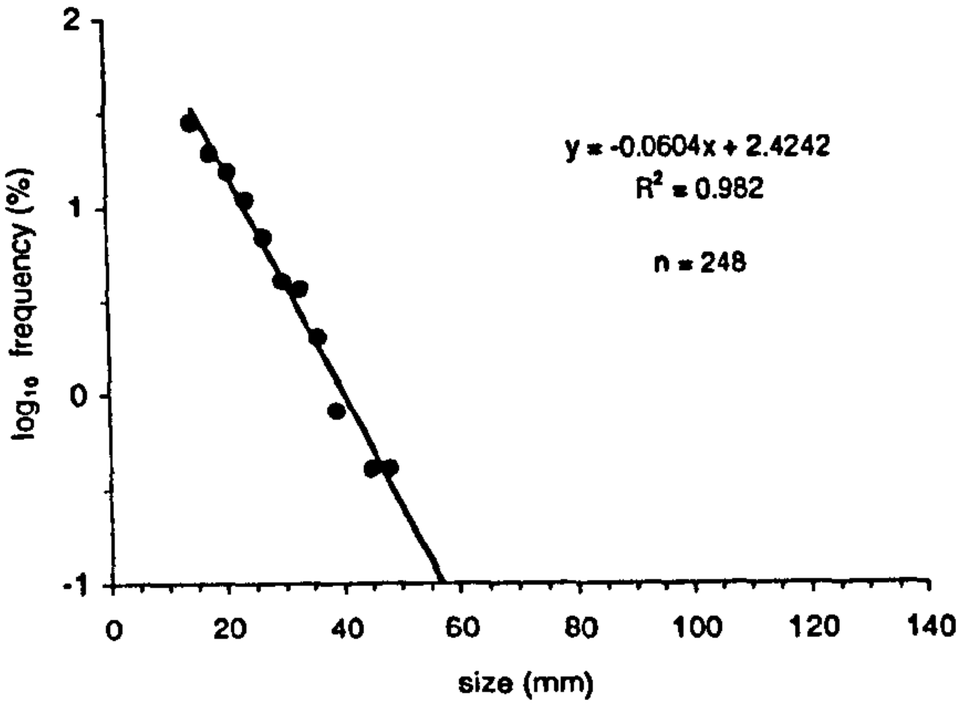

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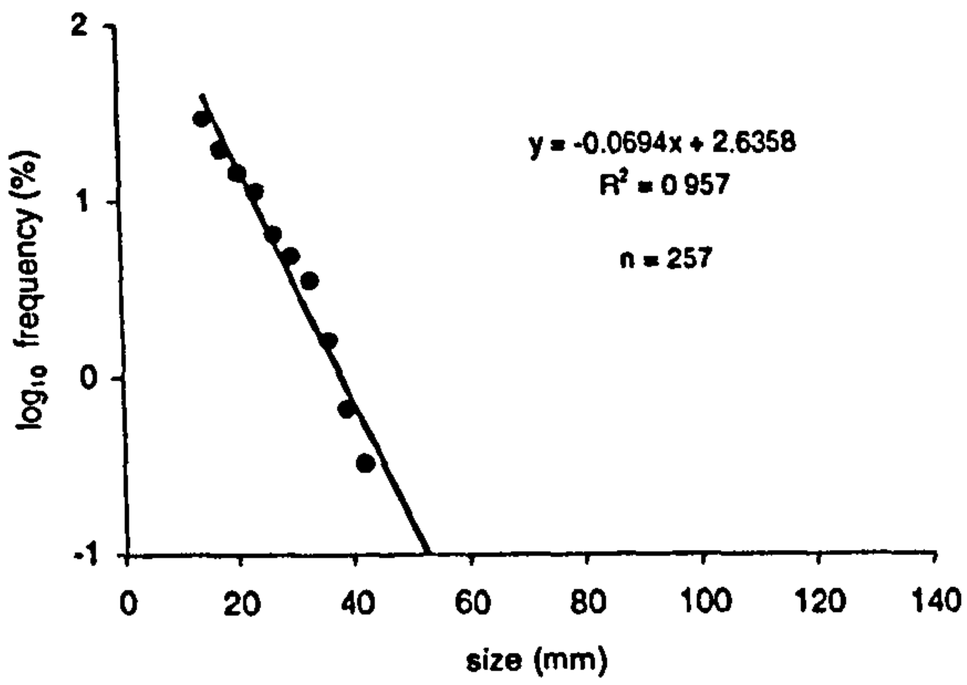

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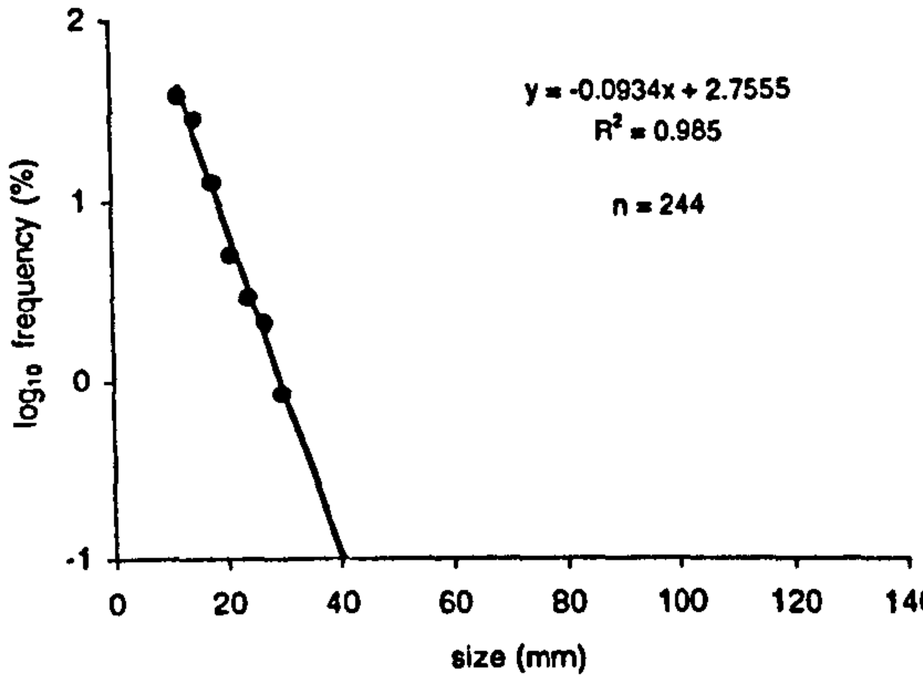

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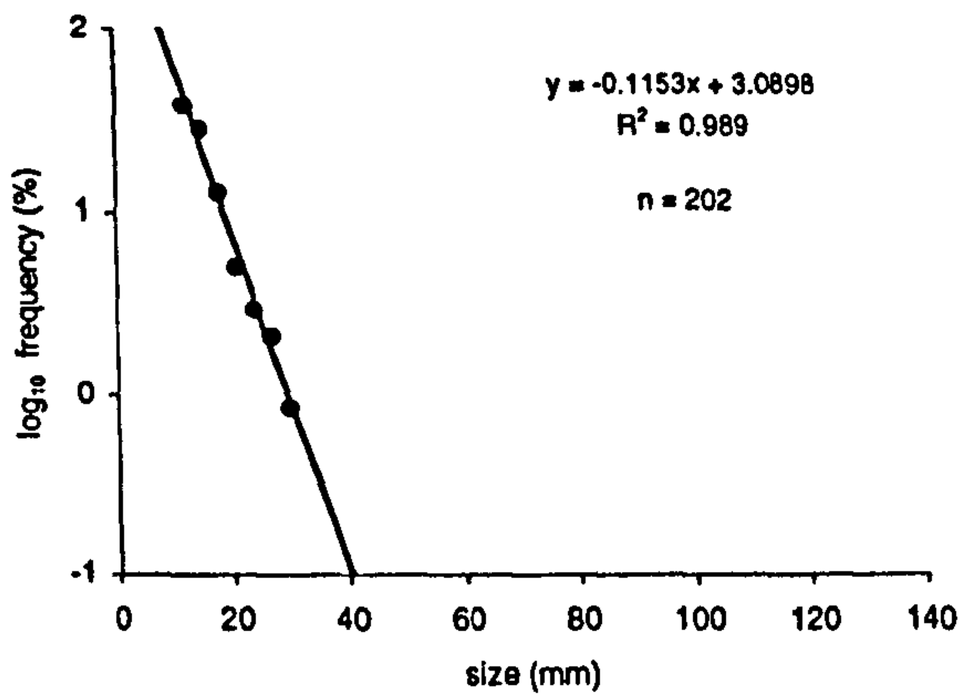

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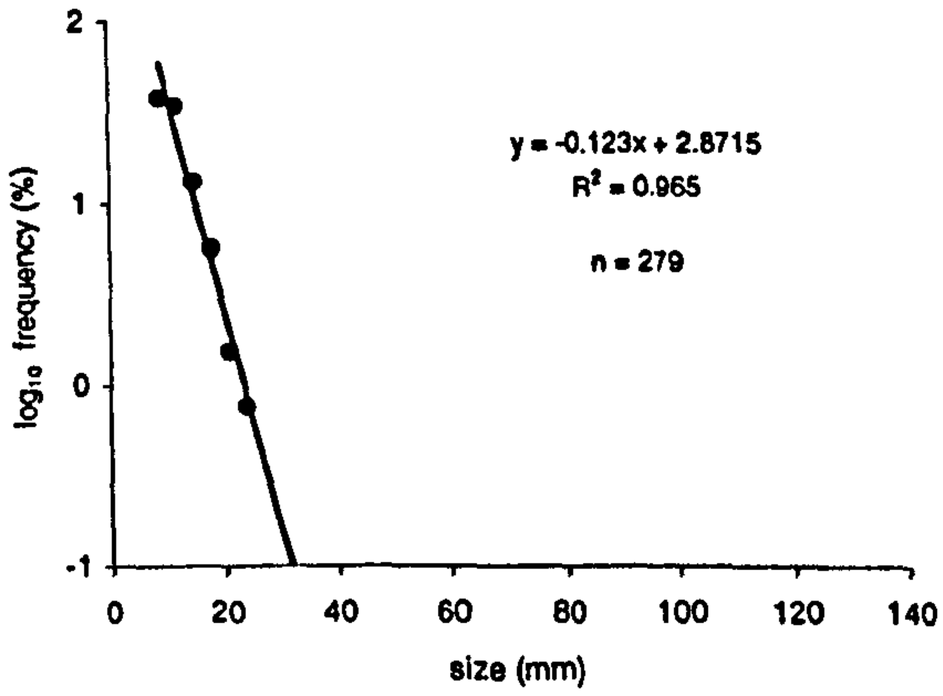

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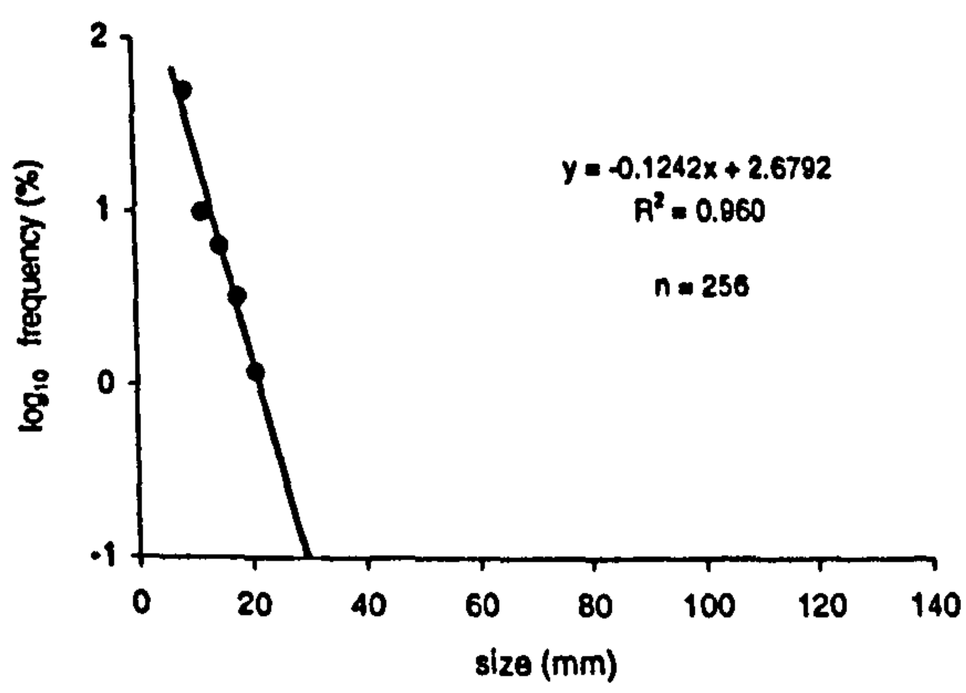


Figure 8.14 Lichen size-frequency graphs from minor moraines in Skyndidalur

Table 8.4 Lichenometric data from moraines in Hoffellsdalur

Moraine	Gradient of size-frequency distribution (-ve)	Date of surface exposure (based on age-gradient curve)	Largest lichen diameter (mm)	Date of surface exposure (based on age-size curve)
H4	0.0253	1796	82	1790
H5	0.0272	1816	66	1856
H5b	0.0317	1835	54	1891
H6	0.0375	1862	60	1875
H6	0.0370	1857	63	1866
H6b	0.0385	1864	49	1903
H7	0.0408	1870	51	1899
H7b	0.0461	1880	54	1890
H7b	0.0453	1882	55	1888
H8	0.0540	1898	42	1917
H8b	0.0500	1892	50	1901
H8b	0.0549	1899	46	1912
H8c	0.0642	1912	39	1923
H9	0.0814	1928	30	1938
H10	0.0829	1930	35	1930
H11	0.0776	1926	36	1928
H12	0.1171	1948	26	1943
H13	0.1527	>1960	21	1948

Table 8.5 (below) Lichenometric data from moraines in Skyndidalur

Moraine	Gradient of size-frequency distribution (-ve)	Date of surface exposure (based on age-gradient curve)	Largest lichen diameter (mm)	Date of surface exposure (based on age-size curve)
S4	0.0315	1835	72	1832
S4*	0.0263	1805	68	1848
S4b	0.0286	1817	74	1825
S5	0.0373	1857	60	1875
S5b	0.0358	1853	65	1859
S5e	0.0370	1854	60	1875
SS	0.0363	1858	53	1893
SN	0.0362	1854	62	1869
S6	0.0398	1866	56	1885
S6b*	0.0453	1880	50	1901
S6b**	0.0407	1870	56	1885
S6b^{un}	0.0700	1919	38	1922
S7	0.0549	1899	45	1912
S8	0.0730	1923	38	1924
S8**	0.0654	1915	39	1922
S8b	no data	no data	no data	no data
S9	0.0833	1931	34	1931
S9b	no data	no data	no data	no data
S9c	0.0793	1927	33	1932
S9d	0.0834	1931	33	1932
S9e	0.0858	1932	32	1934
S9f	0.0999	1940	30	1937
S10	0.2832	>1960	10	1965

Table 8.6 Lichenometric data from moraines in Fossdalur

Moraine	Gradient of size-frequency distribution (-ve)	Date of surface exposure (based on age-gradient curve)	Largest lichen diameter (mm)	Date of surface exposure (based on age-size curve)
Moraines relating to Lambatungnajökull				
L1	no data	no data	no data	no data
L2	0.0509	1892	44	1912
L3	0.0862	1932	36	1928
L4	na	na	na	na
L5	na	na	na	na
L6	no data	no data	no data	no data
L7	no data	no data	no data	no data
Moraines relating to glacier on Grasgiljatindur				
G1	0.0383	1864	59	1878
G2	no data	no data	no data	no data
G3	0.0528	1898	43	1910
G4	0.0788	1926	33	1932
G5	0.1261	1951	24	1945

Table 8.7 Lichenometric data from minor moraines in Skyndidalur

Moraine	Gradient of size-frequency distribution (-ve)	Date of surface exposure (based on age-gradient curve)	Largest lichen diameter (mm)	Date of surface exposure (based on age-size curve)
Minor *arcuate* moraines between S7 and S8				
8	0.0604	1907	44	1913
17	0.0694	1917	40	1920
Minor *sawtooth* moraines between S9 and S10				
a	0.0934	1936	31	1934
d	0.1153	1947	30	1937
m	0.1230	1949	24	1945
p	0.1242	1949	21	1948

Notes

no data – the moraine was not visited
na – the moraine surface was unstable therefore lichens were absent
* – lichens sampled from the crest of the moraine
** – lichens sampled from the ice-distal flank of the moraine
un – the ice-proximal flank of moraine S6b is oversteepened and unstable, subsequently the lichenometric data do not accurately represent the timing of surface exposure.

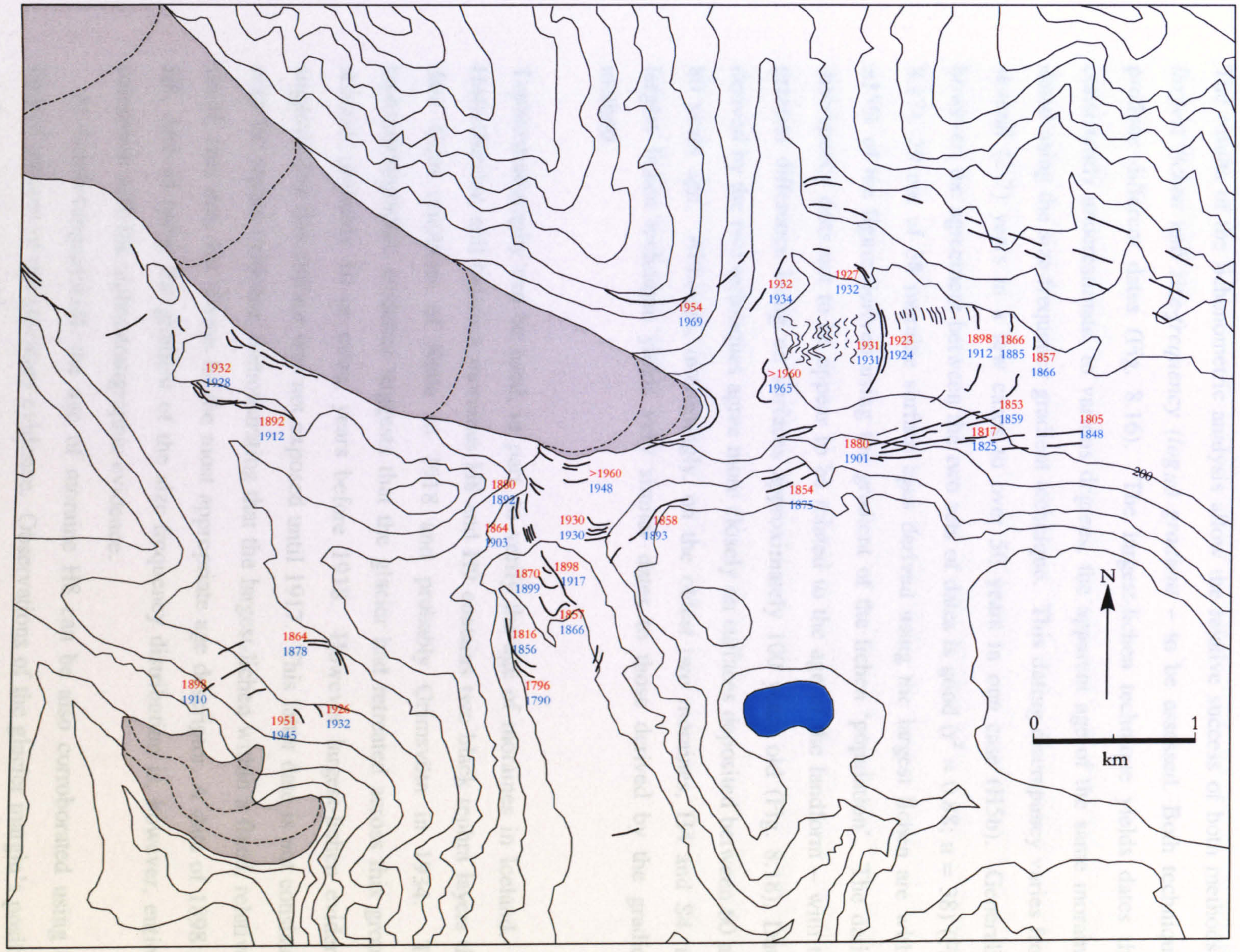


Figure 8.15 Surface exposure ages (AD) of moraines in the study area derived using two different lichenometric techniques. Those dates derived using the *largest lichen* (blue) consistently underestimate those derived using the *gradient of the lichen size-frequency (log_{10}) distribution* (red).

Use of the *largest lichen* and *gradient* techniques in moraine dating studies – a critical assessment

The results of the lichenometric analysis allow the relative success of both methods – *largest lichen* and *size-frequency (log_{10}) gradient* – to be assessed. Both techniques produce different dates (Fig. 8.16). The largest-lichen technique yields dates that consistently underestimate, to various degrees, the apparent age of the same moraines dated using the size-frequency gradient technique. This dating discrepancy varies from several (2-7) years in a few cases to over 50 years in one case (H5b). Generally, however, the agreement between the two sets of dates is good ($r^2 = 0.88$; $n = 38$) (Fig. 8.17). 29 out of 38 moraine surface ages derived using the largest lichen are within ±15% of the figure derived using the gradient of the lichen 'population'. The dating discrepancy does not to be appear to be related to the age of the landform – with the greatest difference being on surfaces approximately 100 years old (Fig. 8.18). Dates derived by the two techniques agree more closely on surfaces deposited between 60 and 80 years ago. Although, interestingly, on the oldest two moraines, H4 and S4, the largest lichen technique yields very similar dates to those derived by the gradient method.

Tephrastratigraphy can be used, in part, to verify the age of moraines in Iceland. In Hoffellsdalur, soil between moraines H8 and H9 contains two black tephra layers that date from eruptions of Katla in 1918 and probably Grimsvötn in 1934. The tephrastratigraphic evidence suggests that the glacier had retreated across this ground several, probably 10 or more, years before 1918. However largest-lichen evidence suggests that this surface was not exposed until 1917. This latter date is not consistent with the tephra evidence, demonstrating that the largest-lichen within a fixed, relatively small, area may not always be the most appropriate age descriptor. A date of 1898 for H8, derived using the gradient of the size-frequency distribution is, however, entirely consistent with the tephrastratigraphic evidence.

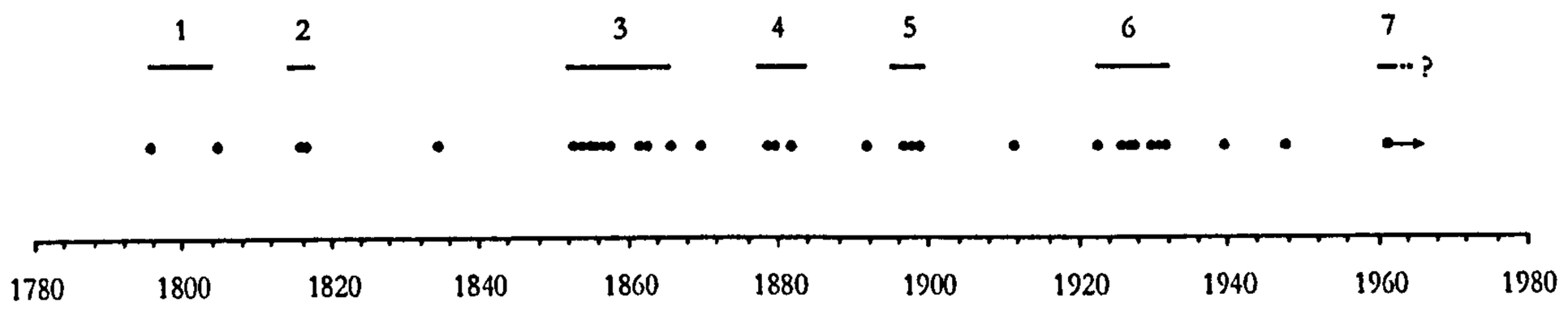
At Lambatungnajökull, the age of moraine H8 can be also corroborated using the limited amount of documentary evidence. Observations of the glacier margin's position at the end of the 19th century (1890s) correspond closely with the position of the

moraines at the head of Hoffellsdalur (Thorodssen, 1913-1915; Thorarinsson, 1943). The largest of these moraines (H8) was exposed in the late 1890s according to the gradient of the lichen size-frequency distribution. This is further support of the validity of the dating technique.

Overall, the gradient of the size-frequency distribution tends to represent the date of older surfaces more accurately than the largest lichen within the distribution. However, on younger surfaces deposited in the last ~60 years these results suggest that the largest lichen is a more precise age indicator due to the exponential form of the age-gradient dating curve. Crucially, the chronology, and ultimately the climatic interpretation, of glacier fluctuations in the study area will be different depending on the technique used.

From this point in the thesis, lichenometric age-estimates for the older moraines, formed over ~80 years ago, are based on the *gradient* technique; whilst for those formed less than ~60 years ago, the *largest-lichen* technique is preferred. For consistency, those moraines formed between 60 and 80 years ago, where both techniques compare well, are also dated based on the *gradient* technique.

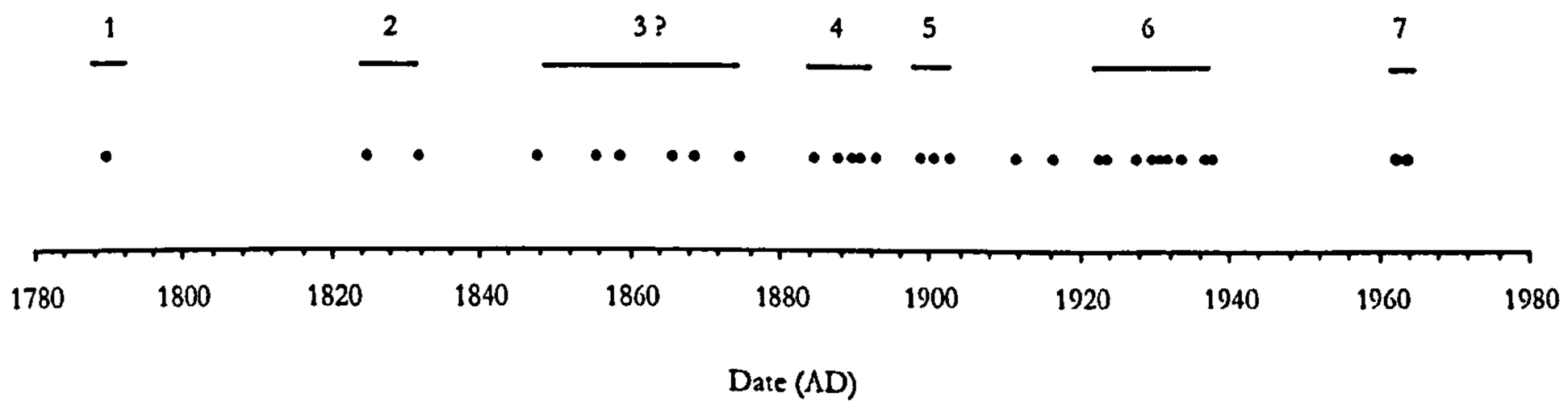
Figure 8.16 A comparison of dates derived using the *gradient* and *largest-lichen* techniques.



(above) The date of surface exposure of the moraines relating to Lambatungnajökull – derived using the *gradient* technique. Note that the dates tend to cluster at certain periods over the last 200 years.

Advances of the glacier margin, and consequent moraine formation, would have occurred shortly before the following times:

1 – 1796 - 1805
2 – 1816
3 – 1855 - 1865 (possibly two episodes)
4 – 1880
5 – 1895 - 1900
6 – 1923 - 1932 (probably two episodes : 1923 and 1928-32)
7 – 1965 ?



(above) The date of surface exposure of the moraines relating to Lambatungnajökull – derived using the *largest-lichen* technique. Note that the dates in the early half of the record (pre-1880) are spread across several decades.

Advances of the glacier margin, and consequent moraine formation, would have occurred shortly before the following times, according to this technique:

1 – 1790
2 – 1825 - 1832
3 – 1848 - 1875 (possibly two or three episodes)
4 – 1885 - 1892
5 – 1898 - 1903
6 – 1923 - 1938 (probably two episodes : 1923-24 and 1930-33)
7 – 1965

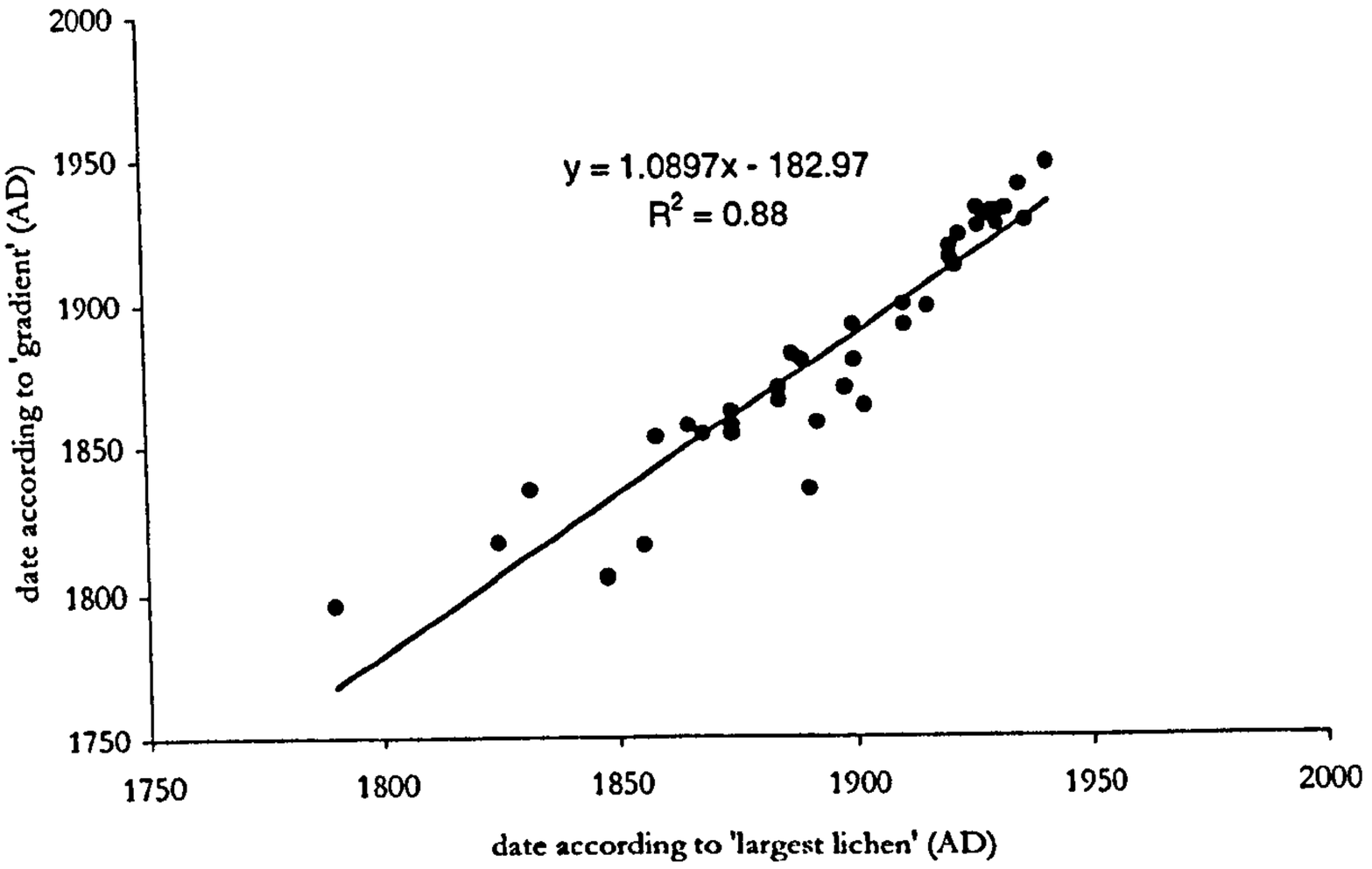


Figure 8.17 The relationship between surface age derived using the largest lichen in a fixed area and surface age derived using the gradient of the size-frequency (log_{10}) distribution. Data are from 36 moraines relating to the former extent of Lambatungnajökull.

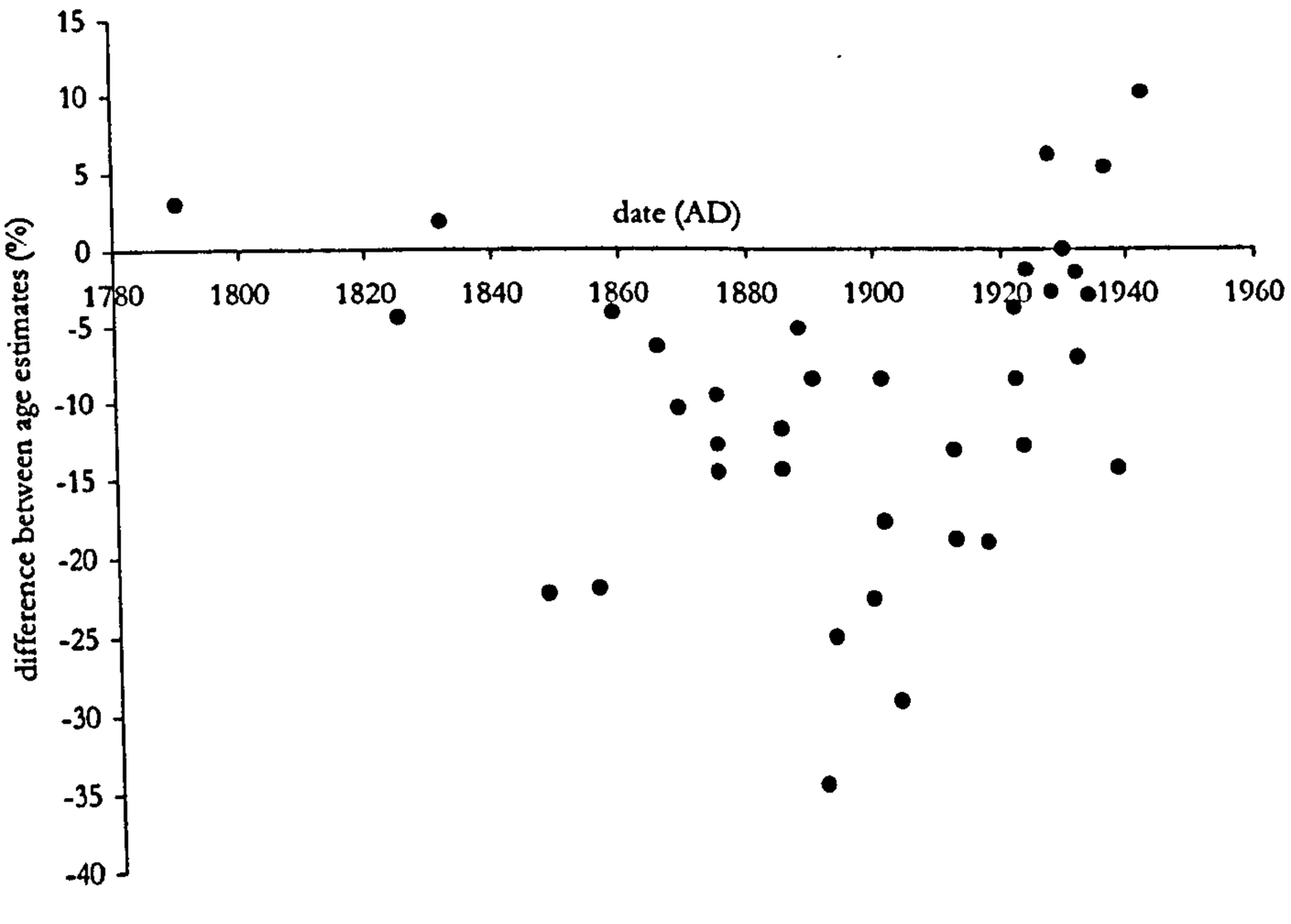


Figure 8.18 The difference between surface ages derived using two lichenometric techniques – *largest-lichen* and *gradient of the size-frequency (log_{10}) distribution.* The differences are expressed as percentages of the *gradient* estimate. Generally, the age according to the *largest lichen* underestimates the age according to the *gradient* by c. 5-20%. Note the high scatter of values c. AD 1900.

Summary of Chapter 8 – Lichenometrical studies

Lichenometric dating curves are an empirical, indirect way of relating lichen size to lichen age. Based on this assumption, dating curves offer an efficient way of studying the growth rates of slow-growing lichens over periods of many decades. *Age-size* curves derived using 'largest-lichens' growing on surfaces of known age approximate to the optimum growth history of lichens in a given area, and may be used to date features formed within the last few centuries. A more rigorous method of lichenometric dating uses a large number, or 'population', of lichens. The gradient of the lichen population size-frequency (log_{10}) distribution is related the age of the lichen population – with older populations possessing a shallower gradient. Calibrated *age-gradient* curves can be used to date lichen-bearing surfaces exposed within the last ~300 years in southern Iceland, although the dating precision decreases with substrate age. Use of both techniques in partnership will ultimately increase dating accuracy and strengthen the existing lichenometric procedure.

Moraines relating to the former extent of Lambatungnajökull have been dated using both the *largest lichen* and *size-frequency gradient* techniques. Use of the largest lichen tends to under-estimate the ages of moraines relative to the gradient approach, particularly those deposited c. 100 years ago. However, 75% of the dates derived using the largest-lichen approach are within ±15% of the dates derived using the gradient of the size frequency distribution. Tephrastratigraphic and documentary evidence support the use of the gradient of the lichen 'population' distribution on surfaces more than ~60 years old. Younger surfaces deposited in the last 60 years or so are more accurately dated using the largest thallus within a large lichen population.

Lichenometry provides a minimum date for the timing of surface exposure. Consequently, the glacier advances or stillstands of Lambatungnajökull that deposited the larger moraines occurred a few years before the following times (AD):

Table 8.8 Summary of moraine ages

Hoffellsdalur	*Skyndidalur*
1796	1805
1816	1817
1854-64	1853-58
1870-80	1866-80
1892-99	1898
---	1923
1928-32	1931-34
1965	1965

Interestingly, the outermost moraine dated with lichenometry indicates that Lambatungnajökull was more extensive during the 1790s than in the late 19th century. This finding, supported by the presence of a previously unreported late-18th-century moraine fragment in front of Fjallsjökull, may require a revision of the *Little Ice Age* glacial chronology in southern Iceland. Further implications surrounding the timing of these, and other, glacier advances will be discussed in Chapter 10.

Chapter 9 : Lichenometric Investigations and Implications

Introduction

The temporal variations in growth rate that may occur over the life history of a lichen have been addressed in Chapter 5. This chapter seeks to examine systematic variations in lichen growth rates as a result of micro- and macro-environmental factors. An understanding of the ecological conditions responsible for different rates of lichen growth would greatly improve the application of lichenometry as a dating tool. Moreover, lichen growth rates could provide an untapped source of information regarding past environmental change.

Whilst in Iceland, several experiments were set up to examine the effects of micro-environment on lichen growth. The results and implications of these are discussed in this chapter. Regional variations in lichen growth rates across Iceland have also been studied. Furthermore, an examination of previous lichenometric research conducted in Northwest Europe and Greenland sheds light on the relationship between climate and lichen growth rates over the last 100 years.

Direct measurements of lichen growth

Presently, there are no published studies of directly-measured lichen growth from Iceland. As part of the lichenometric investigations conducted in 1998 and 1999 it was decided to set up a long-term monitoring project to determine the absolute growth rates of crustose lichen thalli in southeast Iceland. The study is chiefly concerned with the yellow-green *Rhizocarpon* lichens, in particular those species belonging to the *geographicum* group (as classified by Runemark, 1956) (ie. *Rhizocarpon* Section *Rhizocarpon*, as classified by Cernohorsky, (1977)). 41 healthy, non-competing thalli of *Rhizocarpon* Section *Rhizocarpon* (not identified to the species level) were selected for long-term monitoring. The thalli ranged in diameter from 3 mm to 95 mm. Points around the thalli were marked using

permanent, waterproof paint and a fine 0.13 mm Rotring^{TM} pen. Measurements were made (to the nearest 0.1 mm) from each mark to the outermost edge of the hypothallus. At least three measurements were made per thallus. Each lichen was photographed with a 35-mm fixed-lens camera in July 1999 (Fig. 9.1). The chosen sites were carefully described allowing them to be revisited in the future. The measurements are listed in Appendix II. A further 100 or so *Rhizocarpon* lichens were photographed and measured (to the nearest 1 mm) in 1998 and 1999 on boulders, bedrock or gravestones across southeast Iceland. These were not marked but could be revisited in the future to ascertain an approximate growth rate.

Innes (1982) has shown that it is not normally possible to measure significant size changes in slow-growing *Rhizocarpon* thalli during the tenure of a 3-year research programme. It may take 5-10 years for meaningful results to be obtained in certain cases, depending naturally on the original size of the thallus and the environment in which they are growing. However, it is hoped that noticeable size increases will occur over a period of several years and that subsequent investigations will be able to monitor the growth of these particular specimens in the future.

Fortune and a degree of foresight have allowed an indication of lichen growth rates in southeast Iceland to be determined. Lichens measured to the nearest millimetre and photographed by the author in July 1996, whilst on undergraduate fieldwork, were re-measured in July 1999 (Fig. 9.2). Measurements of only 17 yellow-green *Rhizocarpon* thalli show that growth of this crustose lichen is very slow. Due to the precision of the original measurements it is not possible to make any meaningful conclusions regarding the absolute growth rate of the lichens under study. Half of the 17 *Rhizocarpon* thalli grew less than 1.0 mm in diameter, which is the same as the precision of the measurements and therefore not statistically significant. A further two thalli could not be confidently identified. However, three *Rhizocarpon* thalli were at least 2.0 mm larger in 1999 than in 1996. On the basis of this limited investigation, it is clear that longer-term studies are required to directly establish the growth rate of *Rhizocarpon* in southeast Iceland. Optimum radial growth rates of *Rhizocarpon* (4-24 mm in diameter) are probably less than 0.5 mm a^{-1} at this site and in many cases considerably less.

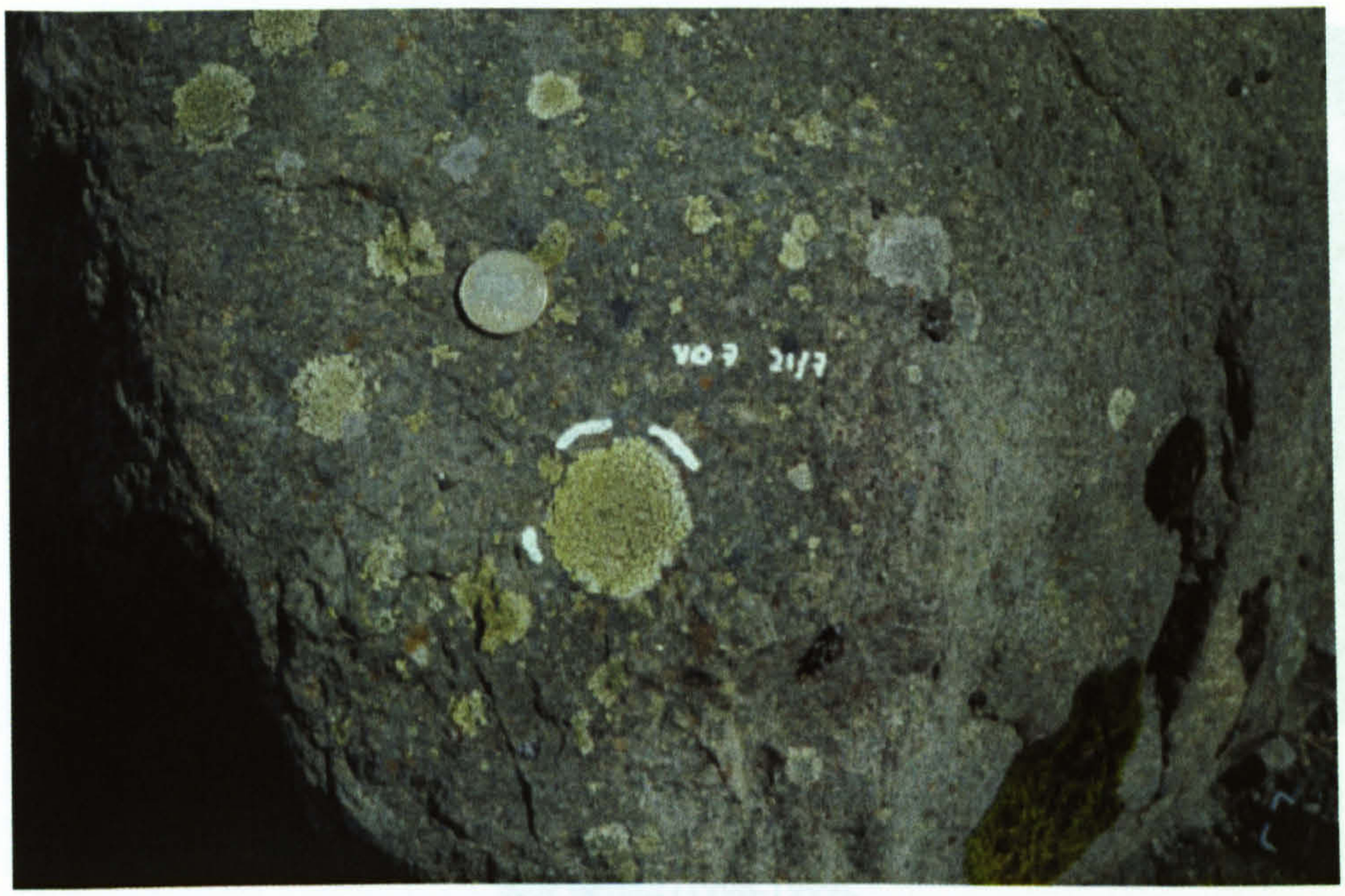
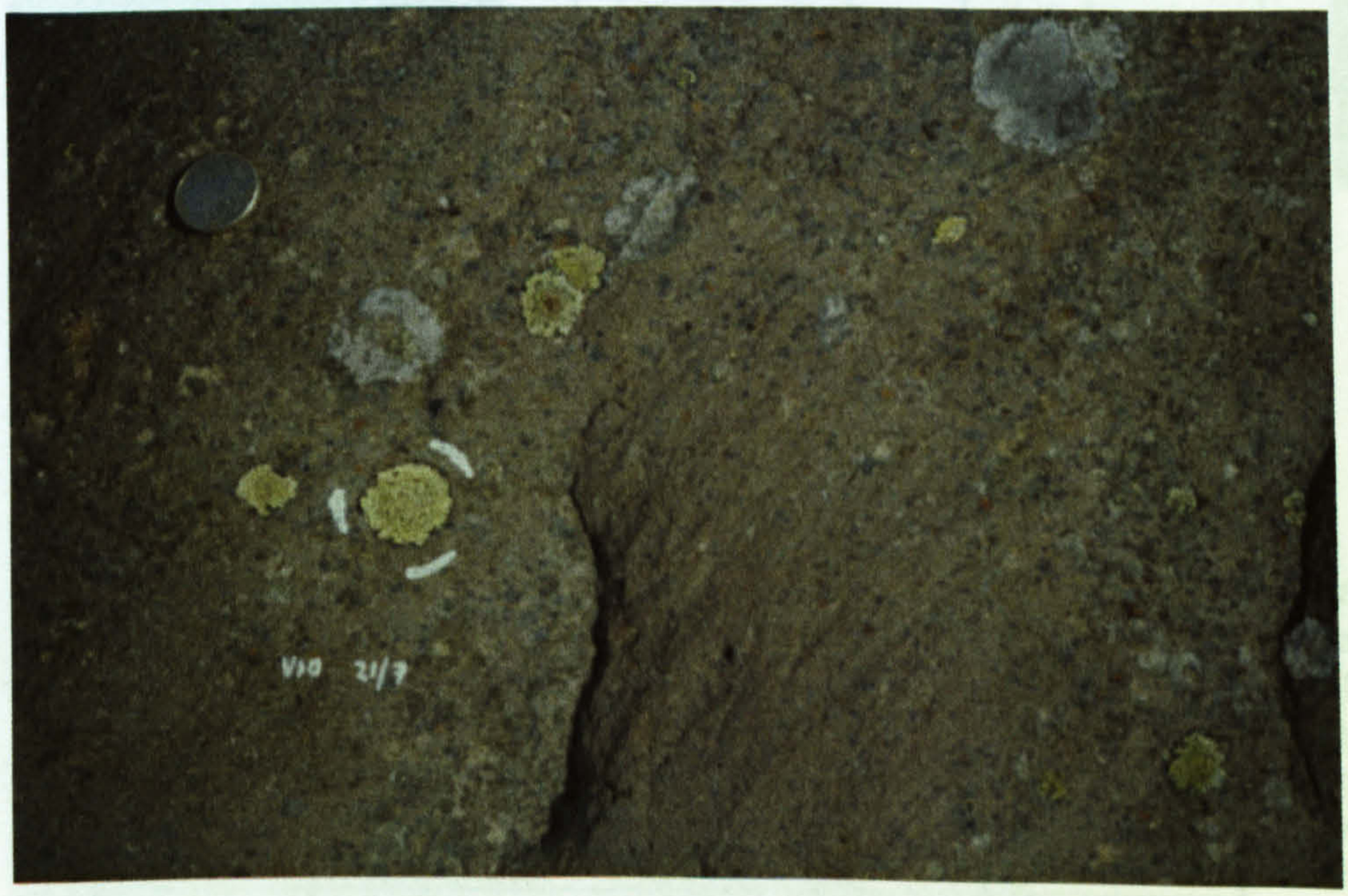


Fig. 9.1. Direct measurements of lichen growth rates on glacially-deposited boulders, Virkisjokull.

Note: - 100 Icelandic Krona coin for scale (=33 mm).

(i) (above) V07 - Approximately circular *Rhizocarpon* section *Rhizocarpon*, 55 mm in diameter. Precise measurements of the margin were made relative to several marks around the thallus periphery (21.7.99). Radial growth should be detectable after a few years.



(ii) (above) V10 - Circular *Rhizocarpon* section *Rhizocarpon* thalli 30 mm in diameter, with wide hypothallus and lobate margin. Measured on 21.7.99. Note also the coalescent *Rhizocarpon* thalli (centre).

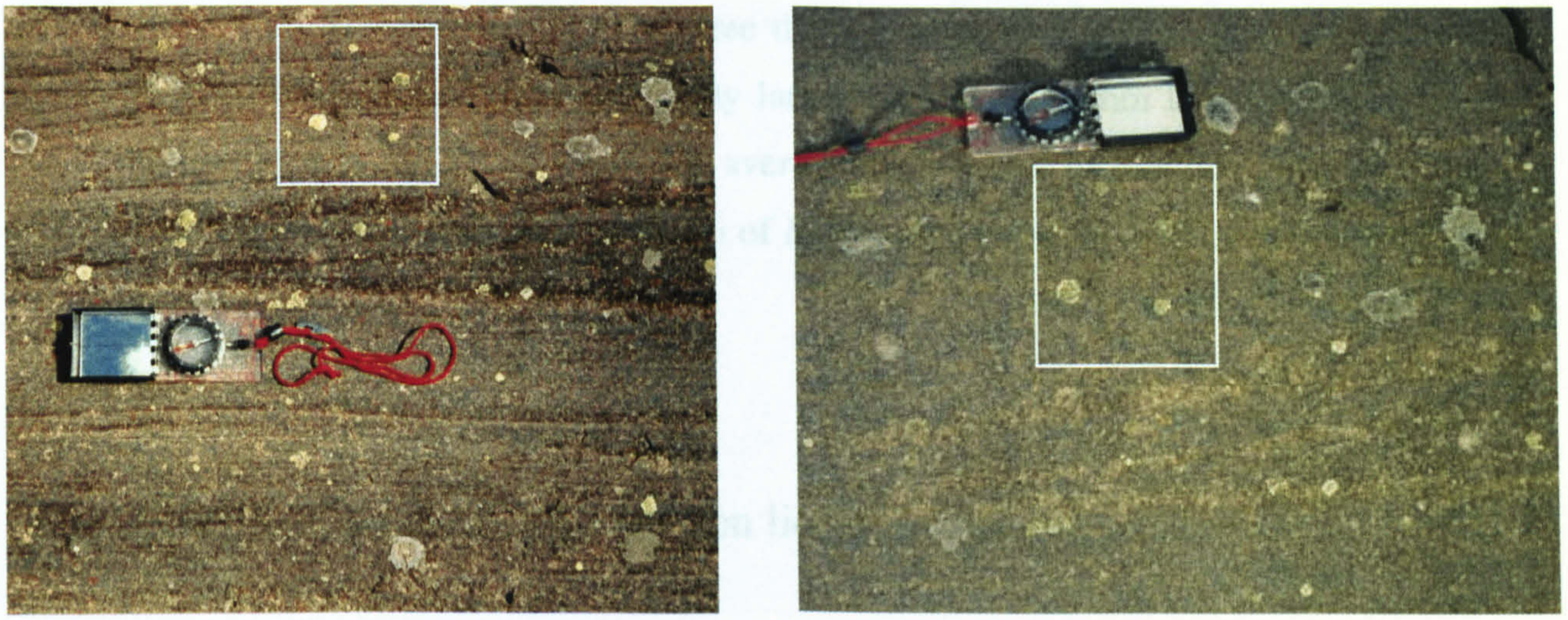


Figure 9.2 Directly-measured growth of crustose lichens at Virkisjökull, southeast Iceland, over a three-year period (1996-1999).

(i) *Rhizocarpon* thalli measured to the nearest millimetre on 29.7.96.

(ii) the same thalli, 3 years later - 4.8.99. Note the compass for scale

Thallus	Lichen family	Maximum diameter (mm)		Mean diameter increase over 3 years (mm)
		1996	1999	
VW-RM-01	section *Rhizocarpon*	21	23.2	2.2
VW-RM-02	section *Rhizocarpon*	6	6.3	0
VW-RM-03	section *Rhizocarpon*	12	12.6	0.6
VW-RM-04	section *Rhizocarpon*	16	17.4	1.4
VW-RM-05	section *Rhizocarpon*	15	17.1	2.1
VW-RM-06	section *Rhizocarpon*	4	5.5	1.5
VW-RM-07	section *Rhizocarpon*	10	na	na
VW-RM-08	section *Rhizocarpon*	21	21.6	*0.6*
VW-RM-09	section *Rhizocarpon*	13	13.0	*0*
VW-RM-10	section *Rhizocarpon*	22	22.2	0.2
VW-RM-11	section *Rhizocarpon*	14	15.0	*1.0*
VW-RM-12	section *Rhizocarpon*	17	18.0	*1.0*
VW-RM-13	section *Rhizocarpon*	10	12.8	*2.8*
VW-RM-14	section *Rhizocarpon*	14	13.7	*0*
VW-RM-15	section *Rhizocarpon*	4	na	*na*
VW-RM-16	section *Rhizocarpon*	14	15.3	1.3
VW-RM-17	section *Rhizocarpon*	17	16.6	0
VW-RM-18	Buellia canescens	37	na	na
VW-RM-19	Buellia canescens	46	46.5	0.5
VW-RM-20	Buellia canescens	22	24.0	2.0
VW-RM-21	Buellia canescens	37	39.0	2.0
VW-RM-22	Buellia canescens	30	34.5	4.5

(left) Measurements of 17 *Rhizocarpon* and 5 *Buellia* thalli on a single boulder in 1996 and again in 1999. The author suggests that subsequent measurements should be carried out at this site in several years' time to ascertain more meaningful growth rates.

The growth rates of five crustose lichen thalli, tentatively identified as *Buellia canescens*, have also been ascertained from the photographs taken in July 1996 and 1999. The maximum diameter of three thalli increased by at least 2.0 mm over this period. One of these was noticeably larger – being 34.5 mm in diameter in 1999 compared with 30 mm in 1996. On average, this species of *Buellia* appears to grow more rapidly than neighbouring thalli of *Rhizocarpon* subgenus *Rhizocarpon*, at least in this particular environment.

Micro-environmental controls on lichen growth

The influence of micro-climate on the growth on saxicolous lichens has been studied by Armstrong (1973, 1975, 1976, 1977). Armstrong monitored the growth of several foliose lichen species (incl. *Parmelia glabratula, Parmelia conspersa* and *Physcia orbicularis*) over consecutive years. He concluded that the aspect distribution of these species was influenced primarily by the micro-environment of the particular rock surfaces. Armstrong (1976) proposed that the seasonal growth patterns were influenced by the amount of moisture available, more specifically the frequency of wetting and the rate of drying experienced by the thallus.

The effect of aspect

Whilst working in Iceland, the author noticed numerous examples of preferential lichen growth on one facet of a rock relative to the other. A survey was conducted in the vicinity of Lambatungnajökull to examine the relationship between lichen density and aspect. The results show that crustose lichens are less common on north-facing rock surfaces (Fig. 9.3). Furthermore, the largest yellow-green *Rhizocarpon* specimens were often found on south-facing rock surfaces. These results indicate that southerly aspects may be better suited for lichen colonisation and optimum growth. It should be mentioned that the role of surface inclination, which was not incorporated in this study, complicates matters somewhat. In view of these findings, an experiment to monitor the long-term growth of two populations of crustose lichens under contrasting environmental conditions was set up in July 1999 (Fig. 9.4).

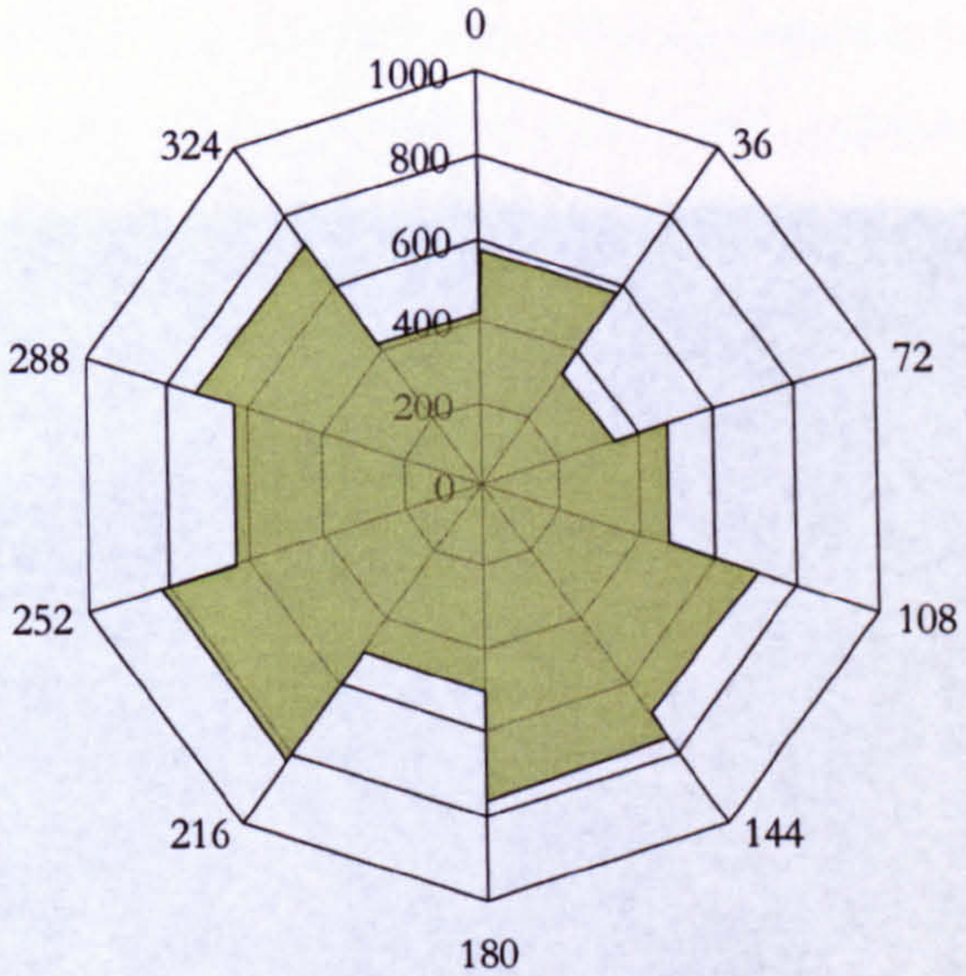
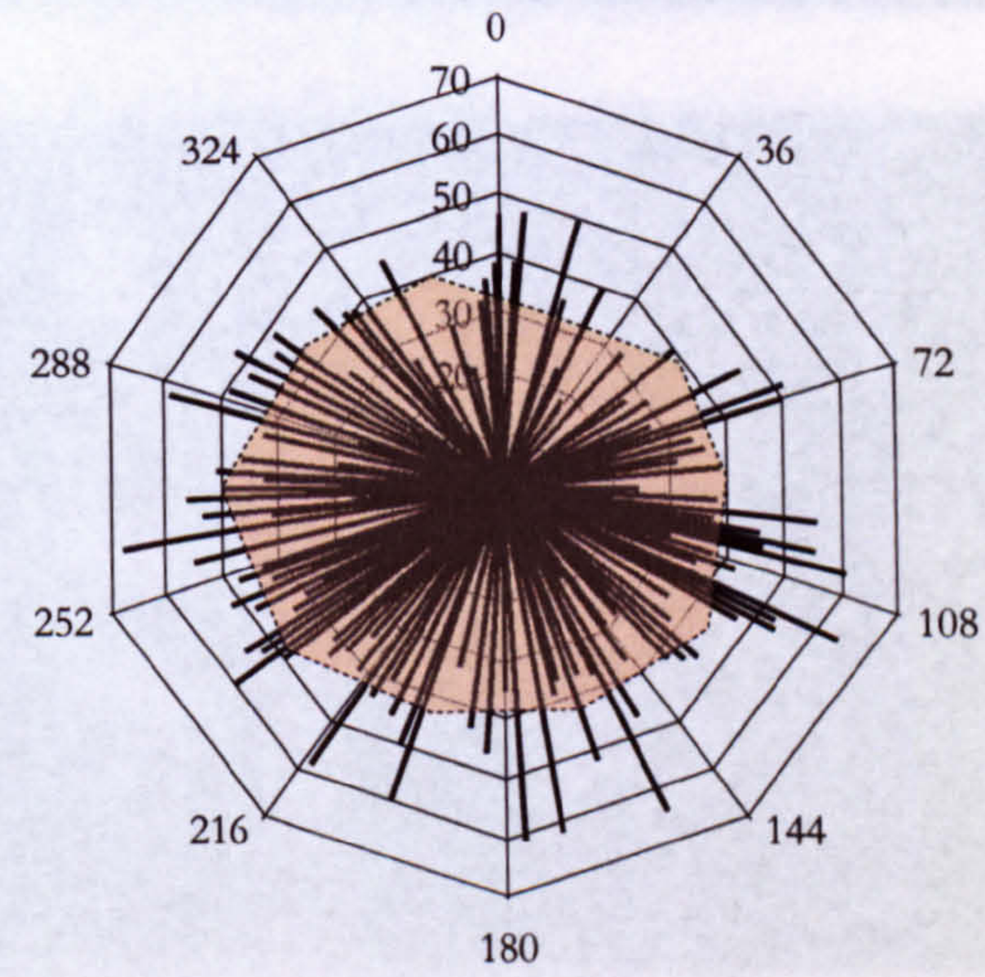


Figure 9.3

(i) The relationship between surface orientation (aspect) and lichen abundance on the foreland of Lambatungnajökull.

The diagram shows the number of yellow-green *Rhizocarpon* lichens (>5 mm) growing on surfaces of different aspect. Over 200 boulders, of various exposure age, were sampled. Aspects have been grouped into eight class intervals.

Rhizocarpon are more abundant on southward-facing rock surfaces in the vicinity of Lambatungnajokull. Fewer thalli were present on N and E aspects.



(ii) The relationship between surface orientation (aspect) and maximum lichen size on the foreland of Lambatungnajökull.

The diagram shows the diameter (mm) of the largest yellow-green *Rhizocarpon* lichen on surfaces of various aspect (0-360°). The orange polygon shows the average 'largest-lichen' diameter in each sector.

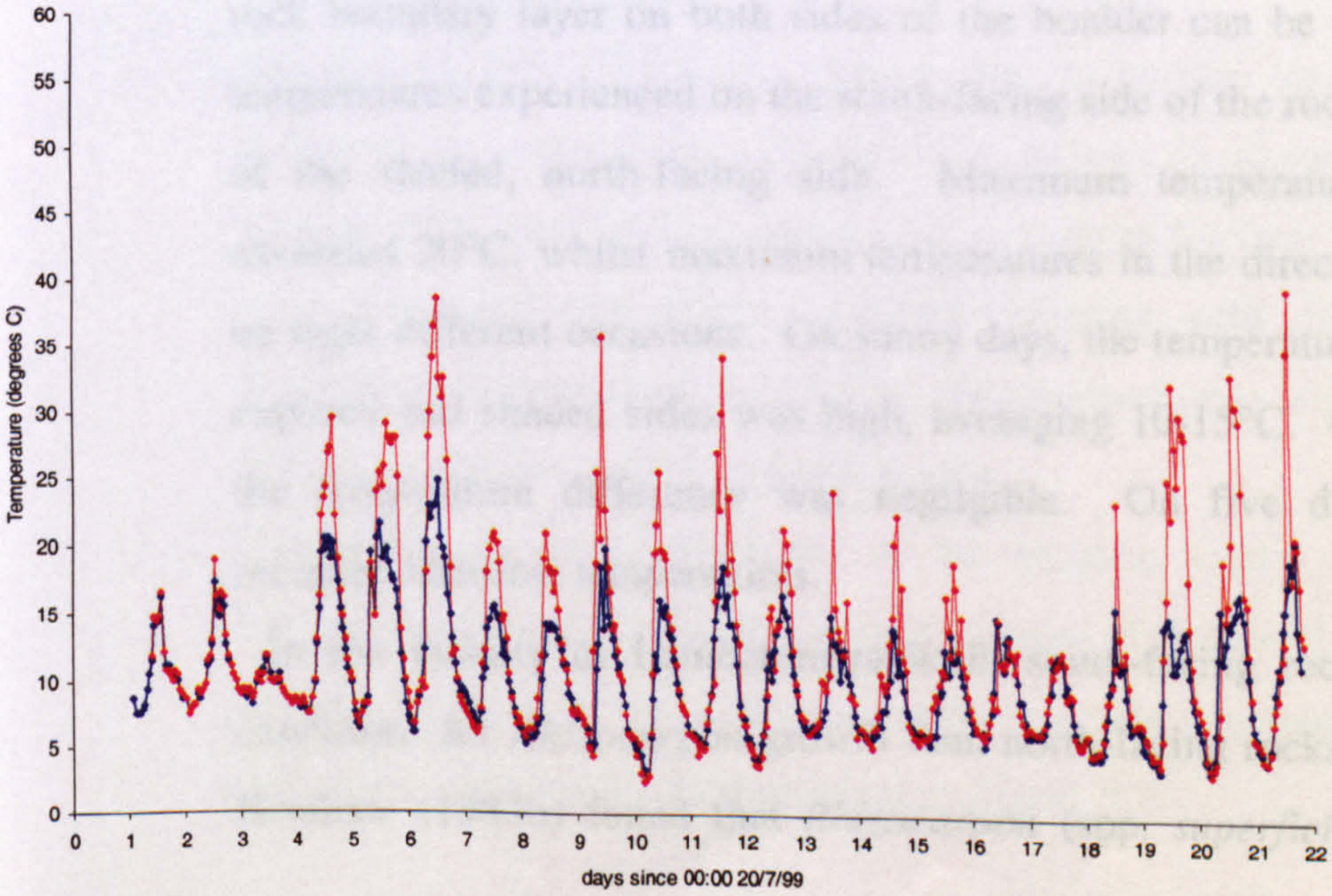
There is a tendency for the largest *Rhizocarpon* thalli to be encountered on south- and west-facing rock surfaces. The average 'largest-lichen' on N and E aspects is generally smaller than the average on S- and W-facing surfaces, irrespective of exposure age.

(above) Lichens growing on a south-facing rock surface, ~1500 m from the snout of Lambatungnajökull. In contrast, the shady, northern aspect has few thalli. Rocks with no measurable aspect (ie. horizontal surfaces), such as the cobble on the left, could not be included in the *lichen growth/ aspect* study.

Figure 9.4 Micro-environmental factors affecting lichen growth – An experiment to monitor the thermal environment on rock surfaces of opposing aspect *and* the consequent growth of saxicolous lichen populations.

(left) The location of the experiment site. Lambatungnajokull is in the background. The crest of the rock strikes almost due E-W.

(left) Unscreened temperature probes were attached at the same height to both sides of the boulder. The south-facing rock surface has approximately 10 times more *Rhizocarpon* thalli growing on it than the shady north-facing side.



(left) Hourly temperature readings from north-facing (blue line) and south-facing (red line) rock surfaces. Measurements were taken between 20/7/99 and 14/8/99.

A glacially-deposited, non-vesicular, basaltic boulder in the E-W trending valley of Skyndidalur was selected. The boulder, partially buried by glaciogenic and fluviatile material, presents two opposite-facing, approximately equal-sized facets. The long axis of the rock strikes 102°, hence the two surfaces face almost due north and south. The north-facing side is permanently shaded and does not receive any direct sunlight. Furthermore, as the rock strikes E-W the role of up- and down-valley winds will also be minimised. The diameter of every yellow-green *Rhizocarpon* lichen (greater than 5 mm) was measured to the nearest millimetre on each side of the rock. No *Rhizocarpon* Section *Alpicola* thalli were identified on either face of the boulder. Five different-sized thalli, including the largest, on the south-facing side were marked, photographed and measured to the nearest 0.1 mm. The same was done on the north-facing side.

144 Section *Rhizocarpon* thalli (>5 mm) were identified on the south-facing side of the rock; whilst only 25 thalli were observed on the opposite face. The largest lichen on the south-facing side had a greatest diameter of 20.0 mm; the largest on the north-facing side measured 16.7 mm

On 29th July 1999 a Tinytalk® datalogger with a thermistor sensor was fixed to each side of the rock using fabric tape and waterproof adhesive (Fig. 9.4). Each datalogger was secured at the same height. The thermistors were not shaded from direct solar radiation. The dataloggers were set up to record the air temperature 5 mm above the rock surface every hour.

After three weeks (20 whole days) the dataloggers were removed from the rock and the data was downloaded. The difference between the temperature of the air-rock boundary layer on both sides of the boulder can be seen in Figure 9.4. The temperatures experienced on the south-facing side of the rock greatly surpassed those of the shaded, north-facing side. Maximum temperatures in the shade rarely exceeded 20°C, whilst maximum temperatures in the direct sunlight exceeded 30°C on eight different occasions. On sunny days, the temperature difference between the exposed and shaded sides was high, averaging 10-15°C. On cloudy, overcast days the temperature difference was negligible. On five days the two dataloggers recorded identical temperatures.

In the vicinity of Lambatungnajökull, south-facing rocks generally offer better conditions for *Rhizocarpon* growth than north-facing rocks (Fig. 9.3). Coxson and Kershaw (1983b) found that *Rhizocarpon* (spp. *superficiale*) has a broad thermal

operating environment with an optimum between 0 and 10°C. Photosynthesis rates also vary with illumination levels (Ahmadjian, 1967). The preference of *Rhizocarpon* for south-facing slopes is therefore probably a function of the longer thermal operating period and higher light intensities throughout the year. Rock surface temperatures in excess of 45°C would present a considerable degree of heat stress. However, MacFarlane and Kershaw (1980) have highlighted how some lichens may adjust their photosynthetic capacity to accommodate changing environmental conditions. *Rhizocarpon,* with a broad photosynthetic response curve, can maximise carbon gains depending on the typical frequency and amplitude of the operating environment. In Iceland, south-facing rocks, presenting the greatest range of temperature and moisture conditions throughout the year, offer the most suitable micro-habitat for epilithic lichen communities.

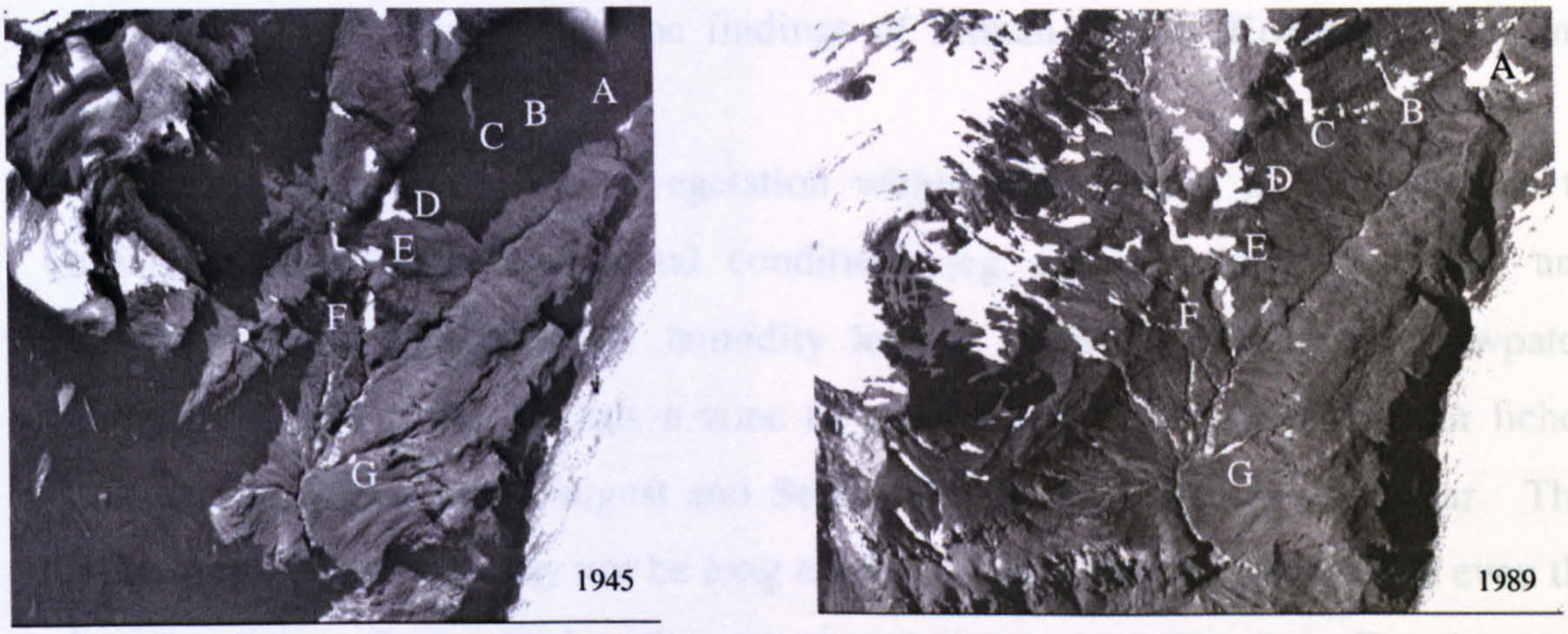
Wirth (1972) and Pentecost (1980) also found *Rhizocarpon* (spp. *geographicum*) preferred south-facing slopes in central Europe and upland Britain respectively. However, this preference for south-facing rocks is not universal. Other site-specific factors such as the degree of moisture available for photosynthesis from rainfall, snow or dew may offset the effects of higher heat or light levels. The rate and nature of thallus development on a rock surface over time will be ultimately determined by a subtle combination of micro-environmental factors including light intensity, temperature, and exposure to wind and moisture, as first suggested by Barkman (1958), and supported by Orwin (1972) and Armstrong (1973).

The effect of snowcover

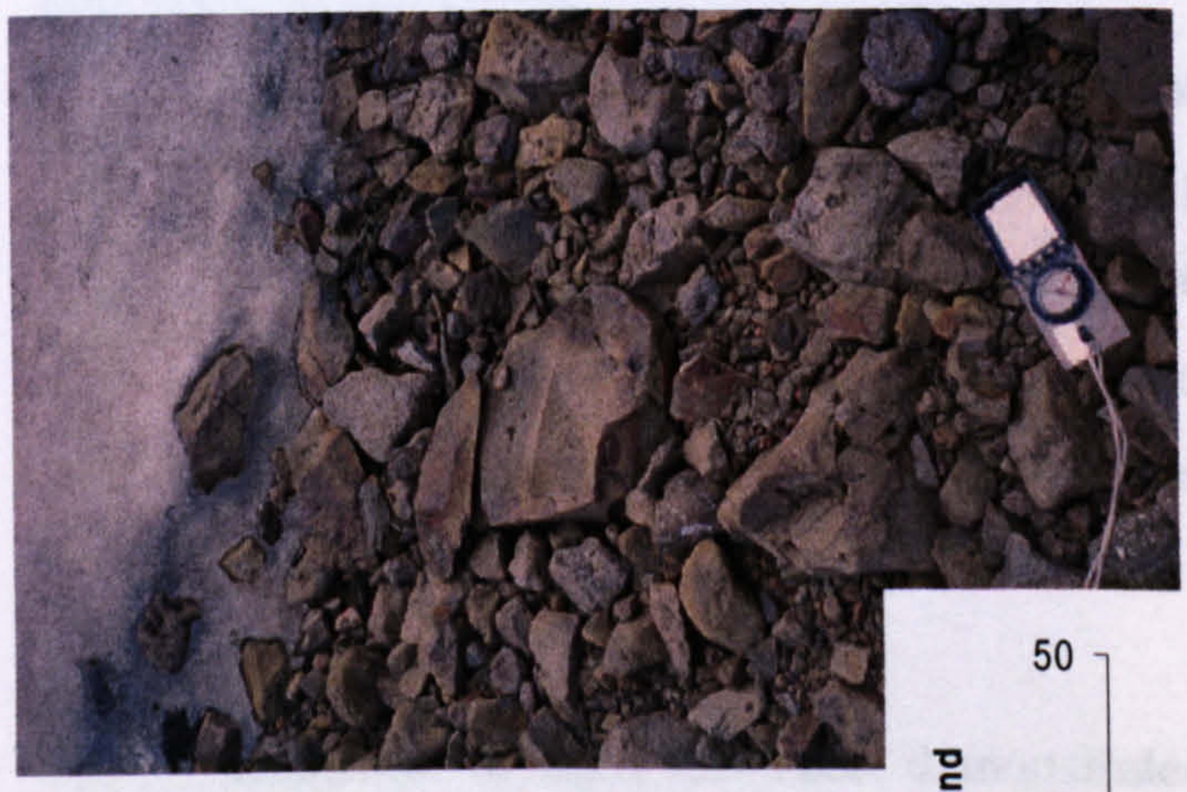
Benedict (1990b, 1991) has shown that in certain situations lichen growth is hampered by snowcover. *Rhizocarpon* lichens in Colorado require a snow-free window of at least 10-12 weeks a year in order to survive. Pitman (1973) noticed in west Greenland that lichen diameters increase with distance away from semi-permanent snowpatches. A similar phenomenon was noticed in upland areas of Iceland during the course of this research

Snowpatches were monitored in the hanging valley of Fossdalur, adjacent to Lambatungnajökull, over the 13-month period of fieldwork. In August 1999, a small lichenometric study was conducted around a large snowpatch in the upper reaches of the valley. The results are presented in Figure 9.5.

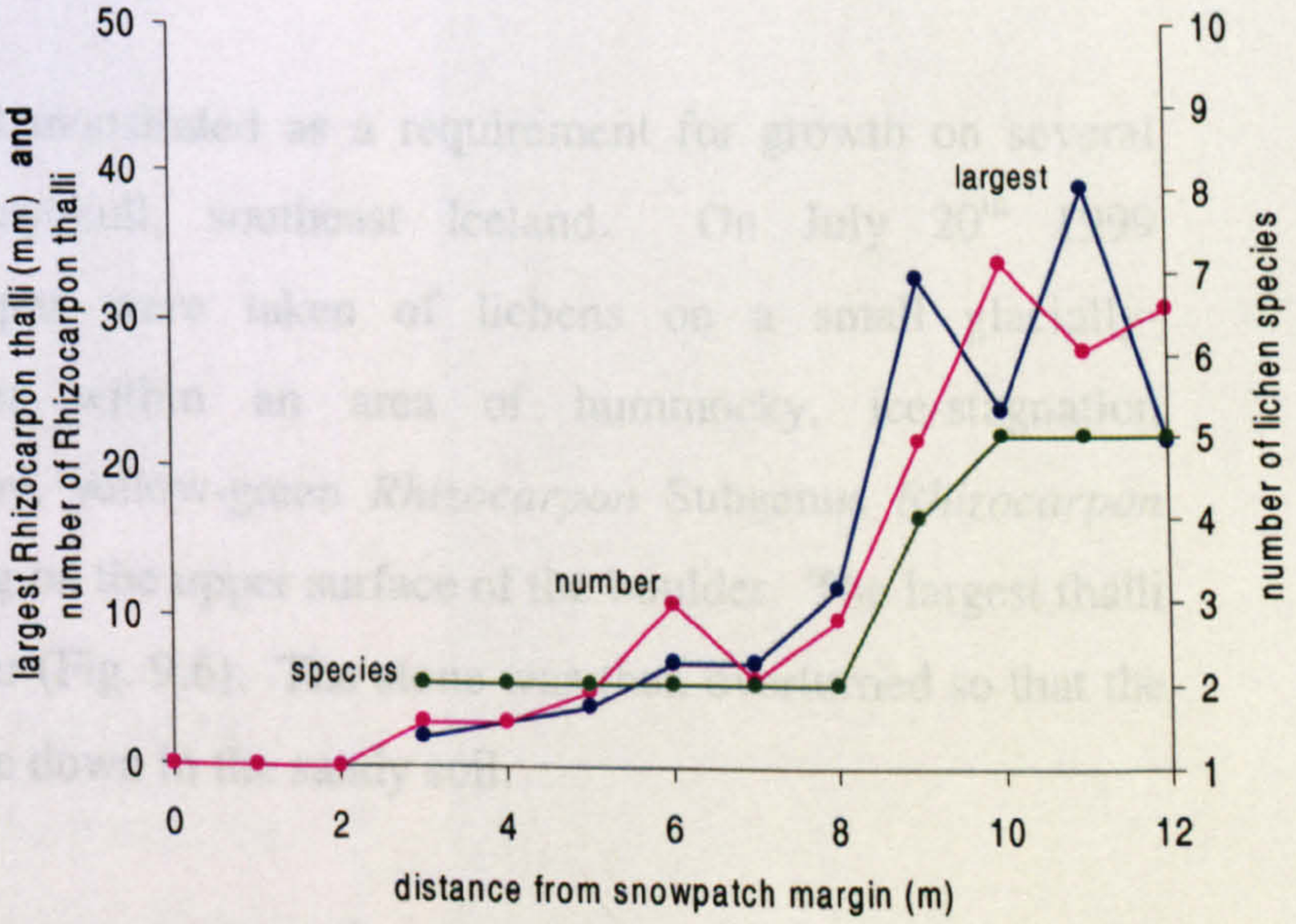
Figure 9.5 Micro-environmental factors affecting lichen growth – the effects of snowcover



(above) Aerial photographs of Fossdalur in the summer of 1945 (left) and 1989 (right) showing the location of various semi-permanent snow patches (A-G). North is to the right.



(left) Lichen-free ground adjacent to snowpatch C (23.8.99).



(right) Results of lichenometric survey performed at the margin of snowpatch C.

Lichens were absent from cobbles and boulders in the snowpatch and also from a concentric zone, ranging from 1-5 m in width, around the snowpatch margin. Transects away from the snowpatch margin detected a general increase in maximum *Rhizocarpon* diameter and also an increase in relative lichen abundance and species diversity. These results echo the findings of Pitman (1973), Koerner (1980) and Benedict (1990b).

The lack of epilithic lichen vegetation within snowpatches is probably due to unfavourable micro-environmental conditions (eg. reduced light intensities, and intolerably high moisture and humidity levels). Partial melting of snowpatch margins in the summer reveals a zone of ground that is only available for lichen colonisation during July, August and September (ie. 8-12 weeks) each year. This brief 'growing season' may not be long enough to guarantee the survival of even the hardiest lichen flora. Lichen-free areas and lichen 'trim-lines' in Fossdalur are therefore thought to be contemporary features reflecting the present-day (late 20th century) distribution of late-lying snow.

The effects of darkness and substrate instability

The ability of lichens to survive for long periods without light has been documented from Antarctica. Wise and Gressitt (1965) observed epilithic species growing on exposed rocks at 86°S, where the period of complete darkness exceeds 120 days. In Iceland, located just to the south of the Arctic Circle, there are a few hours of daylight even at the winter equinox.

Exposure to light has been demonstrated as a requirement for growth on several *Rhizocarpon* thalli at Virkisjökull, southeast Iceland. On July 20th 1999 measurements and photographs were taken of lichens on a small glacially-transported, basaltic boulder within an area of hummocky, ice-stagnation topography. Nineteen healthy, yellow-green *Rhizocarpon* Subgenus *Rhizocarpon* lichens (>5 mm) were growing on the upper surface of the boulder. The largest thalli measured 39.2 mm in diameter (Fig. 9.6). The stone was then overturned so that the lichen-covered surface lay face down in the sandy soil.

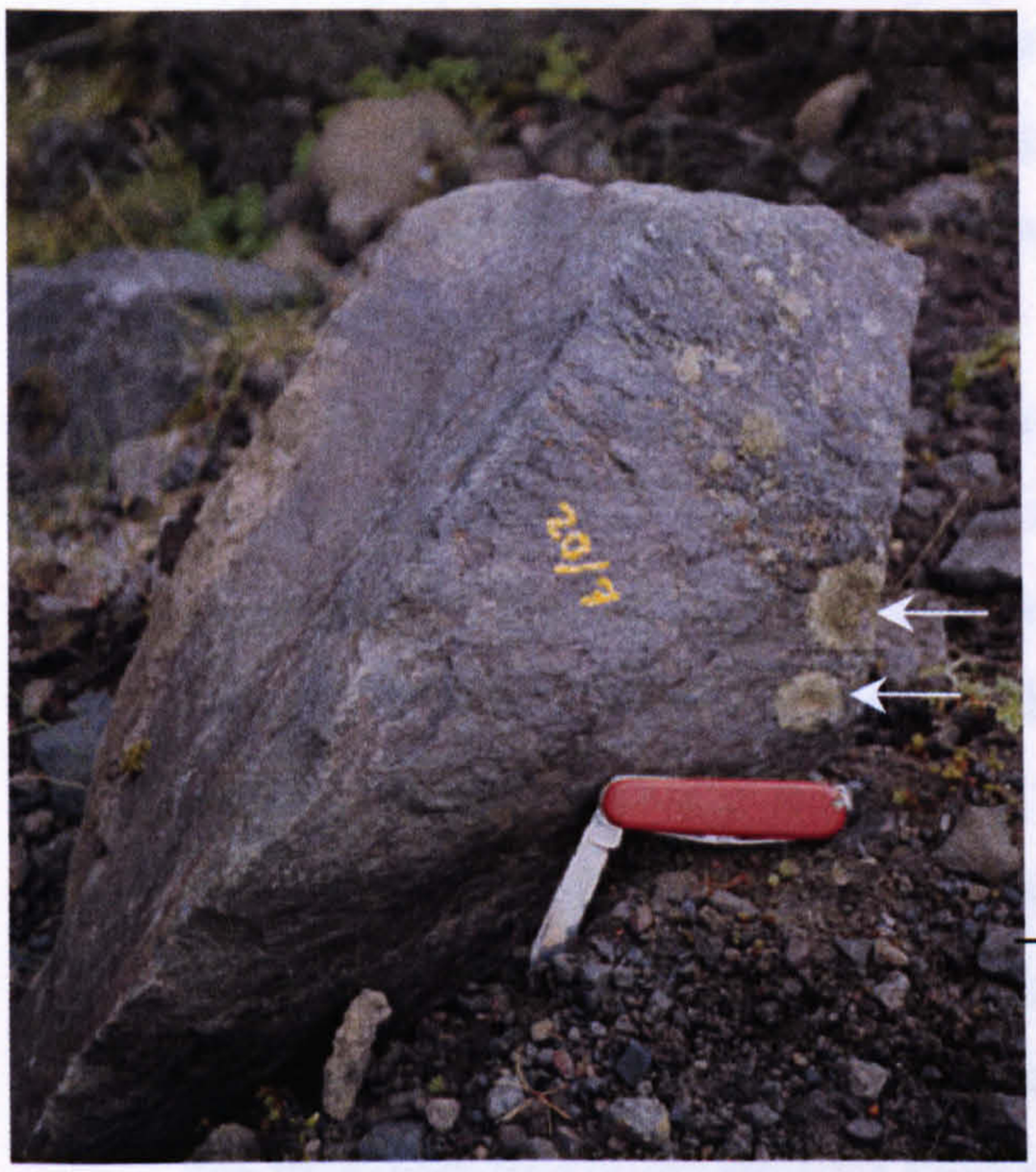


Figure 9.6 Micro-environmental factors affecting lichen growth – an experiment to test the effects of light exposure.

Prior to overturning (20.7.99). Note the numerous, healthy yellow-green *Rhizocarpon* lichens, particularly the two largest thalli at bottom right (arrowed).

(left) After one month the rock was restored the right way-up (20.8.99). Note the discoloured, fragmented appearance of many thalli. The two largest *Rhizocarpon* thalli have died and become detached from the rock surface (arrowed).

(right) Lichens and mosses growing on a large boulder at the foot of an ice-cored slope. The *Rhizocarpon* thalli on the uppermost surface are large (c. 40 mm) suggesting that the rock has not overturned during its 'descent'.

One month later, on the 24th August 1999 the stone was re-visited and examined. The lichen thalli on the underside of the boulder had been in complete darkness for 36 days. Their appearance had changed over the intervening time. Nine of the smaller lichens (<20 mm) had disintegrated or become detached from the rock. The largest specimens looked unhealthy, were darker in colour and had large parts of their thallus missing (Fig. 9.6).

This simple experiment suggests that exposure to sunlight, in summer, is essential for growth of *Rhizocarpon* lichens in Iceland. Relatively short periods of total darkness (4 wks) in poorly-ventilated, humid, conditions will result in thallus atrophication and ultimately death. This finding has implications for lichenometric studies. Lichens growing on unstable substrates that subsequently overturn will die when starved of light. This partly confirms the assertions of Griffey (1978) who stated that lichens growing on supraglacial debris would not survive deposition as the boulder moves or overturns. This would certainly be the case for lichens growing on the face onto which the boulder finally came to rest. However, lichens could survive deposition if they were growing on other faces where conditions may remain suitable for photosynthesis. Similarly, rock debris that slides rather than rolls to its resting place could potentially maintain a lichen population throughout its depositional 'journey'. One such 'ploughing' boulder was observed near the foot of an ice-cored slope at Virkisjökull (Fig. 9.6). Numerous, large *Rhizocarpon* lichens (>40 mm) are growing on the uppermost face of this boulder, suggesting that these thalli have survived deposition and will now probably continue to grow unhindered.

Summary of micro-environmental experiments

Various environmental factors control the colonisation rate and growth potential of epilithic lichens. By means of several simple experiments it has been shown that yellow-green *Rhizocarpon* lichens in southeast Iceland have preferred micro-habitats. South-facing, well-lit, well ventilated, rock surfaces normally exhibit the greatest degree of lichen cover. Whilst, ground that remains snow-covered into late summer exhibits the least degree of lichen cover.

These findings have implications for lichenometry. Locating areas of optimal lichen growth are of interest to workers conducting lichen-dating studies, as it is normally the maximum growth rate which is sought. Additional knowledge regarding

the ability of lichens to flourish whilst being transported and the potential for certain thalli to survive deposition reduces the chance of deriving erroneous lichenometric dates whilst increasing the robustness of the technique.

Macro-environmental (regional) differences in lichen growth

Repeated measurement of thalli over many years is the only way to construct a true lichen *growth* curve. However, this section aims to compare the growth rate of lichens in the mild, wet southeast of Iceland with those in the cooler and drier climate of northwest Iceland. Gravestones were used to construct two regional lichenometric age-size curves. The resulting age-size relationships will be treated as if they approximate to lichen growth, as has been done by various workers (eg. Beschel, 1950, 1961; Carrara and Andrews, 1973; Innes, 1983c).

Methods

Over 650 gravestones in the southeast and northwest of Iceland were examined during May and July of 1999 for the yellow-green *Rhizocarpon* lichen. Where *Rhizocarpon* Section *Rhizocarpon* lichens were observed, the longest axis of each thallus was recorded along with the date inscribed on the gravestone. Lichen measurements were made to the nearest millimetre using a transparent ruler. Very small thalli were identified using a x10 hand lens but only measured if over 1 mm in diameter. Some thalli within other *Rhizocarpon* groups may have been included, as identification of small thalli (<5 mm) is difficult. For each gravestone a note was made regarding the inclination (horizontal, vertical or inclined) and orientation (N, S, E or W) of the stone along with the rock type. The nature of the surface on which the lichens were growing was also carefully noted. Lichens are expected to be larger on the unpolished faces of a gravestone where colonisation can take effect more rapidly. Surface roughness is an important factor influencing lichen establishment and growth, and was deemed particularly noteworthy especially if the results were to be used in a broader dating context.

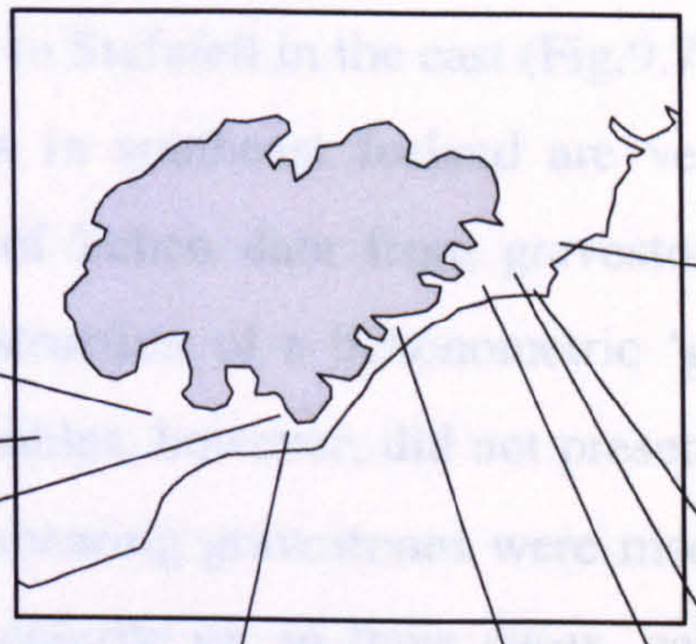
Figure 9.7 Churches visited during lichenometric gravestone surveys in NW and SE Iceland

Núpsstaður kirkja

Kirkjubæklaustur

Sandfell kirkja

A. Southeast



Stafafell kirkja

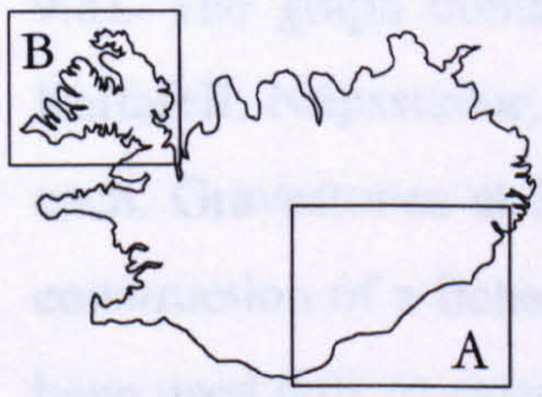
Höfnkirkja

Bjarnarnes kirkja

Brunnhólskirkja

Hofkirkja

Kálfafellsstaður kirkja



B. Northwest

Bolungarvík kirkja

Sæbol kirkja

Isafjörður kirkja

Núpur kirkja

Holtkirkja

Mýrar kirkja

Thingeyri kirkja

Flateyri kirkja

Results

A. Southeast Iceland

Nine churches were visited in Skaftafellsýssla, southeast Iceland, from Kirkjubærklaustur in the west to Stafafell in the east (Fig.9.7).

Lichen-bearing gravestones in southeast Iceland are very few in number. This factor makes the collection of lichen data from gravestones in southeast Iceland difficult and, hence, the construction of a lichenometric 'growth' curve potentially problematic. Lithological variables, however, did not present a problem at any of the churchyards as all the lichen-bearing gravestones were made of basic igneous rock, normally porphyritic basalt, dolerite or, in three cases, gabbro. Considering these limitations, it was decided to analyse the remaining results in order to derive a *Rhizocarpon* 'growth' curve for the region.

The data from the graveyards where Section *Rhizocarpon* lichens had been identified were plotted on a scattergram with age as the independent variable (Fig. 9.8). The graph contains data from eight churchyards although the datasets from Kalfafell, Núpsstaður, Hof and Brunhóllskirkja contain fewer than five data points each. Gravestones at Kalfafellstaður and Stafafell offered the most potential for the construction of a lichen 'growth' curve. Graves at Höfnkirkja, due to their age, have been used only to calculate the colonisation lag time.

B. Northwest Iceland

Eight churchyards were visited in Ísafjarðarsýsla, centred around the inlet of Önundarfjörður. As in the southeast of Iceland, very few yellow-green *Rhizocarpon* lichens were observed on gravestones in the Ísafjörður district. In comparison to Skaftafellssýsla, settlements are more numerous and churches are therefore more densely distributed. Unfortunately, very few old graves were found in the eight churchyards visited and, relative to the number of stones examined, fewer lichens were observed than in the southeast. Out of the 378 gravestones examined only 35 bore *Rhizocarpon* lichens, whilst in the southeast 60 *Rhizocarpon*-bearing gravestones were identified out of a total of 273. This factor makes construction of a meaningful lichenometric growth curve somewhat difficult.

Of the eight sites visited, Bolungarvík cemetery offered the most potential for assessing the rate of lichen growth over time. Interestingly, no Section *Rhizocarpon* thalli were found in Bolungarvík churchyard, 400 m from the cemetery. Thingeyrí

graveyard, situated close to sea-level, with over 100 stones, yielded only nine *Rhizocarpon*-bearing stones, and Ísafjörður churchyard, built in the 1960s, could only be used to help determine colonisation rates. The results of the gravestone survey in ísafjarðarsýsla have been plotted on a single scattergram (Fig. 9.8).

From the data available (55 stones) it has been possible to construct a maximum-growth (age-size) envelope for Section *Rhizocarpon* growing on gravestones in southeast Iceland (Fig. 9.9). A similar maximum 'growth' (age-size) curve has been constructed for northwest Iceland, also using the single largest-lichen measured on each of the *Rhizocarpon*-bearing gravestones (Fig. 9.9).

The lichen 'growth' curves

A. Southeast Iceland

According to measurements of lichens on gravestones in southeast Iceland, the growth rate of *Rhizocarpon* Section *Rhizocarpon* lichens is generally constant over the last eight decades. Although there is an apparently linear relationship between substrate age and lichen diameter, the 'growth' curve probably represents only part of a greater curve with an exponential form (see, for example, Fig. 8.3). In southeast Iceland, according to this study, linear growth lasts for at least 80 years, during which time the thallus increases in diameter by roughly equal radial increments from c. 10 mm up to c. 50 mm. The single lichen over 50 mm in diameter, on a gravestone dating from AD 1866, falls some way below the optimum 'growth' curve and may be growing (or have grown) more slowly. Unfortunately, no firm conclusions can be drawn regarding the growth rate of larger thalli (>50 mm) due to the lack of supporting data. There is also a suggestion that the smallest thalli undergo an initial phase of accelerating growth lasting for approximately 15 to 20 years. This would correspond with the juvenile or 'pre-linear' phase of growth identified by Hale (1974), Armstrong (1975) and Proctor (1977). However, in this case, the 'pre-linear' phase may simply be an artefact of insufficient sampling of young surfaces.

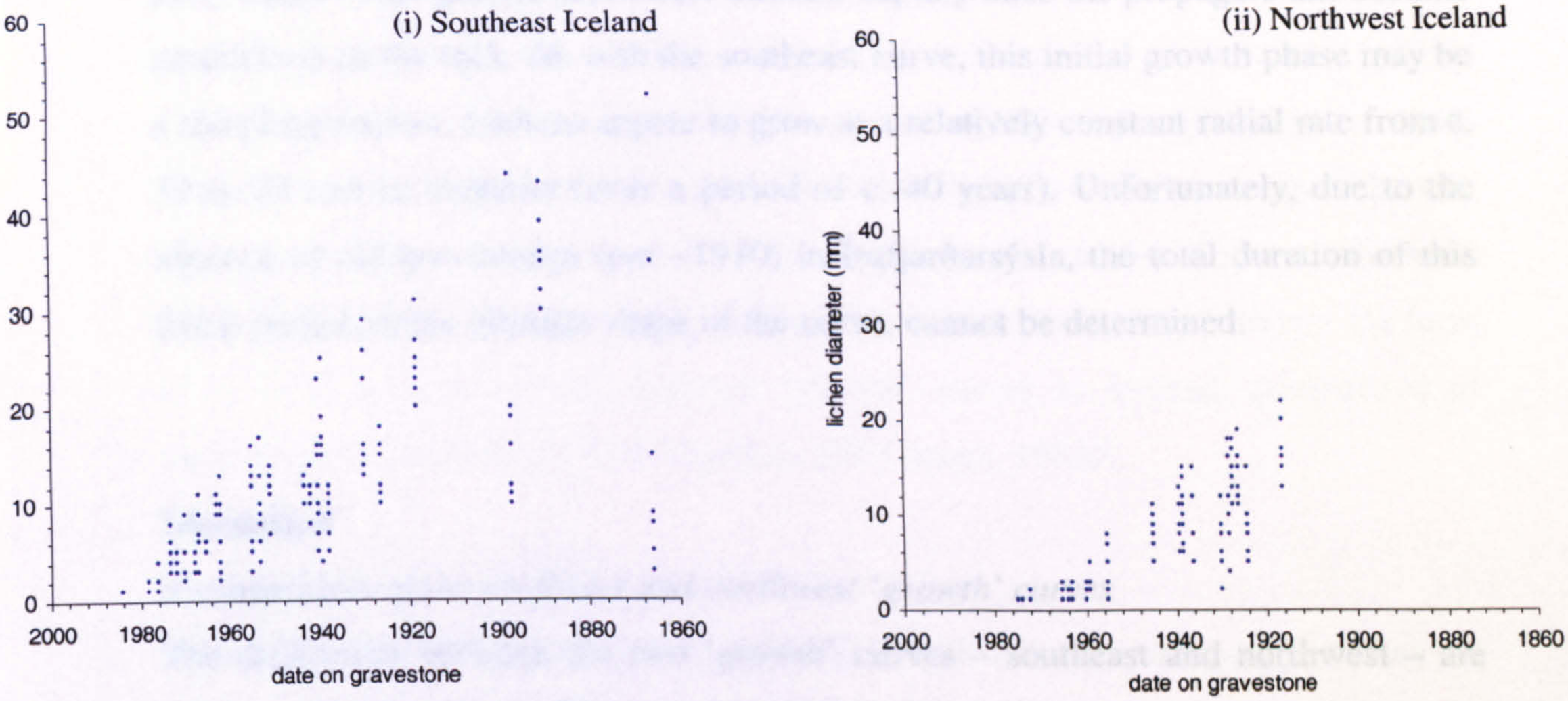


Figure 9.8 (above) Results of lichenometric surveys on gravestones in Iceland.

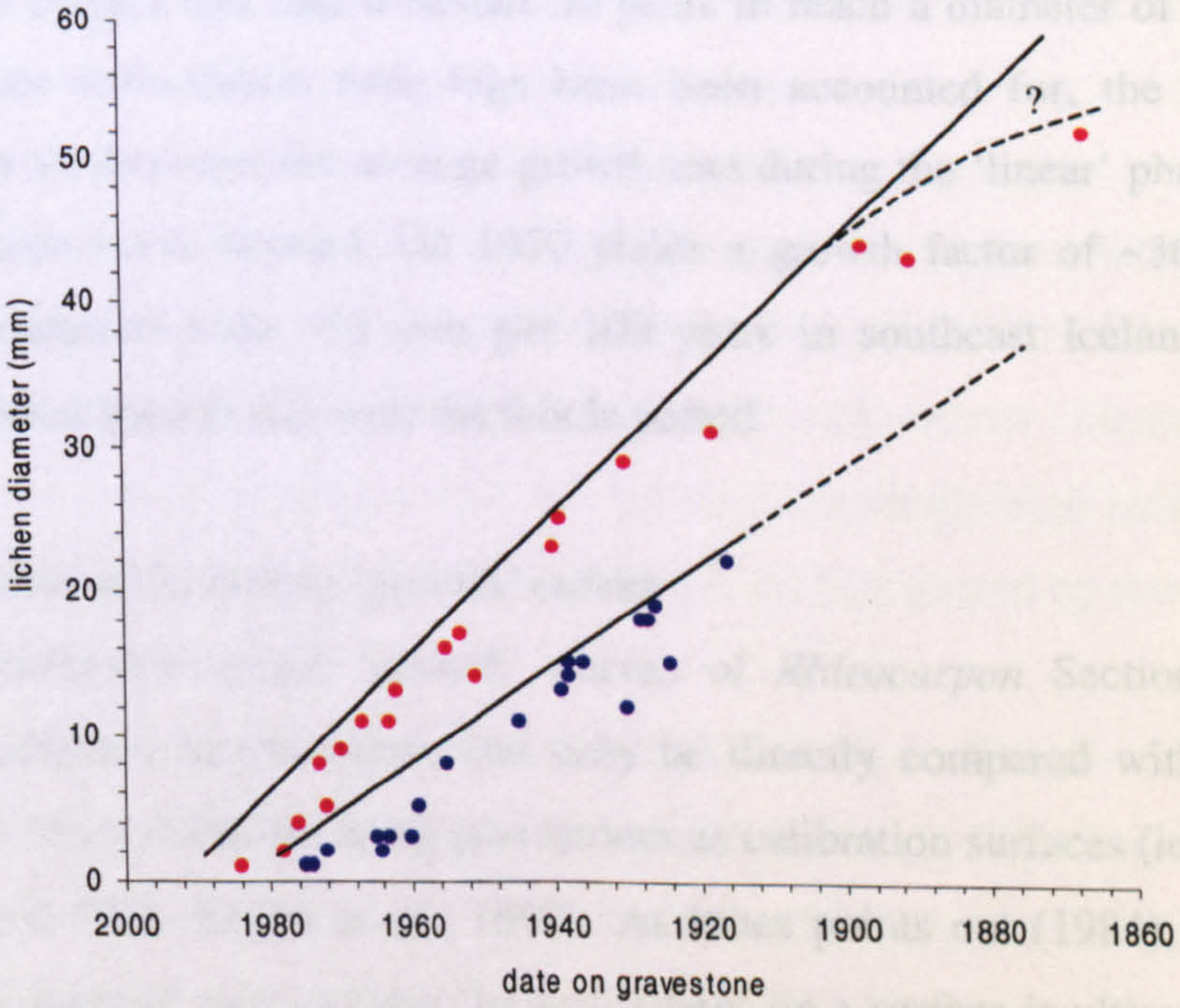


Figure 9.9 (above) Optimum 'growth' envelopes for *Rhizocarpon* Section *Rhizocarpon* lichens growing on unpolished gravestones in southeast (red) and northwest Iceland (blue). The lines give an indication of the maximum average 'growth' rate over the last c. 100 years (ie. mm/100 yr).

B. Northwest Iceland

According to measurements of *Rhizocarpon* Section *Rhizocarpon* lichens on gravestones in northwest Iceland the growth of these lichens is also apparently linear over a period of 70 years. There is an indication of a juvenile, 'pre-linear' phase (<10 mm) where radial growth rates must increase rapidly once the propagule has become established on the rock. As with the southeast curve, this initial growth phase may be a sampling artefact. Lichens appear to grow at a relatively constant radial rate from c. 10 to 23 mm in diameter (over a period of c. 40 years). Unfortunately, due to the absence of old gravestones (pre ~1910) in Ísafjarðarsýsla, the total duration of this linear period, or the ultimate shape of the curve, cannot be determined.

Discussion

A comparison of the southeast and northwest 'growth' curves

The differences between the two 'growth' curves – southeast and northwest – are clear. Yellow-green *Rhizocarpon* lichens become established on unpolished gravestones after only c.10-15 years in southeast Iceland and can be 10 mm in diameter after only 15 years of growth. In northwest Iceland colonisation takes longer. Lichens normally require at least 20 years to become established on an unpolished surface and take a further 30 years to reach a diameter of 10 mm. Once the different colonisation time lags have been accounted for, the most obvious differences are between the average growth rate during the 'linear' phase. Projecting the northwest curve beyond AD 1910 yields a growth factor of ~36 mm per 100 years – compared with ~52 mm per 100 years in southeast Iceland, assuming a constant radial growth rate over the whole period.

A comparison with existing 'growth' curves

The indirectly-determined 'growth' curves of *Rhizocarpon* Section *Rhizocarpon* lichens presented in this study can only be directly compared with those curves derived by other scientists using gravestones as calibration surfaces (ie. Jaksch, 1970; Kugelmann, 1991; Evans *et al.*, 1999). As Innes points out (1984), the size – and hence the 'growth' rate – of the 'largest-lichen' on a surface is ultimately a function of the area searched. 'Growth' curves calibrated using larger surfaces (ie. moraines, alluvial terraces, etc.) are not directly comparable with those derived in this study.

The 'growth' rate of 0.56 mm a^{-1}, derived using gravestones in southeast Iceland by Evans *et al.* (1999), appears very similar to that derived in this study. The 'growth' rate derived by Kugelmann in northern Iceland (0.44 mm a^{-1}) is slightly higher, although similar, to that identified by the present author in northwest Iceland; whilst Jaksch's (1970) 'growth' rate of 0.73 mm a^{-1}, for southern Iceland, based on only five gravestones, is considerably higher.

If the new 'growth' curves derived in this study were applied as a geo-chronological dating tool, they would need to be adapted to take into account the size of the sampling area normally dealt with in lichenometric studies. However, the form of the *Rhizocarpon* 'growth' curves derived from gravestones should mirror the form of all indirectly-derived *Rhizocarpon* 'growth' curves in Iceland, irrespective of sampling area (Matthews, 1974; Topham, 1977; Innes, 1983c).

Regional differences in lichen growth

A study by Topham (1977) into the growth rate of *'Rhizocarpon geographicum'* on gravestones in Scotland drew several conclusions. Topham attributed the difference observed between the growth rate of thalli in western and eastern Scotland to the marked precipitation gradient from west to east. Lichens in sites receiving more rainfall showed more rapid growth than those in drier, easterly locations. Innes (1983) has pointed out that these growth rate differences may also be related to air temperature, height above sea level or possibly due to differences in lichen age at the three study sites.

A comparison of other studies utilising gravestones to derive lichen growth rates (eg. Beschel, 1950, 1961; Carrara and Andrews, 1973; Calkin and Ellis, 1980; Porter, 1981), shows that growth is more rapid in wetter and warmer climates. Similarly, a general decrease in growth rate can be seen at increasingly high latitudes, probably due to the cooler and drier conditions, although the part played by snowcover may be a significant one.

From the results presented in this study a difference in the growth rate of Section *Rhizocarpon* lichens between the northwest and southeast of Iceland can be seen. In northwest Iceland it takes ~70 years, once established, for thalli to grow at the optimum rate to a diameter of 23 mm, compared to ~40 years in the southeast. This can be expressed as a difference of approximately 40% between the growth rates of Section *Rhizocarpon* thalli in the two regions. The reasons for this difference appear

to be climatic and may reflect the 'continentality' of the climate, as first proposed by Beschel (1961). Higher annual precipitation (2000 mm *cf.* 1400 mm) combined with higher mean annual air temperatures (4.5°C *cf.* 3.3°C) will promote more rapid growth in southeast Iceland. An increased frequency of rain-bearing, westerly winds and higher relative humidity, probably also serves to promote lichen growth in the more maritime climate of southeast Iceland. Whereas the high frequency of freezing temperatures and cold northerly winds, coupled with the effect of increased snowcover and an effectively shorter growing season, probably play a vital role in suppressing lichen colonisation and retarding growth rates in the higher latitude region.

Conclusions

Although the limited number of lichen-bearing gravestones in Iceland has made the construction of 'growth' curves difficult, the following conclusions can be drawn from this study:

1. Yellow-green *Rhizocarpon* lichens are relatively rare on gravestones in SE and NW Iceland, probably due to the location of most churchyards close to the sea where sub-arctic/alpine lichen species – such as *Rhizocarpon* – do not flourish.
2. Lichen growth rates and colonisation rates vary considerably within a single region and even within a single graveyard in Iceland. These variations are seen as a result of differences in micro-climate and micro-habitat among the numerous sites.
3. Colonisation lag times generally depend on the roughness of the stone surface. Rougher surfaces are colonised before smooth, polished ones in NW and SE Iceland.
4. Colonisation lag times on unpolished gravestones are 30-50% longer on average in NW Iceland compared with SE Iceland.
5. The two lichenometric 'growth' curves, derived from gravestones, span the last 130 years and 90 years, in SE and NW Iceland respectively. In southeast Iceland, a 'linear' growth phase lasts for at least 80 years; in the NW this 'linear' phase lasts at least 50 years, although its exact duration has not been determined.

6. There is a suggestion that the smallest and, in SE Iceland, the largest *Rhizocarpon* lichens grow more slowly than those between ~10 and ~50 mm in diameter.
7. Lichen growth rates are on average ~40% faster in southeast Iceland than northwest Iceland. A maximum growth factor of 52 mm per 100 years is derived for largest-lichens of Section *Rhizocarpon* on gravestones in SE Iceland, assuming a linear relationship over the whole period. A growth factor of 36 mm per 100 years is extrapolated for NW Iceland.
8. Lichen growth rates appear to be related to climate. Yellow-green *Rhizocarpon* lichens grow faster in SE Iceland as a result of the more maritime climate than in the NW.

Summary – macro-environmental (regional) differences in lichen growth

Age-size ('growth') curves of Section *Rhizocarpon* lichens have been carefully constructed based on measurements of thalli growing on gravestones in southeast and northwest Iceland. The results of this work have two important implications for lichenometric dating studies in Iceland:–

- Firstly, the scarcity of gravestones in SE Iceland and the scarcity of lichens growing on gravestones in NW Iceland resulted in the construction of two lichenometric 'growth' (age-size) curves based on a limited sample size. Such curves are likely to under-represent optimal lichen growth rates in that environment by a considerable degree.
- Secondly, and more importantly, these results suggest that growth rates of Section *Rhizocarpon* vary across Iceland and that these differences in lichen growth rate can be attributed to the degree of climatic 'continentality'. A more maritime climate, with higher precipitation, warmer mean annual temperatures, and limited snow cover, in the SE promotes lichen growth. Therefore, it is not practical to extrapolate locally-derived dating curves over large distances when environmental conditions differ considerably. Rather, it is important that local dating-curves be constructed when embarking on lichenometric-dating exercises in different climatic regions.

Environmental controls on lichen growth rates

A North Atlantic case study

Introduction

Previous attempts to relate lichen growth to climate have been performed on various timescales. Beschel (1956, 1957, 1961) found that growth rates (expressed as mm per century) of crustose lichens in Alpine Italy and west Greenland were related to mean precipitation totals. The striking decline in growth rates of *Rhizocarpon* in drier, inland sites relative to those nearer the coast prompted Beschel (1961: Fig 1.) to conclude that growth rates were broadly related to the 'continentality' of the climate. Karenlampi (1971) found a significant correlation between growth of the foliose lichen *Cladonia alpestris* and mean daily precipitation in subarctic Finland. Lawrey and Hale (1977) found a correlation between growth rates and mean monthly precipitation in several foliose species in Maryland, USA. Repeated measurements of *Parmelia caperata* in Devon, UK, showed that growth was most rapid in those weeks experiencing greater cloud cover and a higher frequency of rain days (Fisher and Proctor, 1978). Working in the mountains of Colorado, Benedict (1990a) found the strongest correlation between daily growth rates and the duration of daylight, snow-free, thallus moisture (i.e. the period during which photosynthesis could occur). Benedict believed that this factor accounted for c. 92% of seasonal growth variability in *Xanthoparmelia* thalli. Weaker correlations were also found between mean air temperature and snowcover conditions. Interestingly, Benedict did not find a correlation between growth rate and precipitation, in the dry continental climate of Colorado.

It was decided to examine which environmental factors have most influenced growth rates of *Rhizocarpon* lichens over the last 100 years in Northwest Europe and Greenland. Such a study would serve to strengthen the link identified by Beschel (1961) between lichen growth rates and climate.

Methods

Using previously published growth rates from numerous studies conducted in the Northeast Atlantic region (c. 50°-70°N and 20°E-50°W) linear regression analyses were performed. Ten primary environmental factors were evaluated:

1. Mean annual air temperature (°C)
2. Mean annual precipitation (mm a^{-1})
3. Mean annual temperature range (°C) [*continentality*]
4. Annual precipitation divided by annual temperature range (mm°C^{-1}) [*oceanicity*]
5. Average number of rain days (days a^{-1})
6. Average number of fog days (days a^{-1})
7. Average duration of snow lie (days a^{-1})
8. Average relative humidity (%)
9. Average number of gale days (>Force 9) (days a^{-1})
10. Latitude (ie. summer day length) (°N)

Unfortunately meteorological data regarding other variables such as cloudiness, snow-depth and precipitation rates (mm/minute) are not generally available for most weather stations and could not be included in the regression analysis. These latter factors may be important for lichen growth but those ten variables, chosen above, are most likely to influence growth over the timescale in question. Most lichenometric studies have been concerned with dating the recession of glaciers in recently deglaciated areas (eg. Austrian Alps, montane Norway, Iceland, coastal Greenland). Although field methods vary between authors, the derived growth rates allow comparison to be made between the studies in all but a few cases.

Directly-measured radial lichen growth rates are often quoted in *mm per year.* However, in lichenometric studies, where growth rates are derived indirectly, such expressions are inappropriate. The term *lichen factor* has been used in lichenometric studies in reference to the size of the largest lichen on a 100-year-old substrate (Beschel, 1958, 1961). Use of the *lichen factor* allows comparison between localities with different length colonisation periods and juvenile growth stages (often referred to as the 'Great Period'). For this reason, *lichen factors* (mm/100 yr) used here are not merely a multiplication of the published 'growth rates' (ie. mm/yr x 100).

Table 9.1 The lichenometric studies used in the regression analysis.

Country	*Author(s)*	*Location*	*Lichen factor (mm/100 yr)*
Iceland	Gordon and Sharp (1983)	Skálafellsjökull	84
	Maizels and Dugmore (1985)	Sólheimajökull	85
	Kugelmann (1990)	Tröllaskagi	44
	Evans *et al.* (1999)	Heinabergsjökull	80
	Evans *et al.* (1999)	Brúárjökull	42
	Sigurvinsson (*unpub.*)	Kaldalónsjökull	40
	Bradwell (*this study*)	Hólárjökull	67
Norway	Bornefeldt and Osterborg (1958)	Nigardsbreen	60
	Matthews *et al.* (1979)	Nigardsbreen	68
	Haines-Young (1983)	Nigardsbreen	52
	Erikstad and Søllid (1986)	Nigardsbreen	65
	Ballantyne (1990)	Lyngsalpene	40
Greenland	Beschel (1961)	Gødhavn	15
	Beschel (1961)	Ikattssaq	32
	Ten Brink (1973)	Strømfjørdshavn	17
	Ten Brink (1973)	Orkendalen	7
Scotland	Innes (1983c)	Roybridge	104
	Innes (1983c)	Ballachullish	96
	Innes (1983c)	Edzell	45
	Innes (1983c)	Bankfoot	40
Svalbard	Werner (1990)	NW Spitsbergen	14

It was hoped to draw on all the published *Rhizocarpon* lichenometric studies. However, differences in sampling method prevent the inclusion of some of the published growth rates in the analysis. For example, those studies that use the shortest lichen axis (Thompson and Jones, 1986; Gudmundsson, 1998b) could not be used, as the relationship between the largest and shortest axes in these studies is not known. Generally, the lichenometric studies are comparable where they entail using the longest axis of the largest or five largest lichens on a surface as an indication of age. The question of sample area is a complicating factor when seeking to compare lichenometric dating curves (Innes, 1984). Most of the studies listed in Table 9.1 use entire moraine ridges as the sampling area, covering areas typically in excess of 1000 m^2 each. Exceptions to this are the studies conducted in Scottish graveyards by Innes (1983c), where the sampling area is small. These 'limited-area' studies have only been included where no other calibration is available in a distinct climatic region.

Table 9.2. Climatic data (1931-1960) from lichenometric study sites

Study	Nearest met. station[1]	mean annual temperature (°C)	mean annual precipitation (mm)	mean annual temperature range (°C)	no. of rain days per year	no. of fog days per year	snow lie weeks per year	relative humidity (%)	gale days (>F 9) per year
Gordon / Sharp (1983)	Holar / Fagurhólsmyri[2]	4.9	2000[a]	11.2	200	10.5	8.8	78	2.5
Maizels / Dugmore (1985)	Vík	5.7	2260	11.0	230	40	12	82	7.2
Kugelmann (1990)	Akureyri[3]	-2.0	1600[a]	12.6	45[e]	35[e]	30[e]	80[e]	12[e]
Evans *et al.* (1999)	Holar / Fagurhólsmyri	4.9	2000[a]	11.2	200	10.5	8.8	78	2.5
Evans *et al.* (1999)	Grimsstaðir[4]	1.2	1200[a]	14.0[a]	40	5	24	84	2.3
Sigurvinsson (unpub.)	Isafjörður	4.1	1100	11.8	177	15	36.4	78	15
Bradwell (this study)	Holar / Fagurhólsmyri	5.7	2000	11.2	200	10.5	8.8	78	2.5
Andersen / Sollid (1971)	Bjørkehaug i Jostedal[5]	4.5	2000	17.0	215	6	25	81	8
Matthews *et al.* (1979)	Bjørkehaug i Jostedal	4.5	2000	17.0	215	6	25	81	8
Haines-Young (1983)	Bjørkehaug i Jostedal	4.5	2000	17.0	215	6	25	81	8
Erikstad / Sollid (1986)	Bjørkehaug i Jostedal	4.5	2000	17.0	215	6	25	81	8
Ballantyne (1990)	Tromso	2.4	850[a]	16.0	110	5	30	76	2
Beschel (1961)	Gødhavn	-2.5	400	22.0	40	16	30	78	10
Beschel (1961)	Gødthaab[6]	-1.4	720	17.5	55	20	20	80	16
Ten Brink (1973)	Søndre Strømfjord[7]	-6.0	240	35.7	30	10	38	66	8
Ten Brink (1973)	Søndre Strømfjord[8]	-7.0[e]	200[a]	40.0[e]	25[e]	6[e]	40[e]	64[e]	8[e]
Innes (1983c)	Fort William	10.2	2200	10.0	225	4	1.5	77	10
Innes (1983c)	Fort William	10.2	2200	10.0	225	4	1.5	77	10
Innes (1983c)	Brechin	8.0	850	11.8	165	8	5	80	6.5
Innes (1983c)	Ladybank	8.2	800	11.0	160	6	4.5	79	3.5
Werner (1993)	Ny Ålesund	0.8	380	18.0	38	12	22	75	7

Notes

1 – Nearest representative meteorological station with data referring to the 8 chosen variables, spanning the 30-year reference period (1931-1960).

2 – Data for coastal southeast Iceland are averages of values from the weather stations at Hólar i Hofn and Fagurhólsmyri.

3 – Akureyri weather station (altitude: 25m asl) is situated ~400-800 m below the study site of Kugelmann (1990). Therefore, data for the Tröllaskagi highlands have been generated based on other meteorological measurements made in the area. Figures for air temperature assume an average annual lapse rate of 0.6°C/100 m (Björnsson, 1971) and are extrapolated from Akureyri. Values for all other variables are estimated from the measurements at Akureyri and of Björnsson (1971) taken in the summers of 1967 and 1968. Some uncertainty surrounds these data due to the lack of measurements at this site.

4 – Grimsstaðir, although 100 km from the study site at Brúárjökull, is deemed to be the weather station most representative of conditions around northern Vatnajökull. As no meteorological measurements have been made in the vicinity of Brúárjökull, unadjusted data from Grimsstaðir has been used.

5 – Bjørkehaug i jostedal is the nearest weather station to Nigardsbreen (<15 km away) and is situated at a similar altitude (324m asl). Data only extend back to 1964. All figures relate to the period 1964-1974 and were provided by Norske Meteorlogiske Institutt.

6 – Gødthaab (Nuuk) weather station (operating since 1866) is deemed most representative of coastal west Greenland between 64 and 66°N.

7 – Søndre Strømfjord weather station has only been in operation since 1970. Data relate to the period 1970-1990.

8 – Data for Orkendalen, situated 60 km east of Søndre Strømfjord at an altitude of 200 m asl., have been estimated based on data from Strømfjørdshavn (1970-1990) and those presented in Ten Brink (1973).

a – determined from official climatic maps (1930-1961).

e – estimated value based on measurements made in the vicinity (see footnotes 3 and 7).

In order to analyse the relationship between lichen growth rates and environmental factors, climatic data were obtained from the nearest weather station in each of the 21 study sites listed above (Table 9.2). Unfortunately, lichenometric dating studies are, by their very nature, often confined to remote, mountainous areas some distance from weather-monitoring stations. In certain localities, the nearest weather station is not representative of the climatic conditions at the study site. In such cases (eg. Tröllaskagi, Iceland; Brúárjökull, Iceland; Orkendalen, west Greenland), temperature and precipitation data have been taken from official climatic maps. However, where possible, data from the nearest weather station, relating to a 30-year reference period (1931-1960), have been used in the regression analysis.

Results

The strength of the correlations are listed below:-

Variable	r^2	p
Mean annual air temperature (°C)	0.64	0.05
Mean annual precipitation (mm a^{-1})	0.85	0.02
Mean annual temperature range (°C)	0.65	0.05
Annual precipitation / annual temperature range (mm $°C^{-1}$)	0.95	0.005
Average number of rain days (days a^{-1})	0.75	0.03
Average number of fog days (days a^{-1})	0.43	0.1
Average duration of snow lie (days a^{-1})	0.55	0.1
Average relative humidity (%)	0.01	na
Average number of gale days (>Force 9) (days a^{-1})	0.03	na
Latitude (ie. summer day length) (°N)	0.18	na

Growth of yellow-green *Rhizocarpon* lichens over the last 100 years is correlated with several climatic variables. Strong positive correlations are found between mean annual precipitation, air temperature, rain days and lichen growth rate (expressed as mm/100 yr), whilst a strong negative correlation is found between mean annual temperature range and growth rate. All of these correlations are significant at the 95% level. Other, weaker relationships exist between growth rate and fogginess (+ve) and snow lie (-ve), although these are only significant at the 90% level.

Of the ten variables, precipitation accounts for 85% of the variation in lichen growth rates across the Northeast Atlantic region. This result re-affirms the initial findings of Beschel (1954, 1957). However, the exact role of climatic 'continentality' on lichen growth as stated by Beschel (1961) is more difficult to determine as there is no agreed definition of this variable. Beschel may have been referring to the index of 'hygric continentality' proposed by his professor H.H. Gams (1932). This was defined as the quotient of mean annual precipitation and altitude (metres asl) and, when used in Austria, closely described the relative preference of tree species in alpine areas. Most continentality indices, however, depend crucially on mean annual temperature range (Gorczynski, 1920; Conrad, 1946; Barry and Chorley, 1992) The negative correlation ($r^2 = 0.64$) identified here between mean annual temperature range and lichen growth rates in northwest Europe would go some way to support Beschel's (1961) findings.

The environmental variable most highly correlated with lichen growth rates over the last 100 years is the ratio of mean annual precipitation to mean annual air temperature range. Kotilainen first proposed a form of this expression as a most suitable index of 'oceanicity'. *Oceanicity*, or the degree of oceanic influence on the climate, has been used in various botanical and ecological studies in Northern Europe in relation to vegetation distribution (Godske, 1944; Poore and McVean, 1957; Kirkpatrick and Rushton, 1990). Kotilainen's index considers both precipitation and temperature whilst also expressing the even distribution of heat throughout the year by taking the number of vernal and autumnal days into account (ie. when mean T >0 and <10°C). Maritime climates have high oceanicity indices (>100). A modification of Kotilainen's index has been used in the regression analysis (ie. annual pptn/annual temperature range) to estimate the relative oceanicity of the climate in each of the lichenometric study sites.

The ratio of annual precipitation to annual air temperature range (1931-1960), referred to as the *K-index*, is correlated with the size of the largest *Rhizocarpon* lichen growing on a 100-year-old surface at the 99.5% level. Therefore, broadly-speaking, lichen growth rates in Iceland, Greenland, Norway and the British Isles are related to mean annual precipitation and inversely related to mean annual

temperature range. For this reason oceanicity (defined as pptn / MATR) is the single environmental variable in the regression analysis that best correlates with lichen growth rates in the Northeast Atlantic and Greenland. The regression analysis shows that 95% of lichen growth factor variability can be explained by variations in oceanicity (Fig. 9.10).

Discussion – A predictive model for lichen growth rates

Using the strong relationship between oceanicity and *Rhizocarpon* growth rates in northwestern Europe it becomes possible to **predict** the maximum size of a lichen thallus growing on a 100-year old substrate in a given region. Calculation of the *K-index* at a specific site allows a maximum lichen growth factor to be derived using the regression equation ($y = 0.4x + 11.0$). The model is currently limited to northwestern Europe – being the area from which the calibration data came. However, application to areas further a-field – particularly those experiencing similar maritime climates (eg. western Patagonia; South Island, New Zealand) – will also be examined.

A preliminary map has been generated, using the oceanicity/growth rate relationship, showing the **predicted**, potential, maximum lichen growth rates across the Northeast Atlantic region (Fig. 9.11). Although currently operating at a relatively low spatial resolution, due to the lack of dating studies from climatically-diverse environments, this model should prove useful for future lichenometric work in the region. A scientist wishing to construct a lichenometric dating curve in a previously unstudied area could gain an impression of maximum *Rhizocarpon* growth rates prior to visiting the site merely by consulting readily available meteorological data. Furthermore, in the absence of any lichen-bearing dated surfaces from which to calibrate a dating curve, the oceanicity/growth rate relationship could be used to estimate the age of lichens growing on a substrate, if the relevant climatic data is known.

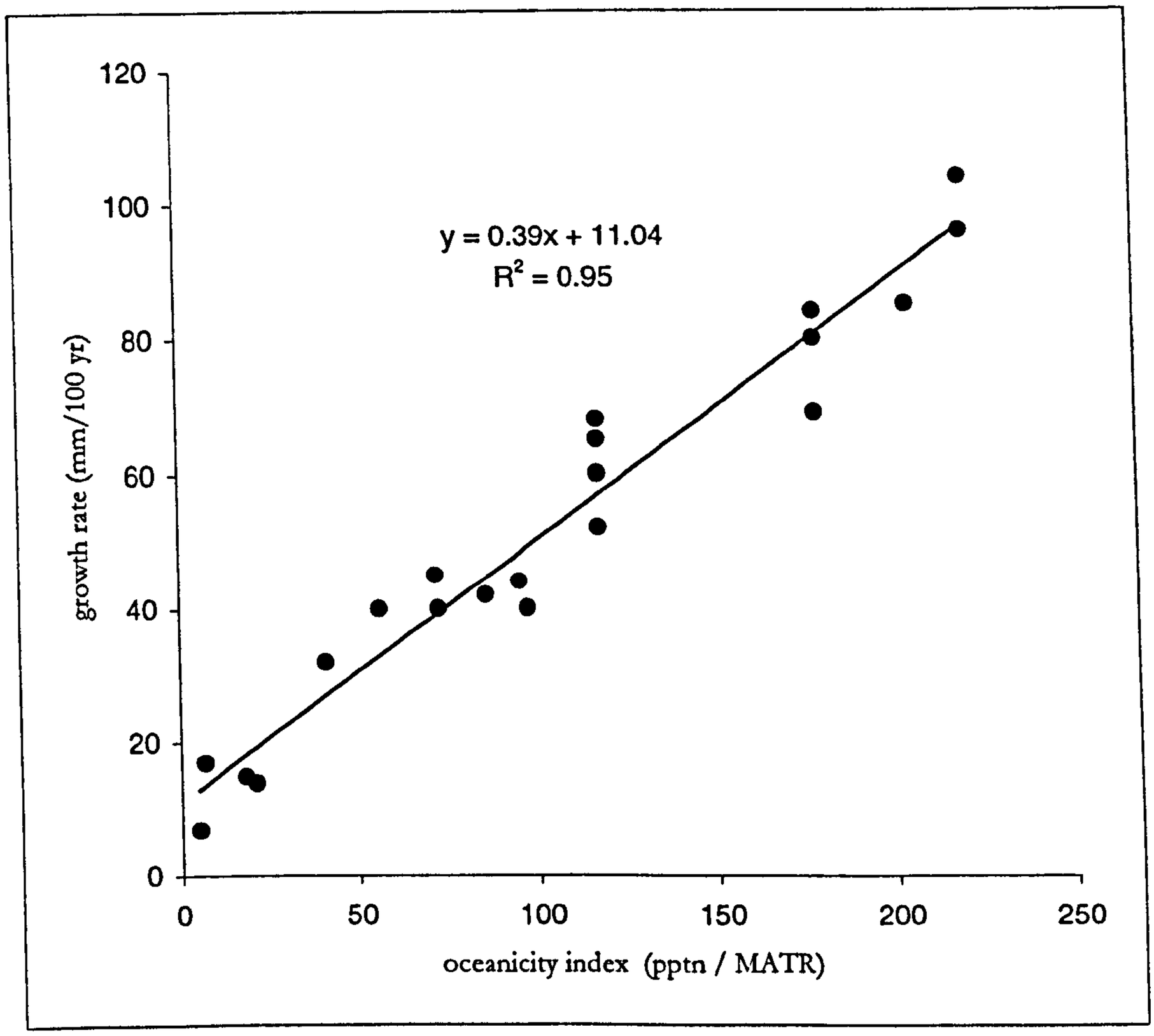


Figure 9.10 The relationship between the growth rate of lichens and climatic 'oceanicity' in Northwest Europe and Greenland.

The size of the largest Subgenus *Rhizocarpon* lichen growing on a 100-year-old surface is used as a direct indicator of growth rate (expressed in mm per 100 years). 'Oceanicity index' is defined as mean annual precipitation (mm) divided by mean annual air temperature range (°C). The graph collates data from 21 lichenometric studies conducted in the North Atlantic region (see Table 9.1).

Note the high r-squared value (0.95) and linear relationship.

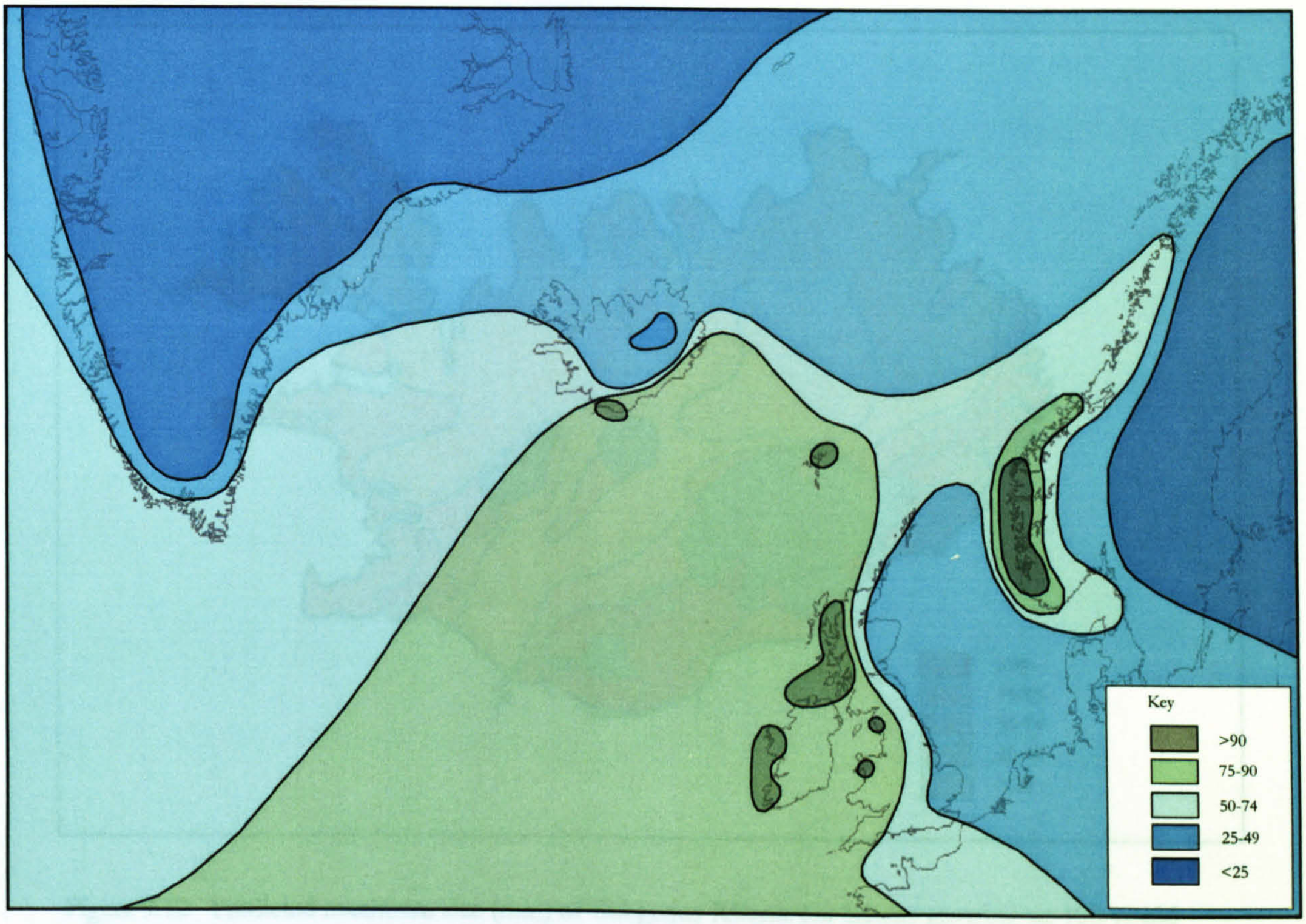
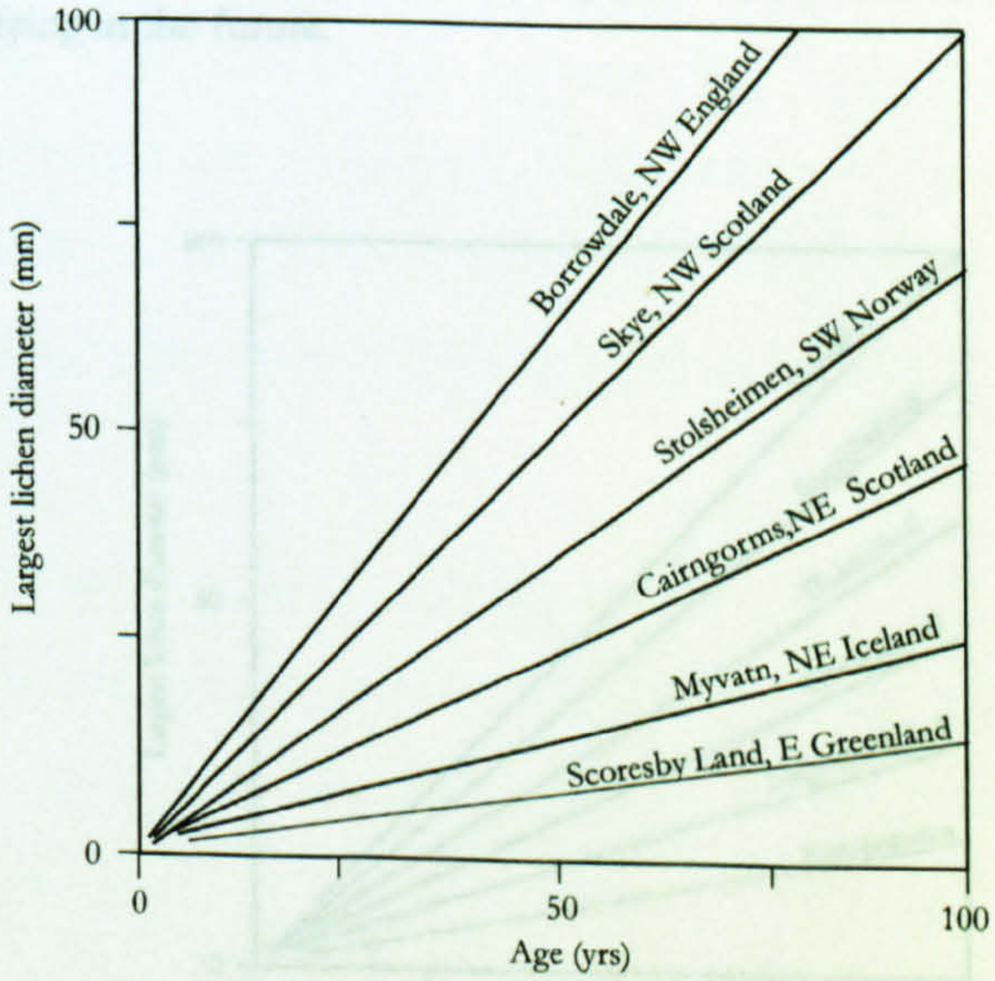


Figure 9.11 Predicted maximum size of Subgenus *Rhizocarpon* lichens growing on 100-yr-old surfaces in the Northeast Atlantic region. Values have been derived using the *growth rate/oceanicity* model established in this study ($y = 0.4x + 11.0$; $r^2 = 0.95$). (Climate data relate to the period 1931-1960.)



(Above) Predicted maximum growth rates of Subgenus *Rhizocarpon* lichens in selected localities around the Northeast Atlantic. Note that the modelled values assume constant growth over the last 100 years and no climatic change.

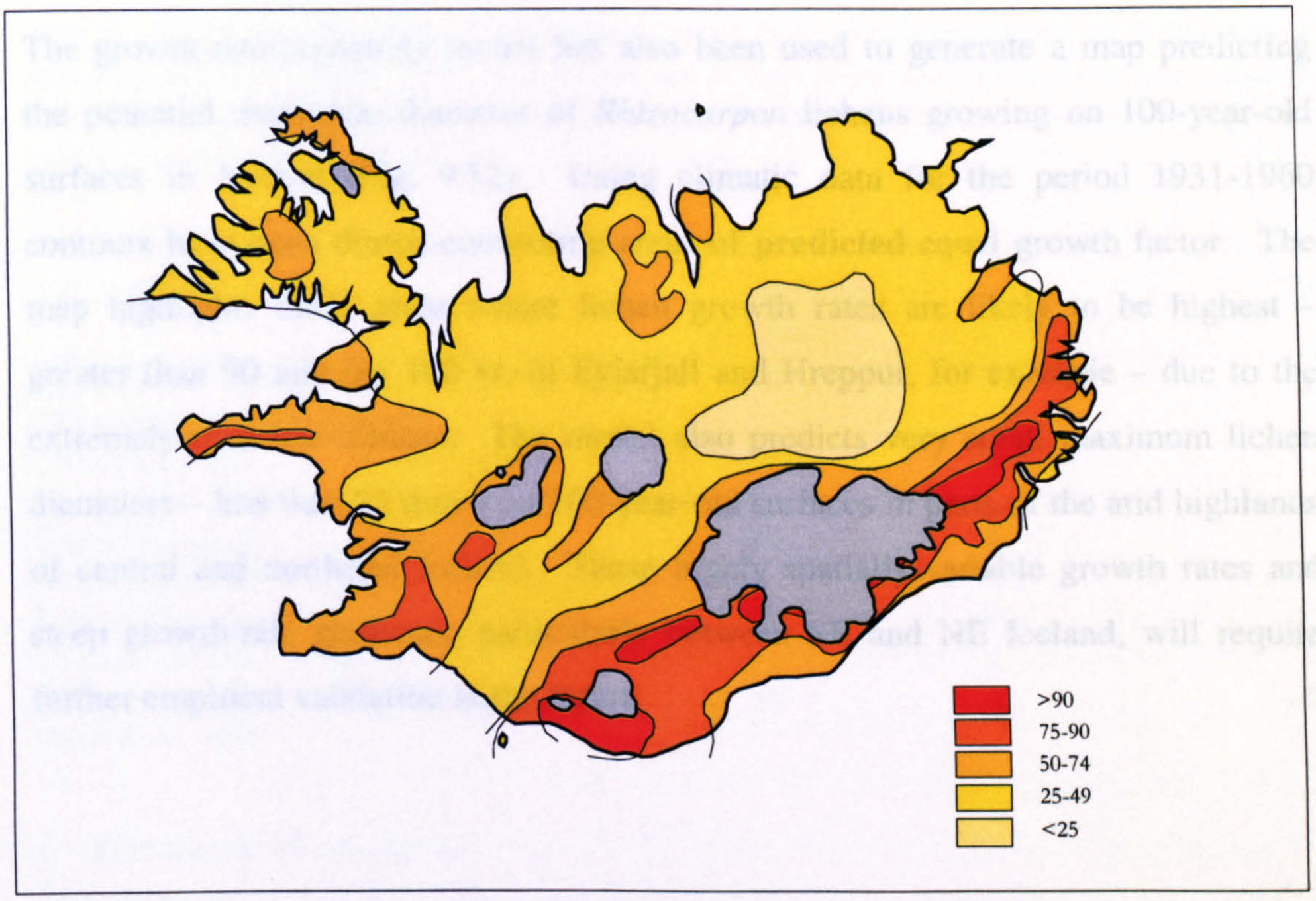
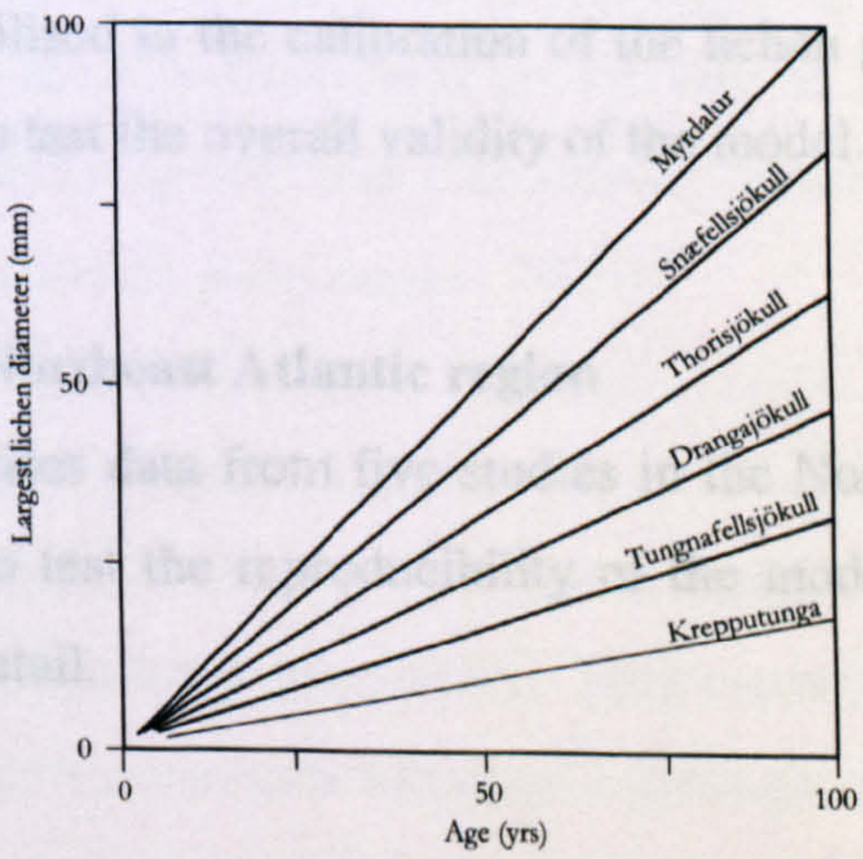


Figure 9.12 Predicted maximum size (mm) of Subgenus *Rhizocarpon* lichens growing on 100-yr-old surfaces in Iceland. Values have been derived using the *growth rate/ oceanicity* model established in this study. (Climatic data relate to the period 1931-1960.)

The map highlights the areas where lichen growth rates are likely to be most rapid. Southernmost coastal localities and the mountainous regions to the SW and NE of Vatnajökull are likely to yield the highest *Rhizocarpon* growth rates – greater than 90 mm per 100 years. In contrast, growth rates in large parts of the central highlands to the north of Vatnajökull are likely to be less than 25 mm per 100 years. These predictions will need verifying in the future.



(above) Predicted maximum growth rates of Subgenus *Rhizocarpon* lichens in selected localities in Iceland. Note that the modelled values assume constant growth over the last 100 years and no climatic change.
Such information could be used to estimate the age of lichen-bearing surfaces on glacier forelands in areas where calibration surfaces are lacking.

The growth-rate/oceanicity model has also been used to generate a map predicting the potential maximum diameter of *Rhizocarpon* lichens growing on 100-year-old surfaces in Iceland (Fig. 9.12). Using climatic data for the period 1931-1960 contours have been drawn connecting areas of **predicted** equal growth factor. The map highlights those areas where lichen growth rates are likely to be highest – greater than 90 mm per 100 yr, in Eyjafjall and Hreppur, for example – due to the extremely maritime climate. The model also predicts very small maximum lichen diameters – less than 25 mm – on 100-year-old surfaces in parts of the arid highlands of central and northeast Iceland. These highly spatially-variable growth rates and steep growth-rate gradients, particularly between SE and NE Iceland, will require further empirical validation in the future.

Testing the predictive lichen growth-rate model

Introduction

It has been proposed, in the previous section, that the maximum growth rates of crustose yellow-green *Rhizocarpon* lichens can be predicted from climatic data (ie. *oceanicity* index = average precipitation / mean annual air temperature range). Using lichenometric data not utilised in the calibration of the lichen growth-rate/oceanicity analysis, it was decided to test the overall validity of the model.

A – Evidence from the Northeast Atlantic region

Table 9.3 below summarises data from five studies in the Northeast Atlantic region which have been used to test the reproducibility of the model. Each case is then discussed separately in detail.

Table 9.3 Testing the lichen growth-rate/oceanicity *model in the Northeast Atlantic region. Maximum* Rhizocarpon *diameters have been predicted based on temperature and precipitation data. Observed maximum lichen sizes are taken from five studies. The difference between the* predicted *and* observed *values are expressed as percentages in the final column.*

location (author of study)	latitude	maximum lichen diameter on 100-yr-old surface		error (%)
		prediction	observation	
Snowdonia, Wales (Winchester, *pers.comm.* 2000)	53°15'N	106	105	<1
Cumbria (east) (Harvey *et al.*, 1984)	54°30'N	70	80	-12
Cumbria (west) (Winchester, 1988)	54°30'N	86	90	-4
Kebnekaise, Sweden (Denton & Karlén, 1973)	65° N	28	35	-20
Jan Mayen (Anda *et al.*, 1985)	71° N	34	24	+42

1. Snowdonia, North Wales

Churchyards located within the extremely wet, mild climate of the Snowdonian mountains provide an excellent opportunity to examine the relationship between climate and lichen growth rates and serve as an ideal test for the growth rate/oceanicity model. Measurements of *Rhizocarpon geographicum* on slate gravestones at Capel Curig and Betws y Coed were made by Winchester in 1996. Maximum lichen growth factors for Capel Curig were found to be 105 mm/100 yr (Winchester, *pers. comm.* 2000). The model predicts a maximum lichen diameter of 106 mm on a 100-year-old surface. By recently re-measuring the same lichen thalli, Winchester has derived a directly-determined growth rate of 1.4 mm a^{-1} over a 4-yr period, confirming her earlier growth rate calculations (allowing for a colonisation time of 15-18 years) (Winchester, *pers. comm., 2000)*

2. Cumbria, NW England (Harvey et al., 1984)

Working in the Howgill Fells of Cumbria, northwest England, Harvey *et al.* (1984), dated recent fluvial deposits using lichenometry. They constructed a dating curve for the area using '*Rhizocarpon geographicum* agg.' lichens growing on gravestones in nearby churchyards. The use of an envelope curve, through the largest five lichen diameters, allows the *maximum* potential growth rate to be determined for a 100 year-old lichen thallus. A growth rate of 80 mm/100 yr has been derived from their

calibration curve (Harvey *et al.*, 1984: 358, Fig. 5). The upper Lune Valley experiences a mean annual temperature range of 11.5°C and mean annual precipitation of 1700 mm, according to data taken from maps published by the UK Meteorological Office (1931-1960). According to the model (Fig. 9.10), this corresponds to a predicted maximum lichen diameter of 70 mm on a 100-year-old surface. The predicted and measured maximum growth rates for the last 100 years are in close agreement, with the predicted value underestimating the 'true' value by only 10 %.

3. Cumbria, NW England (Winchester, 1988)

Measurements of *Rhizocarpon geographicum* subsp. *prospectans* on over 100 gravestones around Derwent Water and Grasmere in Cumbria, by Winchester (1988), yielded a maximum growth rate of 90 mm/100 yr. This study, although only 40 km from the study site of Harvey *et al.* (1984), produced a notably faster growth rate for lichens of the same family growing on the same rocktype. The difference in growth rate can be explained by the precipitation gradient between the two sites. The predicted maximum lichen size on a 100-year-old substrate in Grasmere, in the English Lake District, is 86 mm. This is in very close agreement with the figure derived by Winchester (1988). According to the model, the highest localised lichen growth rates in the British Isles – greater than 100 mm per 100 yr – may be found at the heads of the glacially-eroded valleys that dissect the central Cumbrian massif (ie. Borrowdale, Wasdale and Langdale) and correspond with the areas of highest precipitation (>3000 mm/yr).

4. Kebnekaise, Sweden

Situated 500 km from the Norwegian coast and at an altitude of approximately 1000 m, the Kebnekaise mountains provide a more continental climate in which to test the lichen growth/oceanicity model. The mean annual temperature range at Riksgransen is 22°C (1931-1960) and the annual precipitation over the same period was 940 mm. Observations at Tarfala mountain research station only commenced in 1967, although the data suggests that precipitation totals are very similar to those at Riksgranen (940 mm cf. 990 mm (1967-1980)). The model predicts a largest lichen of 28 mm on a 100-year-old surface in such a climate. Denton and Karlen's (1973)

lichen-dating studies were carefully calibrated using a host of natural and man-made surfaces in the Kebnekaise area. Using *Rhizocarpon geographicum* and *Rhizocarpon alpicola* he derived a growth rate of 35 mm/100 yr (read directly from Fig.14; 1973: 369) for the Kebnekaise mountains. The model prediction underestimates the observed value by 23%. This discrepancy is greater than has been observed at lower latitudes in maritime climates. This result shows that the exact role of various environmental factors (eg. prolonged snowcover, low winter light levels, increased UV levels, increased evapotranspiration etc.) on lichen growth at high altitudes and in arid, high latitude climates need to be further examined.

5. Jan Mayen

Moraine-dating studies on the glaciated volcanic island of Jan Mayen (Anda *et al.*, 1985) have derived growth rates of 24 mm/100 years for *Rhizocarpon* lichens in this sub-polar climate (read directly from the dating curve (1984:Fig.12)). Meteorological records from Jan Mayen extend back to 1921, although some monthly data during the period 1940-42 are missing. Lichens were sampled on moraines in the north of the island around the present-day ice-cap Beerenberg. Data from the nearest weather station is assumed to be representative of the environment in which the lichens were measured. The typical mean annual temperature range (1931-1960) on Jan Mayen is 11.1 °C and the mean annual precipitation (1931-1960) 646 mm. Such a climate should yield a maximum lichen size of 34 mm after 100 years, according to the growth rate-oceanicity model. However, this is considerably (c. 40%) higher than that recorded by Anda *et al.* (1985).

This discrepancy is interesting. But there are several possible reasons for the apparent difference between the maximum lichen diameter predicted for Jan Mayen and that recorded by Anda *et al.* (1985). The Danish-Norwegian lichenometric study relies heavily on the age of certain moraines believed to date from the middle to late 19th century. Previous scientific expeditions to Jan Mayen found the glaciers to be at various stages of retreat in 1882, and 1938 (von Boldva, 1886; Jennings, 1939). Prior to 1882 the positions of the glacier termini can only be estimated. The dating control on Jan Mayen is therefore only weakly constrained and may lead to uncertainties surrounding the exact age of the 19th century moraines. Thorarinsson (quoted in Jennings, 1951) states that some of these moraines may well date from a

previous glacier advance, such as the one identified in Norway which culminated in the mid to late 18th century. Lichen growth rates derived from surfaces of disputed age should be seen as, at best, only estimates.

A second possible reason for the discrepancy between the predicted and the measured lichen sizes is due to climatic change. Lichen growth rates are related to climate, as has been highlighted already. Climatic changes on a decade to century scale will consequently affect the growth rates of slow-growing crustose lichens. Significant changes in precipitation and temperature regime over the past 100 years will result in temporally-variable growth rates, which cannot be taken into account by such a relatively crude model.

The Polar regions are known to experience greater amplitudes of climatic change relative to middle and low latitudes. A shift towards a warmer, wetter, more oceanic climate over the past 100 years has occurred throughout the Atlantic-Arctic sector (Houghton *et al.*, 2000). Jan Mayen in particular has experienced a most remarkable climatic amelioration since records began in 1921 (Lamb *et al.*, 1963; Brazdil, 1988). Mean air temperatures at Jan Mayen rose rapidly over the period from 1921 to 1940 and have continued to rise steadily up to 1993. Correlation with temperatures at Grimsey, north of Iceland, suggests that the current mean temperature (1971-2000) represents an estimated increase of around 4°C since the 1890s. This 20th century warming has been accompanied by a very substantial increase in precipitation, probably as a result of the increased vigour of the general atmospheric circulation over the North Atlantic and relative absence of sea ice in the area since the late 1920s (Lamb *et al.*, 1963; Steffensen, 1982). The exact magnitude of this precipitation increase was c. 80% over the period 1920 to 1960. Mean annual precipitation for the period 1922-1931 was only 410 mm whereas for the period 1950-1959 this figure was 740 mm. Precipitation levels have remained high since the 1950s, typically 700 mm a year, with only a small deviation below average during the cold spell from 1964-1970. The earliest measurements of precipitation on Jan Mayen (1920s) are thought to be representative of the first two decades of the 20th century generally. Data from other North Atlantic sites (eg. Bodø, Grimsey, Teigarhorn and Torshavn) indicate a distinct rainfall minimum in the years 1900-1920. Increased temperatures over the last 60 to 70 years in the vicinity of Jan Mayen also resulted in a larger proportion of precipitation falling as rain. The warmer, rainier climate produces a

notably higher oceanicity index (*K-index*) for Jan Mayen over the period 1931-1960 than for the preceding 30-year period. The relatively mild, wetter climate of the island over the last ~70 years would have resulted in accelerated lichen growth rates. The 20th-century climatic variation described here may partly explain why the predicted values of maximum lichen size, which assume uniform climatic conditions over the 100-year period – but are based on data from the 'climatic optimum' of 1931-1960 – are in excess of those actually observed on the island.

A third reason for the inaccuracy of the prediction may be due to a flaw in the model. The model is calibrated using only one high latitude (>70°N) site and therefore may not truly represent lichen growth conditions in Polar climates. The simple model cannot predict lichen diameters lower than c.10 mm due to the nature of the regression analysis. A more robust model would need more calibration points to determine the exact relationship between climate and lichen growth rates in more arid, polar regions.

In the highest latitudes and at high altitude where most of the precipitation falls as snow the model probably fails to predict lichen growth rates accurately. The role of snow in affecting lichen growth is still undetermined. Light levels and moisture levels are reduced beneath a thick snow covering. These factors probably restrict photosynthesis during spring and autumn whilst serving to protect the thallus from the extreme, freezing temperatures during winter. The period of daylight, snow-free thallus moisture, seen by Benedict (1990a) as crucial for growth, is severely restricted at high latitude and high altitude sites. Hence, snow may play a more crucial role in determining growth rates in polar climates than at lower latitudes. A future, more powerful, predictive lichen growth-rate model may need to take the possible effects of prolonged snow cover into account.

B – Evidence from further afield

Lichenometric research has been conducted in various parts of the world, usually as part of geo-chronological studies into recent landscape change. It is very tempting to examine the applicability of the lichen growth-rate/oceanicity model ($y = 0.4x + 11.0$), calibrated in northwestern Europe, in more distant regions.

Table 9.4 Selected tests of the lichen growth-rate/oceanicity *model from outside NW Europe. The mean annual temperature range and mean annual precipitation at the nearest representative weather station are also shown.* Predicted *lichen growth rates are expressed as the maximum size of* Rhizocarpon *thalli growing on a 100-yr-old surface.* Observed *values are taken directly from the various studies.*

location (author of study)	temp. range (°C)	average precip^{n}. (mm/yr)	maximum lichen diameter on 100-yr-old surface: prediction	observation	error (%)
Front Range, Colorado, USA *(Benedict, 1967)*	21	105	13	14	-7
Northern Cascades, Washington, USA *(Miller, 1969)*	17	2300	65	60	+8
Wrangell Mountains, S. Alaska *(Denton & Karlen, 1973)*	21	300	17	17	<1
Southern Alps, New Zealand *(Burrows and Lucas, 1967)*	14	2800	91	76	+20
South Patagonian Icefield, Chile *(Winchester & Harrison, 1984)*	15	1600	54	55	<1

Summary

It is clear that there is a good match between observed and predicted values of lichen growth rates using the model (Tables 9.3 and 9.4). In the northeast Atlantic region four out of five predicted values of lichen diameter are accurate to within +/-20%. The three lower-latitude sites show the best agreement with the predictions, whilst the prediction for Kebnekaise in Arctic Sweden is less accurate – probably as a result of the cooler, more continental climate. The prediction for Jan Mayen over-estimates the measured lichen growth rate by 42%. The reasons for this discrepancy could be attributed to the remarkable degree of climatic change experienced in the region during the last century and the intrinsic weakness of the model in more arid, polar climates. Further preliminary tests of the model in similar maritime, and more continental, climates from both hemispheres show a good agreement between the predicted and observed maximum size of lichens growing on 100 year-old-surfaces.

Judging by the success of the lichen growth-rate/oceanicity model, there is no reason why the relationship should not be used to estimate maximum yellow-green *Rhizocarpon* growth rates in temperate and sub-polar climates where 'growth' curves cannot be established due to a lack of calibration surfaces. This may allow the age

estimation of surfaces in more remote and isolated areas, without the need to construct local lichen-dating curves.

Conclusions – environmental controls on lichen growth rates

1. Indirectly-determined growth rates of yellow-green *Rhizocarpon* lichens – expressed as mm/100 yr – vary considerably across the North Atlantic region (>100 mm to <10 mm/100 yr).

2. These spatial variations are attributable to differences in climate and are particularly related to variations in annual precipitation, the number of rain days, air temperature and annual temperature range, experienced in different geographical regions.

3. The degree of climatic *oceanicity* (expressed as mean annual precipitation / mean annual temperature range (1931-1960)) accounts for 95% of lichen growth factor variation throughout Northwest Europe and Greenland over the last 100 years.

4. Using the results of the regression analysis, maximum *Rhizocarpon* sizes on 100-year-old surfaces have been predicted for areas where no lichenometric work has previously been undertaken. The climate of southernmost Iceland, the Faeroe Islands, coastal southwest Norway, upland areas of western Ireland, the Highlands and mountainous islands of western Scotland, along with much of Gwynedd and Cumbria, in Great Britain are theoretically most conducive to rapid lichen growth. Where *Rhizocarpon* lichens grow in these localities, maximum growth rates – based on optimum 20th-century climatic conditions – are predicted to be in excess of 90 mm per century.

5. These predictions need verifying. However, due to the lack of existing data on lichen growth rates this model operates at only a relatively low spatial resolution. Local climatic variations due to topography (which affects air temperature, precipitation, windiness, etc.) are likely to result in significantly higher or lower growth rates being observed than the 'typical' values predicted here.

Implications – effects of climatic change on lichen growth rates

Lichen growth rates are strongly linked to the oceanicity of the climate in Northwest Europe and Greenland. Lichen growth rates in different climatic zones, such as northeast and southeast Iceland, can vary greatly – primarily as a result of the available moisture. Climatic conditions today, in the year 2001, are different from those 100 years ago. Comparisons of oceanicity indices over time highlight an increasingly maritime component in Iceland's climate, since measurements began in 1845 (Table 9.5). By extension, it is logical to assume that this climatic change would have affected the relative speed at which lichens, and other plants, grow. Present-day lichen growth rates – in southeast Iceland, for example – should be higher than they were in the late 19th century.

Climatic data suggests that lichen growth rates in Iceland were probably most rapid during the period 1930-1960 than at any time during the last 150 years. This should be remembered when applying the growth-rate/oceanicity model (Fig. 9.11). The model will tend to overestimate maximum lichen growth rates, by c. 10% across the Northeast Atlantic, because it is calibrated using data from the period most favourable for optimum lichen growth (ie. AD 1931-1960).

Table 9.5 Comparison of oceanicity indices at four Icelandic weather stations over the period 1846-1990. A mean annual temperature range of 10°C and mean annual precipitation of 500 mm will yield an index of 50. Note the marked 20% increase in oceanicity from 1871 to 1960 at Stykkisholmur and Teigarhorn and from 1901 to 1960 at Suðureyri.

	1846-1870	*1871-1900*	*1901-1930*	*1931-1960*	*1961-1990*
Stykkisholmur	44	46	50	57	55
Teigarhorn		102	120	119	114
Suðureyri			76	93	89
Hólar			114		109

On the balance of evidence, it is highly likely that environmental change over the last two centuries in Iceland has accounted for secular variations in lichen growth rates. Lichen colonisation rates are also probably influenced by environmental conditions. A high frequency of cold, northerly winds depleted in lichen spores, coupled with the effect of increased snowcover and a shortened growing season, probably played a

vital role in suppressing lichen colonisation rates during the cooler periods of the last 200 years. Climatic conditions during the 20th century have been notably more favourable for lichen growth with a shift to warmer, wetter, more maritime conditions following the end of the *Little Ice Age* (Jonsson and Garðarsson, 2001). However, without careful, long-term, direct measurements of lichen growth in Iceland the effect of climatic change on growth rates cannot be firmly established.

Summary of Chapter 9

Data regarding the micro-and macro-environmental conditions for optimum lichen growth have been collected from Iceland. Generally, yellow-green *Rhizocarpon* lichens prefer south-facing, well lit, well ventilated, stable micro-habitats in maritime southeast Iceland. Prolonged snowcover suppresses lichen colonisation in upland areas. Growth rates of *Rhizocarpon* Section *Rhizocarpon,* indirectly-derived from gravestones, indicate that the 'largest-lichen' growing on a 100-yr-old surface in southeast Iceland will be ~40% larger than in Northwest Iceland. This growth rate difference is almost certainly a function of the warmer, wetter, more maritime climate in the southeast.

The relationship between lichen growth rates and climatic oceanicity in Northwest Europe, suggested by Beschel in 1961, is described and tested here for the first time. This relationship allows growth rates of yellow-green *Rhizocarpon* lichens across the Northeast Atlantic and further a-field to be estimated, using merely readily available climatic data. Where circumstances do not permit the construction of lichenometric dating curves (which are always preferable), a lichenometric-age estimate could be made based purely on the lichen growth-rate/oceanicity model

Finally, as lichen growth rates are related to climate, the relationship between secular climatic change and lichen growth cannot be overlooked. Although only partly substantiated, such a phenomenon casts doubt on all lichenometric-dating studies that use a constant, uncalibrated, growth rate over many centuries and those that extrapolate present-day growth rates to determine the age of ancient lichen-bearing surfaces.

Chapter 10 : Glacier Fluctuations and Holocene Climatic Change

Introduction

This chapter reconstructs the temporal and spatial pattern of glacier fluctuations in the study area, based on the findings of geomorphological, tephrochronological and lichenometrical investigations. The second part of this chapter deals with the climatic implications of the reconstructed glacier fluctuations, with particular emphasis on the last 1000 years.

Three timescales are considered:

- Sub-decadal-scale changes – the NAO operates on this timescale.
- Decadal-scale changes – for example, the warming during the 1920s and 1930s and the cooling in the 1960s.
- Centennial-scale changes – most notably the *Little Ice Age* and *Medieval Warm Period.*

Reconstructed glacier fluctuations

Using moraine limits, tephrochronology, combined with knowledge of soil accumulation rates, and lichenometry it has been possible to reconstruct glacier fluctuations over the last ~10,000 years, as has been done successfully elsewhere in Iceland (eg. Thorarinsson, 1956, 1960; Dugmore, 1987; Stötter, 1990; Häberle, 1991; Gudmundsson, 1998a).

Figures 6.16 and 6.17 (in Chapter 6) show all the moraines identified in the field area. Most of these moraines relate to the former extent of Lambatungnajökull, but the extent of other palaeo-glaciers in the Hornafjörður region have also been delimited.

Age estimates for each feature, based on the presence of tephra layers in the surrounding soil are summarised in Chapter 7 (Tables 7.1 and 7.2). Based on these bracketing dates, the extent of Lambatungnajökull has been reconstructed for several periods during the last ~10,000 years (Figs. 10.1 – 10.4).

Table 10.1 Holocene palaeo-geographical reconstructions:

stage	dates	map
Dalsheiði	> 9000 BP	Figure 10.1
Hoffellsdalur I	>4800 <6000 BP	Figure 10.2
Hoffellsdalur II	>1800 <3500 BP	
Hoffellsdalur III	>1200 <1800 BP	Figure 10.3
Hoffellsdalur IV	~ AD 1780s – *Little Ice Age*	Figure 10.4

Moraine age estimates based on lichenometry are summarised in Chapter 8 (Tables 8.4 – 8.7). Based on these bracketing dates, the extent of Lambatungnajökull has also been reconstructed for several periods during the last 500 years (Figs. 10.5 – 8.10).

Table 10.2 Little Ice Age *palaeo-geographical reconstructions:*

stage	culmination of advance	map
Hoffellsdalur IV	~AD 1780s – *Little Ice Age* Maximum	Figure 10.5
Hoffellsdalur V	AD 1810s	Figure 10.6
Skyndidalur IV	AD 1850s	Figure 10.7
Skyndidalur VI	AD 1890s	Figure 10.8
Skyndidalur VIII	c. AD 1930	Figure 10.9
Skyndidalur IX	c. AD 1965	Figure 10.10

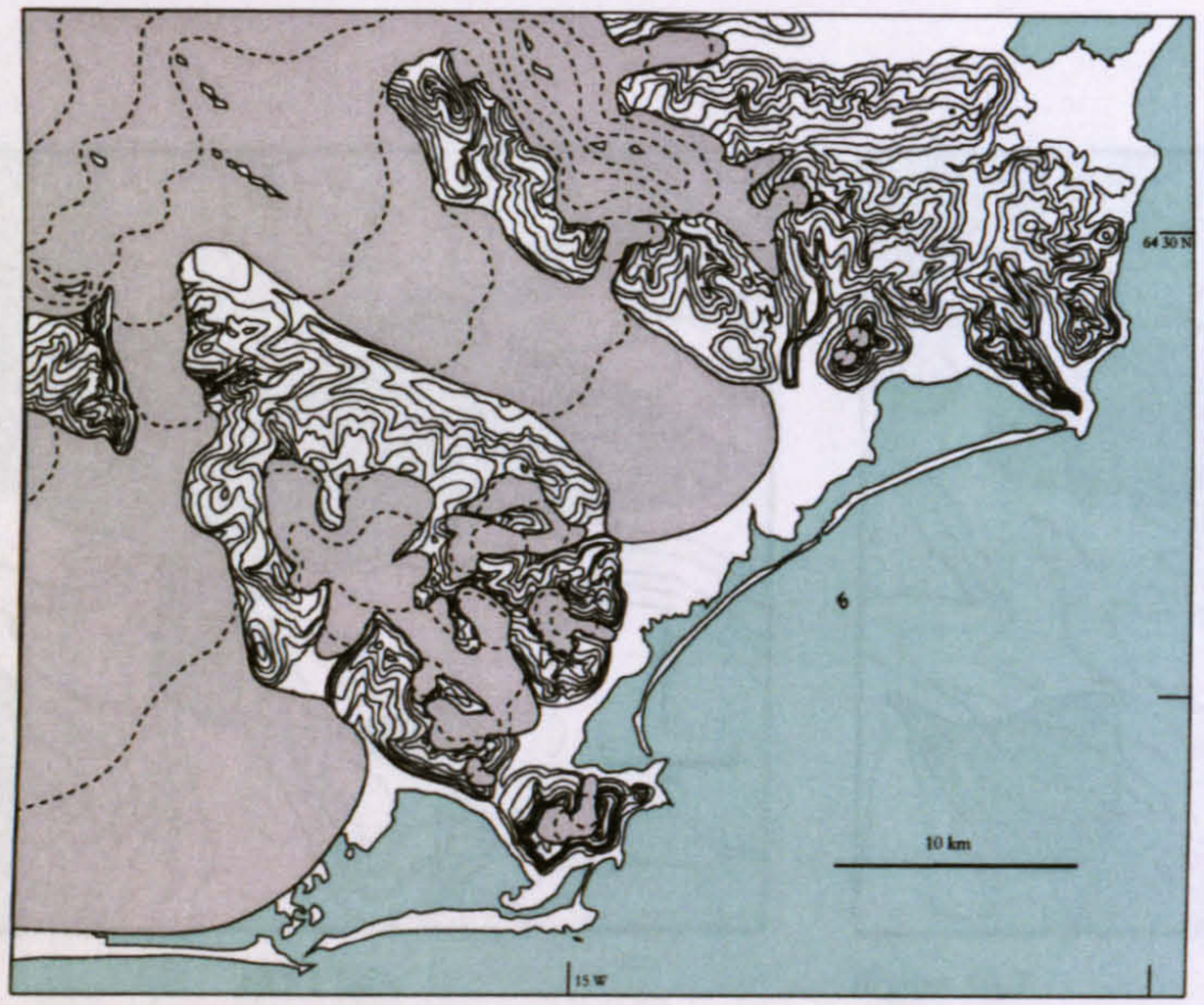


Figure 10.1 > 9000 BP

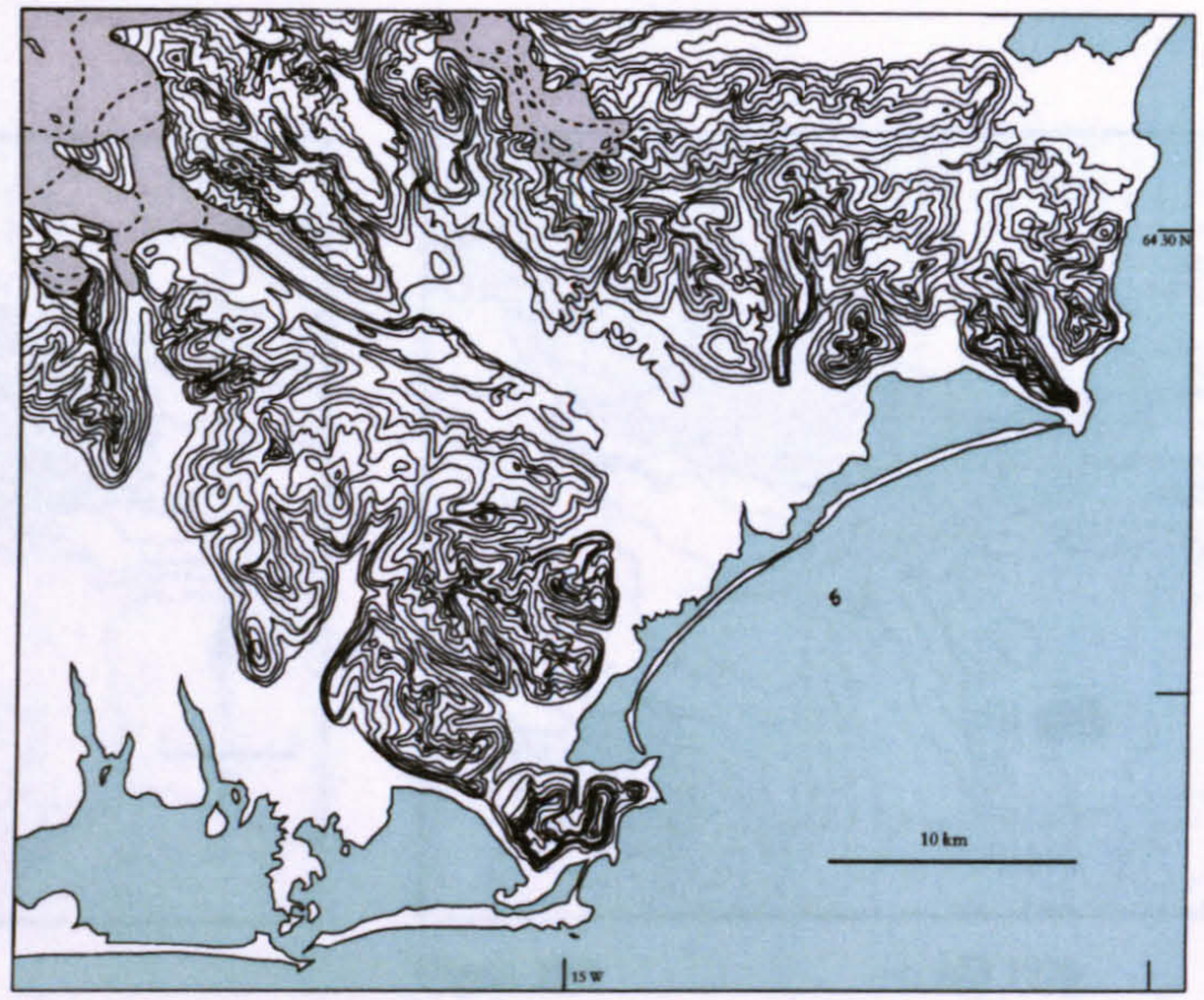


Figure 10.3 c. 1600 BP

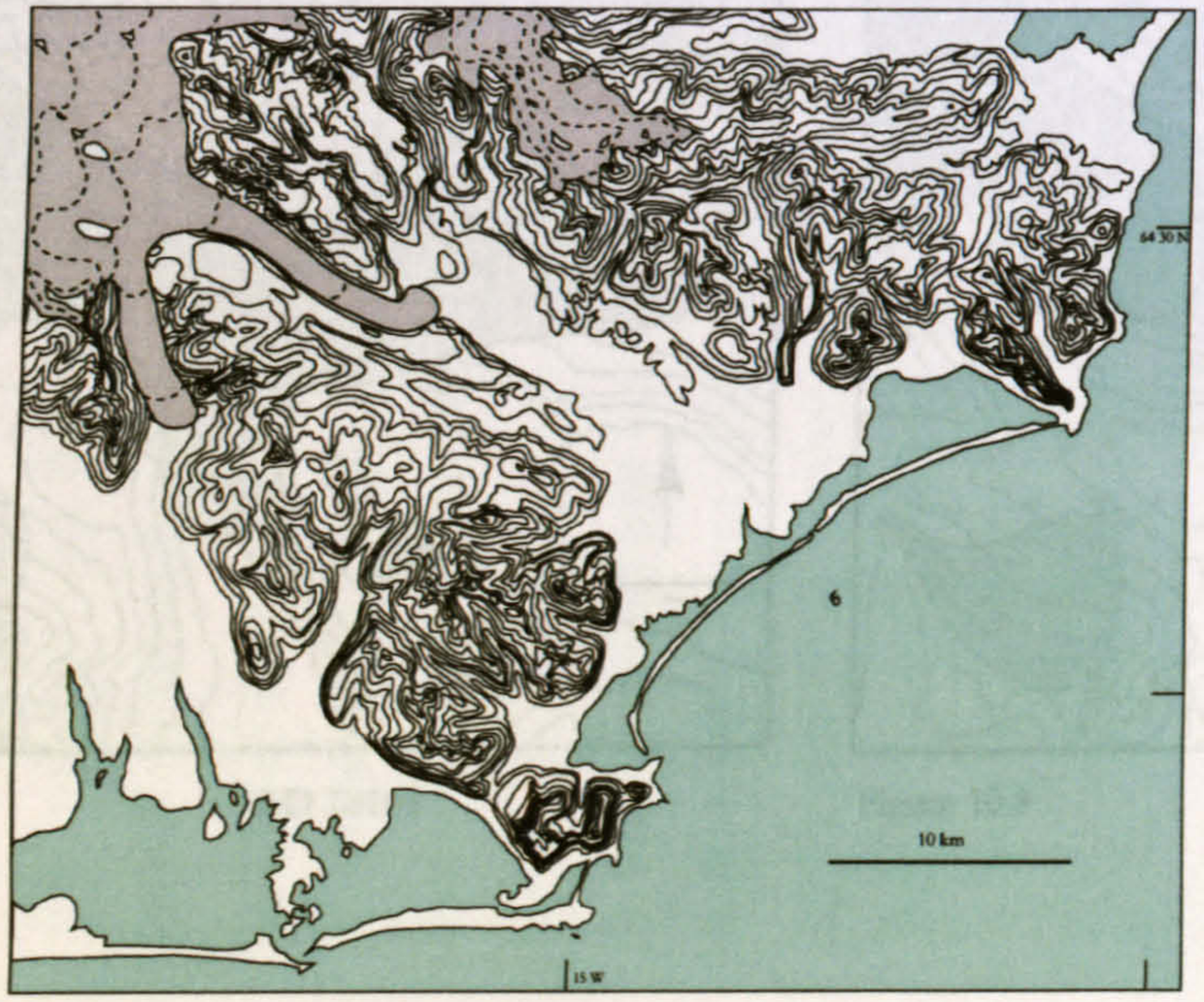


Figure 10.2 c. 5000 BP

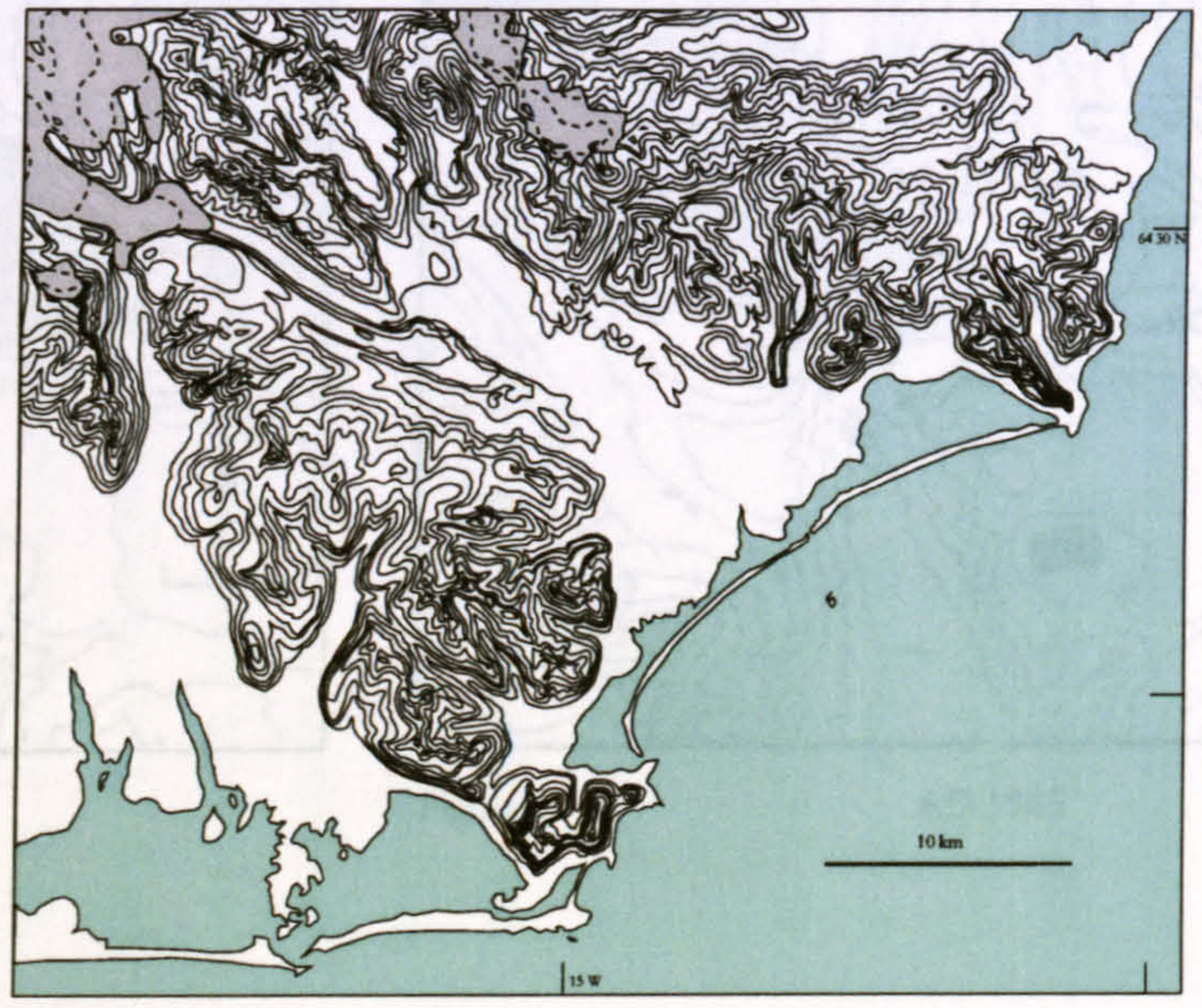


Figure 10.4 c. AD 1780s

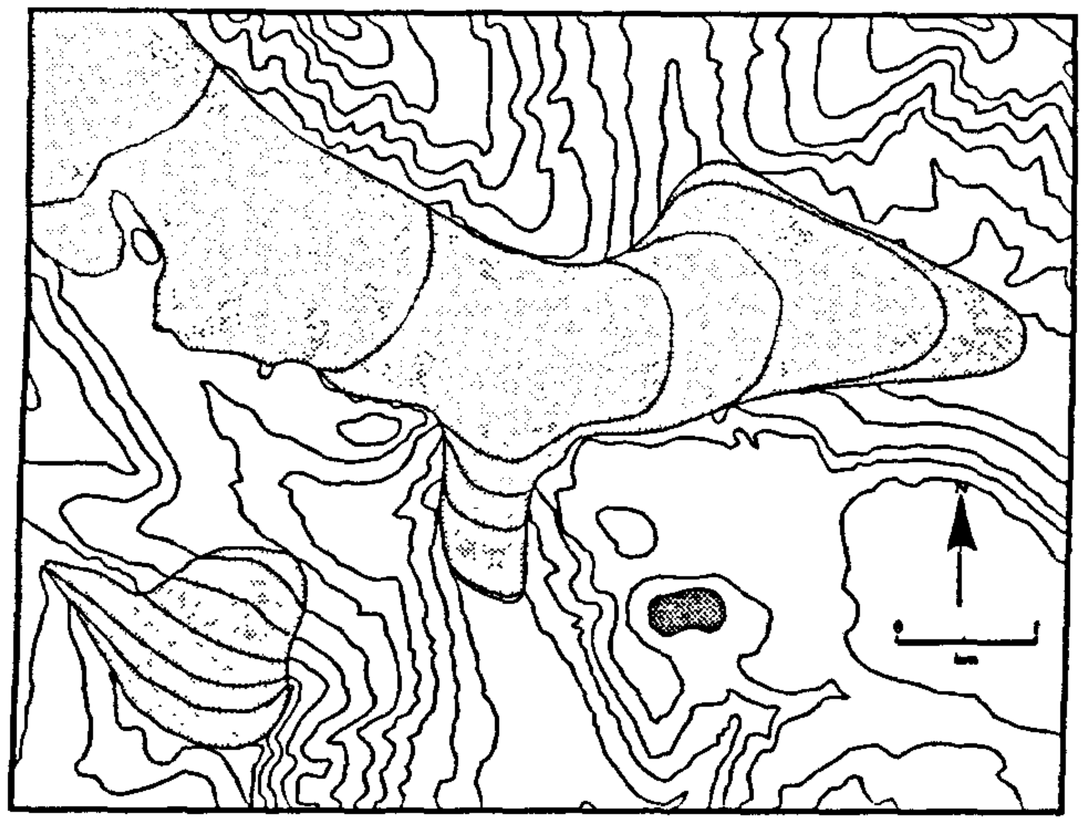


Figure 10.5 AD 1780s

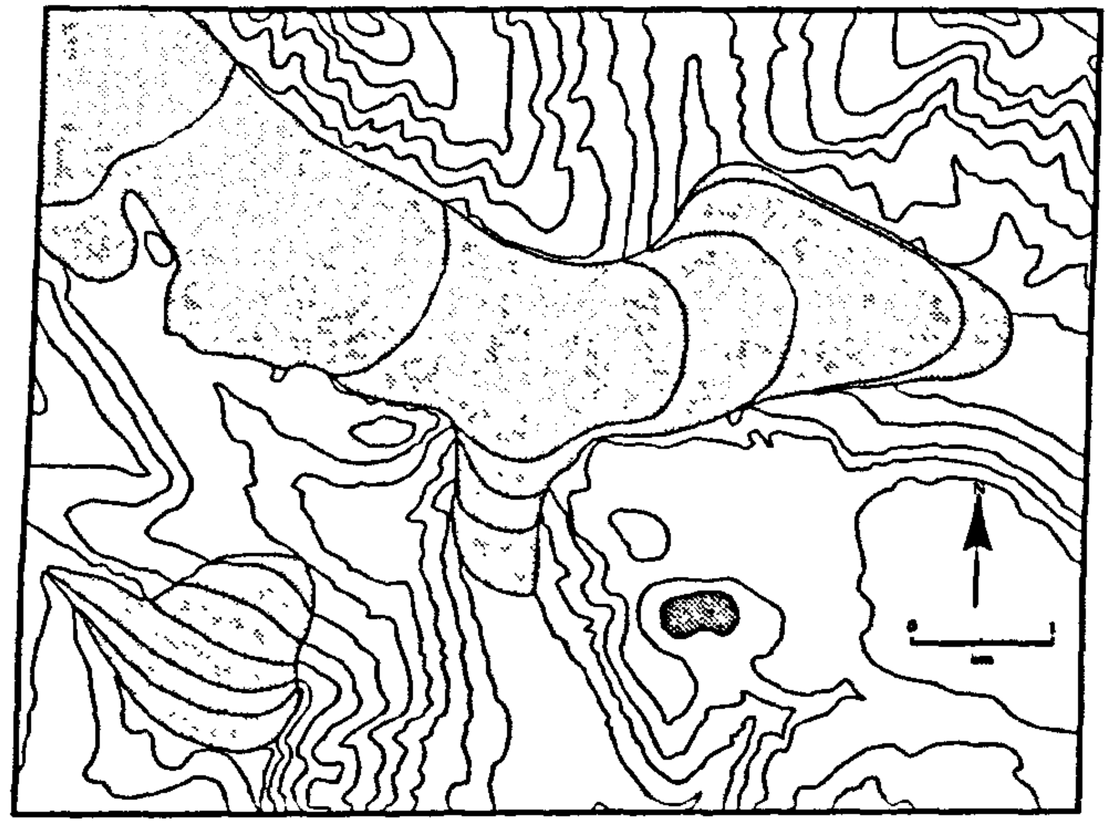


Figure 10.6 AD 1810s

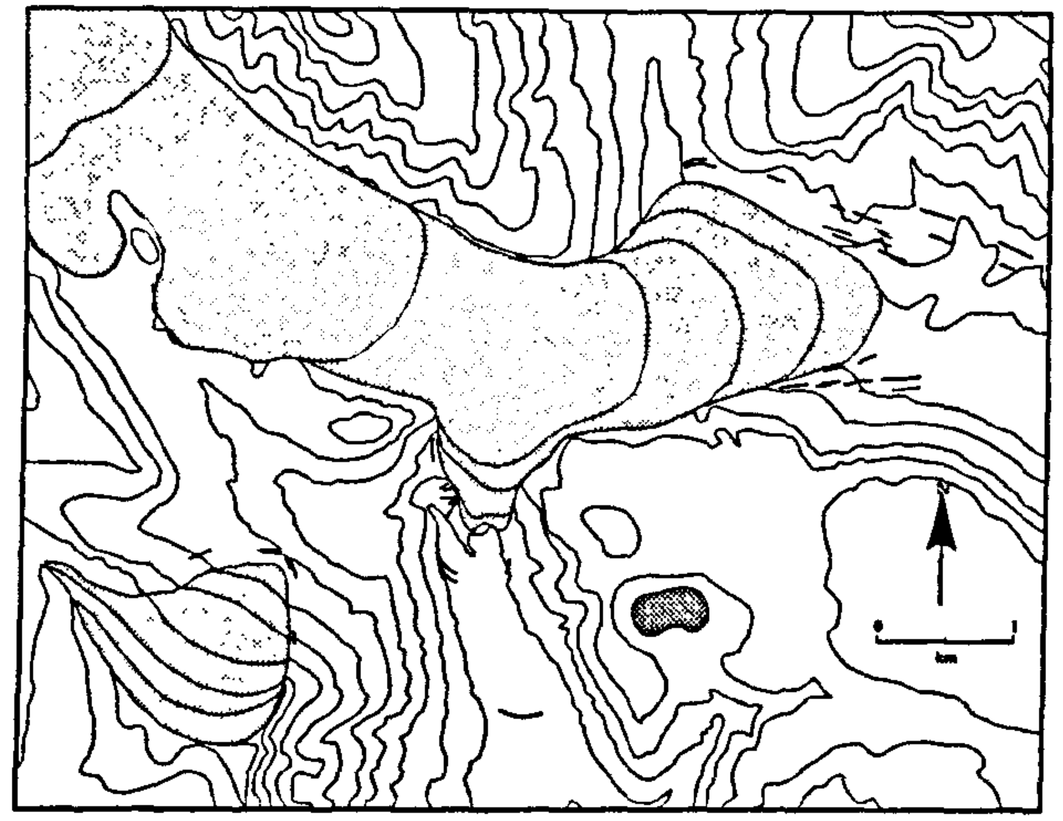


Figure 10.7 AD 1850s

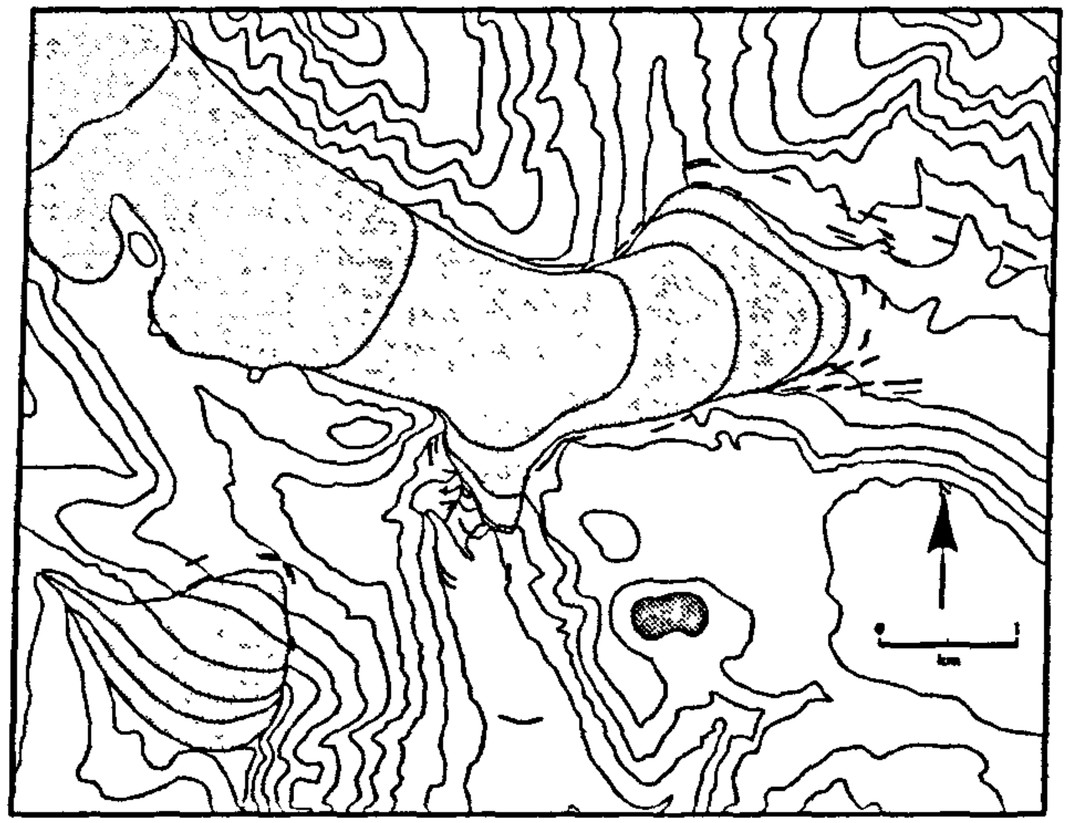


Figure 10.8 c. AD 1890

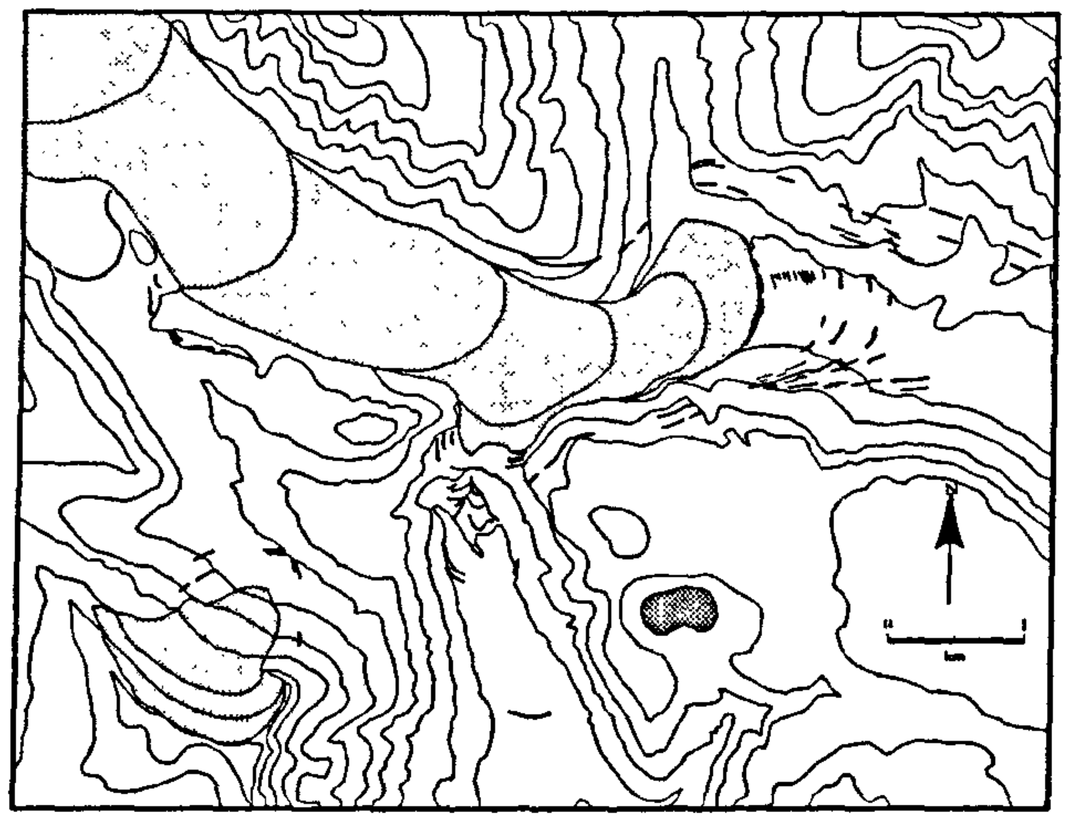


Figure 10.9 c. AD 1930

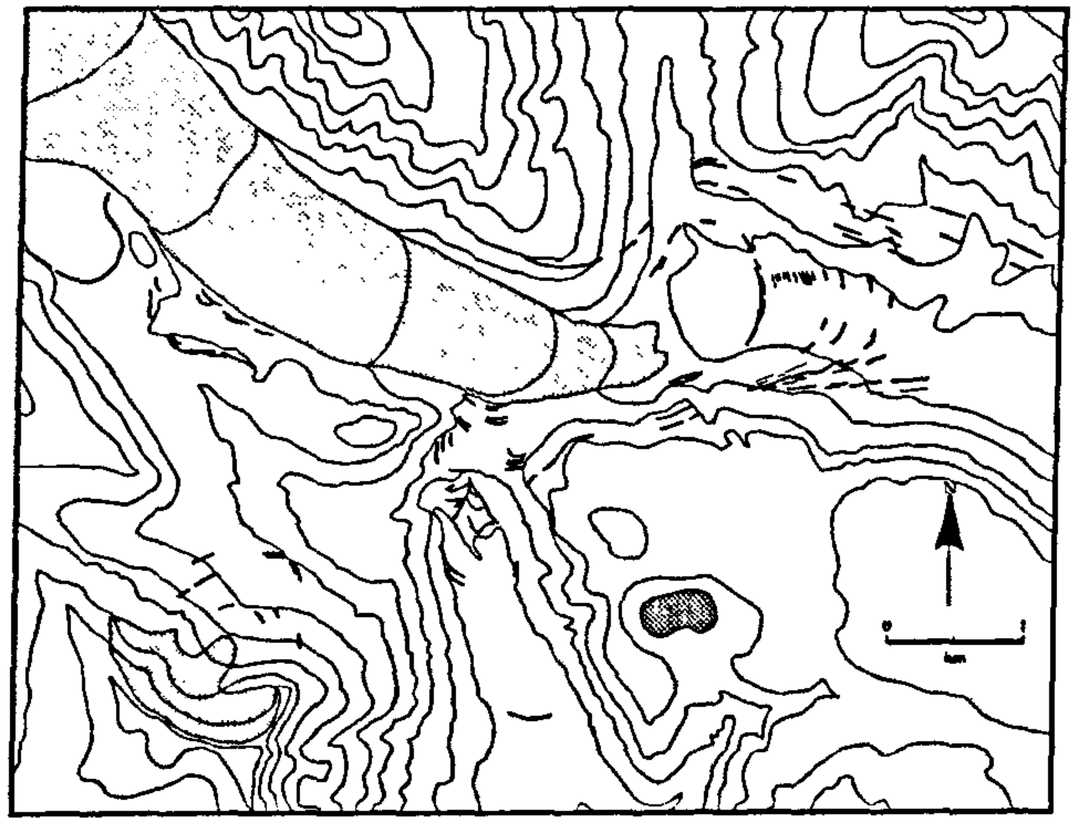


Figure 10.10 AD 1965

It can be seen from the reconstructions of ice-extent in the valleys of Skyndidalur and Hoffellsdalur that Lambatungnajökull has fluctuated greatly since the termination of the Last Ice Age. Four main, discrete, periods of glacier advance or stillstand have been identified in the last ~10,000 years: one at least 9000 years BP, one between ~4800 and ~6000 years BP; two between ~1000 and ~3500 years BP, and one during recent centuries c. AD 1750-1800. The magnitude of these events has become smaller with time.

Over the last ~300 years the margin of Lambatungnajökull has continued to fluctuate. Lichenometric dates yield the minimum age of surface exposure; ie. when the ice retreated from the proximal flank of the moraine. It is assumed that the culmination of the glacier advance responsible for forming the ridge pre-dates this time by a few, probably no more than 10, years. On this basis several periods of re-advance or stillstand have been identified all of which culminated shortly before the following times:- 1795, 1815, 1860, 1875, 1895, 1923, 1930, 1967 and 1982 (Table 8.8).

The timing of glacier fluctuations in Iceland : a discussion

The last 10,000 years

The temporal pattern of glacier fluctuations derived in this research shows a similar pattern to those derived by scientists working in other parts of Iceland (Dugmore, 1987, 1989; Gudmundsson, 1998a; Stötter, 1990; Häberle, 1991) (Fig. 10.11). These new findings indicate that Vatnajökull was more extensive during the mid-Holocene than at any time in the last 1000 years.

Earlier in the 20th century it was a commonly held belief that the recently-deposited moraines in Iceland – dating from the 18th and 19th centuries – probably represented the maximum extension of glaciers in Postglacial time (ie. since 10,000 BP) (Eythorsson, 1935; Thorarinsson, 1936; 1944). This view was modified when Thorarinsson (1956) found moraines of pre-historic age outside those formed during the *Little Ice Age*. Thorarinsson concluded that some of the glaciers of Vatnajökull were a little more

advanced in Sub-Atlantic times (c. 2500 BP) than in recent centuries. This view held for several decades. Regarding the last 10,000 years, Björnsson (1979) argued that deglaciation was rapid soon after the end of the Last Ice Age. Björnsson also proposed that Iceland was largely ice-free during the first half of the Holocene, the period known as the Climatic Optimum or *Hypsithemal*, and that the major ice-caps reformed only 2-3000 years ago.

This new research, along with studies by Dugmore (1987) and Gudmundsson (1998a), show that this is unlikely to have been the case. Dugmore's (1987, 1989) results were the first to indicate that the ice-caps of Mýrdalsjökull and Eyjafjallajökull were more extensive during the mid-Holocene than at any time since. A period of *Neoglaciation* culminated between 7000 and 4500 years BP at Sólheimajökull. Gudmundsson's (1998a) study of glacier fluctuations in Öræfi, southeast Iceland, supports Dugmore's findings. Gudmundsson found terminal and lateral moraine fragments confirming that the glaciers of Öræfajökull were more extensive during the mid-Holocene than in *Little Ice Age* or Sub-Atlantic times. Both Dugmore (1987) and Gudmundsson (1998a) found other, pre-historic, moraines dating to c.3000 BP and c.1200-1800 BP which were also the product of pre-*Little Ice Age* glacier advances.

The glacier fluctuations identified by both Dugmore (1987) and Gudmundsson (1998a) are difficult to interpret climatically, as they may be complicated by other factors. Dugmore (1989) points to changes in the catchment area of Sólheimajökull, as a possible reason for the apparently anomalous behaviour of the glacier. Migration of the ice-divide over time could have caused the glacier to fluctuate in response to topographically controlled thresholds, rather than simply changes in atmospheric conditions (Dugmore and Sugden, 1991). However, new research by Mackintosh (2000) suggests that the role of topography and catchment dynamics may have been overstated. Using a numerical model driven by the GRIP temperature record, Mackintosh (2000) suggests that a temperature change of merely -2°C (from the 1966-1996 mean) is required to successfully simulate the frontal position of Sólheimajökull in the mid-Holocene.

Table 10.3 Glacier advances of Lambatungnajökull discovered in this study.

Stage	*Bracketing dates (BP) or minimum age (AD)*	*Culmination of advance*	*Period*
Dalsheiði	> 9000 BP	> 9000 BP	Pre-Boreal / Younger Dryas
Hoffellsdalur I	4800-6000 BP	c. 5000 BP	Atlantic *Neoglaciation*
Hoffellsdalur II	1800-3500 BP	c. 3000 BP	Sub-Boreal
Hoffellsdalur III	1200-1800 BP	c. 1600 BP	Sub-Atlantic
Hoffellsdalur IV *Skyndidalur III*	<AD 1796	1780s	*Little Ice Age* maximum
Hoffellsdalur V	<AD 1816	1810s	LIA
Hoffellsdalur VI *Skyndidalur IV*	<AD 1855-1865	1850s	LIA
Hoffellsdalur VII *Skyndidalur V*	<AD 1880	c. 1870	LIA
Hoffellsdalur VIII *Skyndidalur VI*	<AD 1895-1900	c. 1890	LIA
Skyndidalur VII	<AD 1923	1920	20th Century - I
Skyndidalur VIII	<AD 1930-1933	1928	20th C - II
Skyndidalur IX	AD 1960-1967	1965	20th C - III
Skyndidalur X	<AD 1982	1980	20th C - IV

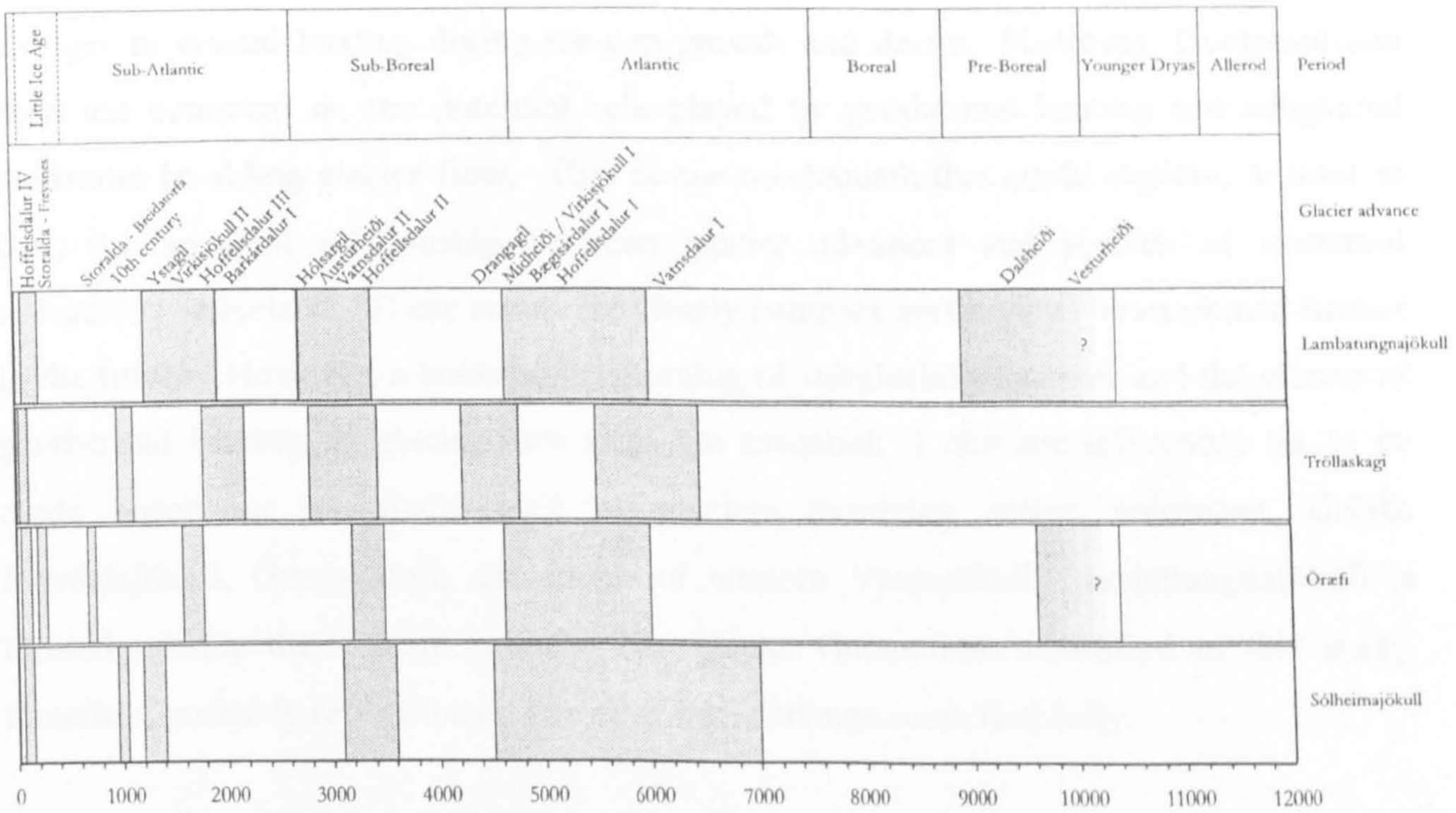


Figure 10.11 A comparison of Holocene glacier fluctuations in Iceland. The diagram summarises the evidence gathered from four regions:
Sólheimajökull - Dugmore (1987)
The glaciers in the Öræfi district - Gudmundsson (1998a)
The glaciers in the Tröllaskagi highlands - Stötter (1990), Häberle (1991)
Lambatungnajökull - Bradwell (this study)

Further uncertainties surround the role of volcanism. Mýrdalsjökull, like Eyjafjallajökull, Öræfajökull, western Vatnajökull, and several small mountain ice-caps in Iceland, overlie active volcanoes. Volcano-glacier interactions could result in a non-climatic record of glacier fluctuations. Few studies have focused on the role of volcanic activity on glacier terminus fluctuations. Thorarinsson (1943) was aware that the marginal fluctuations of western Vatnajökull were greatly affected by subglacial volcanic activity. Sturm *et al.*, (1991) working on Mt. Washington in North America, found that the strong geothermal heat gradient in the underlying bedrock contributed significantly to the production of subglacial meltwater. This basal water acts as a lubricant allowing the glacier to slide more effectively over its bed. Sturm *et al.* (1991) argued that glaciers flowing over active volcanic systems would fluctuate in response to such changes in basal water conditions as well as changes in atmospheric (climatic) factors. Further complications could arise if the shape or size of the glacier catchment changed as the result of a volcanic explosion.

Gudmundsson (1998a) finds a close relationship between the timing of glacier fluctuations and eruptions of the Öræfajökull strato-volcano. He associates this with changes in crustal loading during ice-cap growth and decay. However, Gudmundsson does not comment on the potential role played by geothermal heating and subglacial meltwater in aiding glacier flow. This is one mechanism that could explain, at least in part, the apparent relationship between glacier advances and periods of increased volcanicity in Iceland. These issues are clearly complex and need to be examined further in the future. However, a better understanding of subglacial processes and the effects of geothermal heating on glacier flow rates are essential, if climatic inferences are to be made based on the fluctuations of glaciers overlying active volcanoes. Unlike Mýrdalsjökull, Öræfajökull, and much of western Vatnajökull, Lambatungnajökull is located outside the volcanic zone. The glacier fluctuations identified in this study, therefore, probably reflect the course of climatic change more faithfully.

The last 1000 years

Several studies have focused on glacier fluctuations during the last 1000 years. Some workers have studied the small alpine glaciers of northern Iceland (Caseldine, 1990;

Häberle, 1991; Kugelmann, 1991; Stötter *et al.*, 1999), whilst others have studied the large outlet glaciers in southern Iceland (Thorarinsson, 1956; Dugmore, 1987; Thompson, 1988; Ives, 1996; Gudmundsson, 1998a; Evans *et al.*, 1999). Most have concentrated on the last few centuries – a period known as the *Little Ice Age.*

According to recent studies, where lichenometry has been used to date moraines, the maximum extent of glaciers in Iceland during the *Little Ice Age* was reached between AD ~1850 and 1900 (Gordon and Sharp, 1983; Thompson, 1988; Caseldine, 1990; Kugelmann, 1991; Gudmundsson, 1998a; Evans *et al.*, 1999). Exceptions are found at Tungnahryggsjökull in Tröllaskagi, Svinafellsjökull in Öræfi, and Gigjökull and Steinholtsjökull flowing from the Eyjafjöll volcano. These glaciers were most extensive in AD 1810-1820 (Caseldine, 1990), AD 1727-1755 (Gudmundsson, 1998a) and AD 1721-1821 (Kirkbride and Dugmore, 2001b), respectively. However, uncertainty surrounds the age of moraines in Tröllaskagi beyond the range of the lichen calibration curve (Häberle, 1991), whilst questions surround the climatic significance of glacier fluctuations on active volcanoes with high basal meltwater regimes.

This new research, at Lambatungnajökull, shows a different pattern of glacier fluctuations. Lichens growing on moraine H4 in Hoffellsdalur indicate that this surface was exposed in the late 18th century. Similar geomorphological evidence exists in Skyndidalur. It is therefore likely that Lambatungnajökull reached its *Little Ice Age* maximum in the late 18th century, probably between AD 1780 and 1795 (Fig. 10.5 and 10.12). This period represents the most advanced stage of glaciation in southeast Iceland during the last 1000 years.

The last 70 years

The record of glacier-front fluctuations at Lambatungnajökull, since the late 18th century, compares well with those of other similar-sized glaciers in Iceland since AD 1930. Measurements of Skaftafellsjökull, Fjallsjökull and Skálafellsjökull, taken from the Icelandic Glaciological Database (Sigurdsson, 1998), are shown in Figure 8.13. Both the magnitude of glacier recession and the timing of various stillstands and advance events correspond quite closely between all four selected glaciers. Overall, Lambatungnajökull retreated c.1200 m between ~1930 and 1999. The degree of recession is similar to that

experienced at Skaftafellsjökull (~1300 m) and Skálafellsjökull (~900 m), both similar-sized, non-surging, outlets of southern Vatnajökull. The rate of recession at Lambatungnajökull was greatest during the 1930s and 1940s, and slowed in the 1960s. This is also reflected in the records of other comparable glaciers in southern Iceland (Fig. 8.13). All four glaciers shown in Figure 8.13 re-advanced slightly in the mid 1960s. The ice-front of Skálafellsjokull currently occupies a more advanced position than in 1967. Most glaciers in southern Iceland have fluctuated little over the last 30 years. A small-scale advance or stillstand took place at Fjallsjökull, Skaftafellsjökull and Lambatungnajökull around 1980. Skálafellsjökull data from the 1980s are missing. Since 1990 some glaciers have started to recede again.

The similarity between the 20th-century pattern of fluctuations of Lambatungnajökull, presented in this thesis, and other glaciers indicates that Lambatungnajökull is representative of glacier behaviour in southeast Iceland generally. The climatic implications of these, and other, glacier fluctuations over the last 10,000 years in southeast Iceland will now be examined.

Climatic implications

The waxing and waning of glaciers are usually a good indicator of climatic variations particularly in maritime climates (eg. Röthlisberger, 1986, Grove, 1988; Oerlemans, 1989, etc). Thorarinsson (1943) drew general conclusions regarding the climate of Iceland based on knowledge of glacier fluctuations between ~1700 and 1938. Recently, Johannesson and Sigurdsson (1998) have shown that there is a clear relationship between recorded variations in glacier size (ie. terminus position) and climate, measured at nearby weather stations, during the 20th century. The following discussion on glacier fluctuations and climatic change will focus on three timescales: sub-decadal, decadal, and centennial or longer.

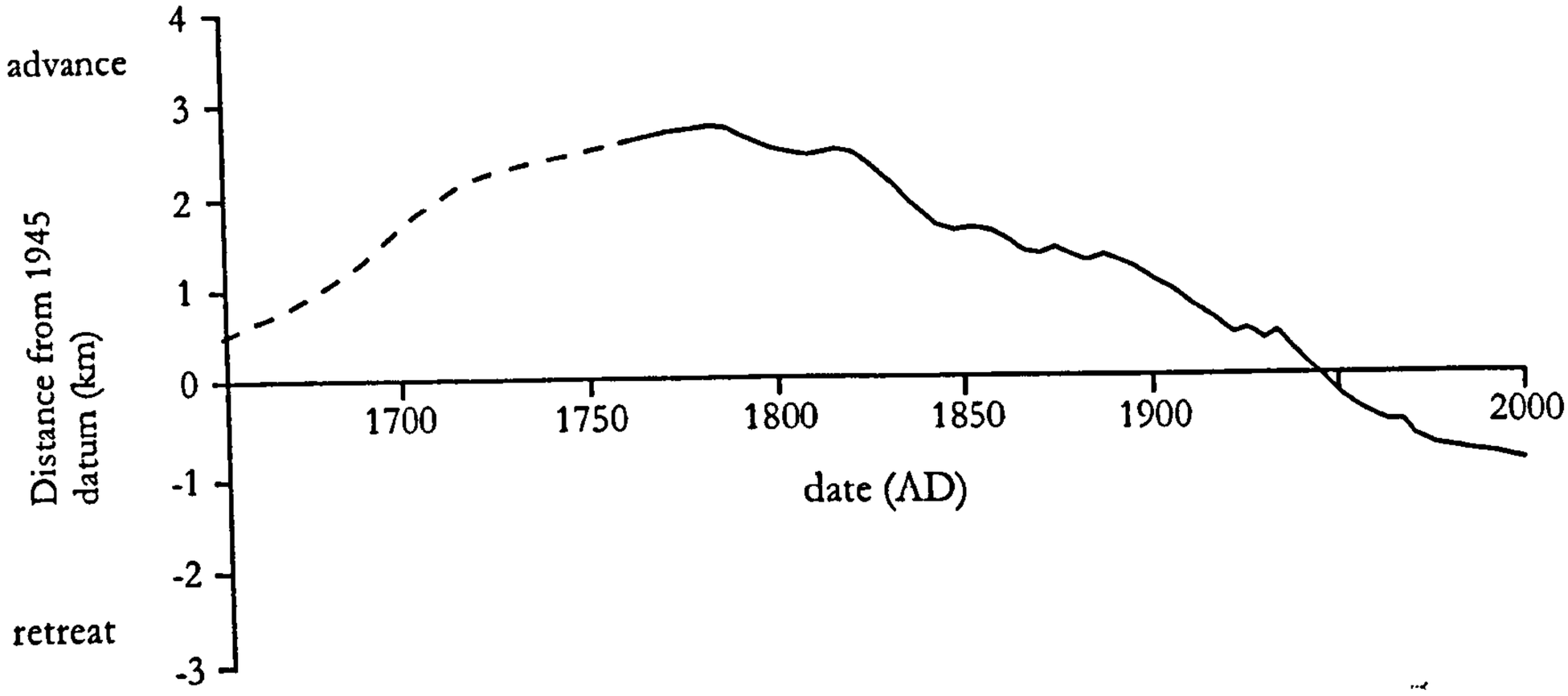
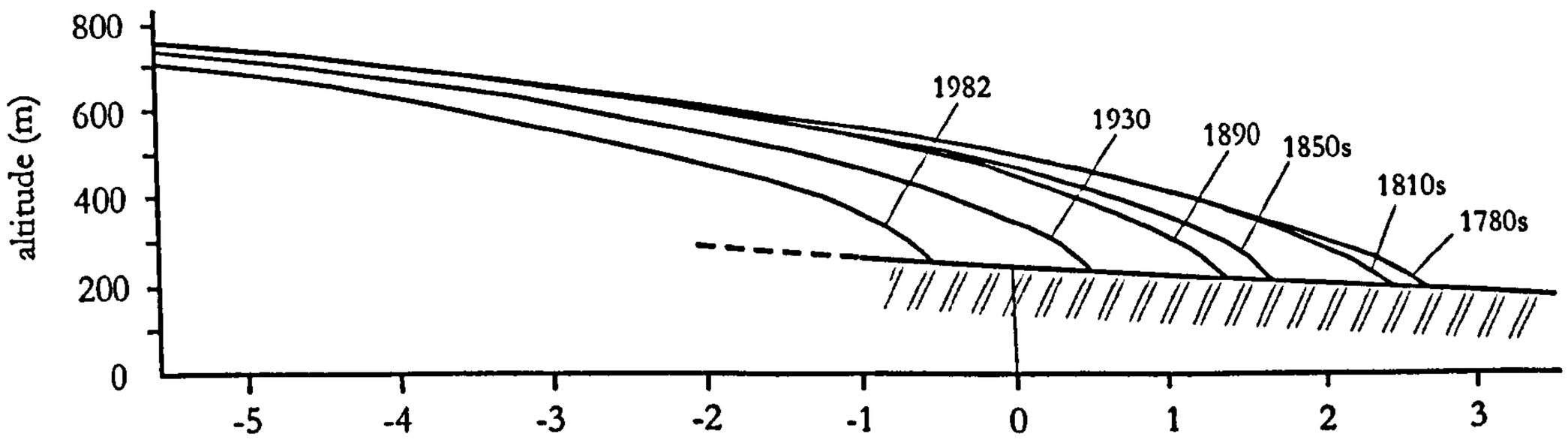


Figure 10.12 Fluctuations of the terminus of Lambatungnajökull during the last 350 years.

Distances are expressed in kilometres from the 1945 terminus position. The position of the glacier margin prior to c. 1780 has been inferred, partly from documentary evidence cited by Thorarinsson (1943), and is more uncertain.



(above) Reconstructed terminus positions of Lambatungnajökull in Skyndidalur. The long profile of the glacier has been inferred assuming a similar ice-surface slope to the present-day. Horizontal distances are expressed in kilometres from the ice-front position in AD 1945.

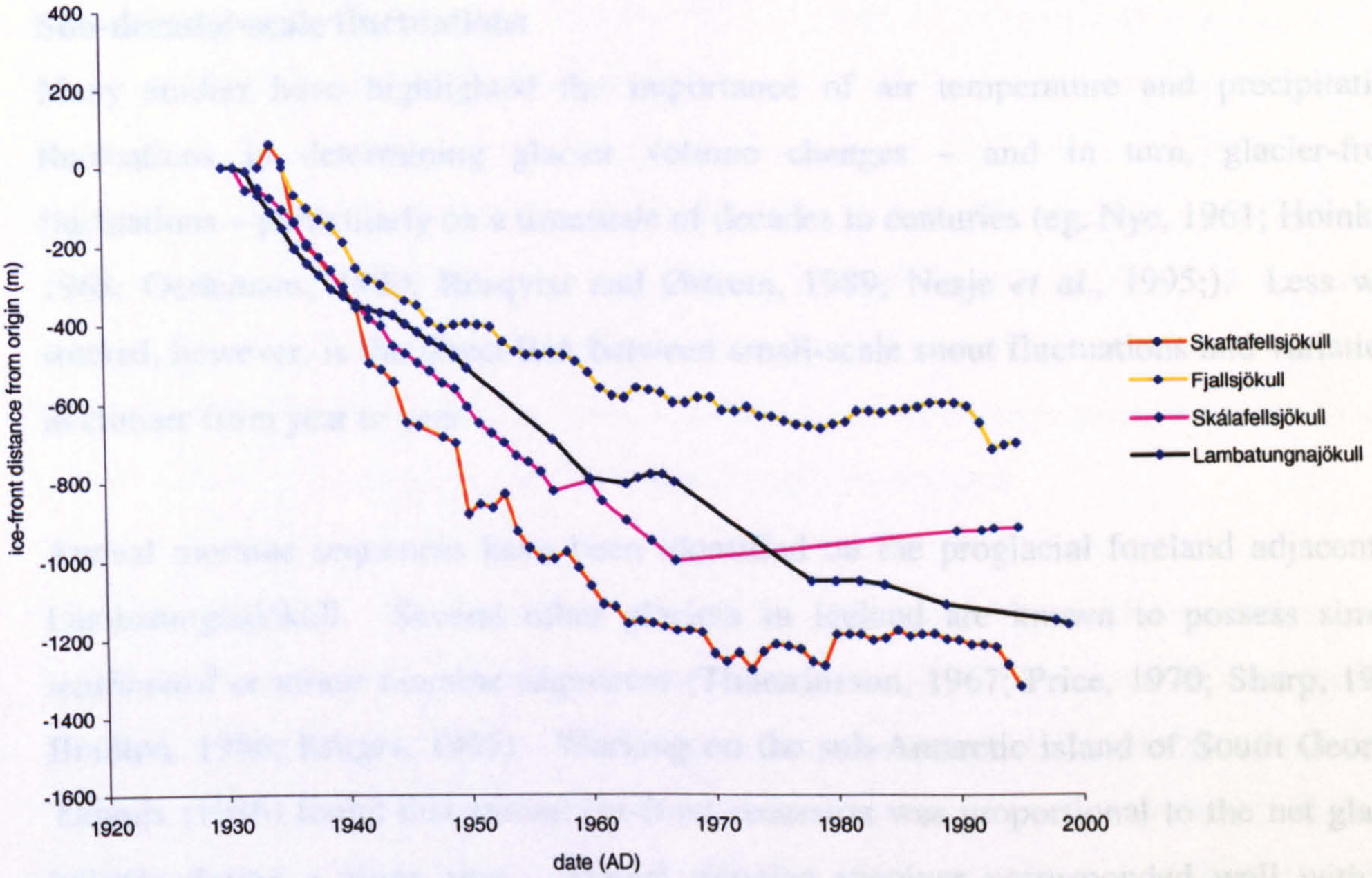


Figure 10.13 A comparison of terminus fluctuations of Lambatungnajökull with those of similar-sized valley glaciers in southeast Iceland since AD 1930 (other data taken from Sigurdsson, 1998).

The data are plotted relative to the terminus position in 1930, keeping with the convention of the Icelandic Glaciological Society.

The measurements are of :

Skaftafellsjökull - (No.2, 3)
Fjallsjökull - (Fitjar)
Skálafellsjökull
Lambatungnajökull - (Skyndidalur)

All four glaciers have retreated considerably over the last 70 years. Skaftafellsjökull has shown the greatest degree of recession – c. 1300 m. Lambatungnajökull has behaved in a similar manner, retreating c.1200 m in total, since 1930. Skálafellsjökull retreated strongly during the 1930s and 1940s but now occupies a more advanced position now than it did in 1965. Fjallsjökull – the broadest of the five glaciers – has shown the least degree of frontal recession.

Sub-decadal-scale fluctuations

Many studies have highlighted the importance of air temperature and precipitation fluctuations in determining glacier volume changes – and in turn, glacier-front fluctuations – particularly on a timescale of decades to centuries (eg. Nye, 1961; Hoinkes, 1968; Oerlemans, 1989; Rosqvist and Østrem, 1989; Nesje *et al.*, 1995;). Less well studied, however, is the direct link between small-scale snout fluctuations and variations in climate from year to year.

Annual moraine sequences have been identified on the proglacial foreland adjacent to Lambatungnajökull. Several other glaciers in Iceland are known to possess similar *washboard* or minor moraine sequences (Thorarinsson, 1967; Price, 1970; Sharp, 1984; Boulton, 1986; Kruger, 1995). Working on the sub-Antarctic island of South Georgia, Timmis (1986) found that annual ice-front recession was proportional to the net glacier balance during a given year. Annual moraine spacings corresponded well with the measurements of net balance over a five-year period. Kruger (1995) made a general comparison between annual retreat of the north margin of Mýrdalsjökull in southern Iceland and mean summer air temperature trends in Stykkisholmur since 1906. His findings implied that these minor ice-front fluctuations were in tune with variations in air temperatures during the ablation season.

The margin of Lambatungnajökull retreated considerably between the late 1890s and 1950. Figure 10.14 is a map of those moraines in Skyndidalur that formed during this time. Three major advances occurred at around AD 1890, 1920 and 1930, each forming a conspicuous moraine ridge. Between 1898 and 1915 the glacier deposited a series of 17 closely-spaced, low elevation, *arcuate* moraines. Geomorphological and lichenometric evidence suggests that these features formed annually. Between 1932 and 1950, the intensely crevassed ice-margin retreated rapidly, forming a second sequence of *sawtooth* annual moraines.

There is a close correspondence between mean summer (June-September) air temperature (measured at Hólar i Hofn and Teigarhorn) and the rate of ice-front recession of Lambatungnajökull (Fig. 10.15a). Between AD 1900 and 1914 this relationship is only a

general one. However, during the years 1932 to 1949 the relationship is striking. The warmest summers in 1933, 1934 and 1936 result in the most rapid glacier recession of between 40 and 60 m/yr; whereas in the cooler years of 1942, 1943 and 1944 ice-front retreat is only 8 m/yr, on average. The implication is that differences in ice-front retreat of Lambatungnajökull during the first half of the 20th century (1900-1915 and 1932-1950) can be largely attributed to variations in mean summer air temperature from year to year. It is likely that such high-resolution annual-moraine sequences, wherever found, could be used as suitable proxies for annual net glacier balance.

The glacier advances of ~1918-1922 and ~1925-1930 involve more than merely summer temperature variations. The advance of ~1920 coincides with a spell of cold weather, when winter temperatures in southeast Iceland were comparable to those of the 1890s (Fig. 10.15b). It is likely that reduced ablation, combined with heavy snowfalls, in the spring and autumn of 1917, 1918 and 1919 resulted in a positive mass balance for those years. Although a considerable degree of ablation probably occurred in the mild summers of 1925-1930, it was clearly insufficient to prevent the glacier margin from advancing. A large amount of winter precipitation in 1922, much of which probably fell as snow, may have promoted a glacier advance following a lag of 3-5 years (Fig.10.15c). However, the exact cause of this re-advance around 1925-30 is still not well understood.

A small advance in ~1963-66 and a stillstand in ~1978-1982 both coincide with periods of low mean annual temperatures, whilst winter precipitation levels remained relatively constsnt (Fig. 10.15d). During these cold years, the length of the ablation season would have been significantly curtailed allowing the glacier margin to advance unchecked.

To summarise, the retreat rate of Lambatungnajökull from year to year is primarily controlled by variations in summer temperature. Advance events appear to be a more complex response to a combination of climatic variables. The glacier is most likely to advance as the result of a few years of below-average annual temperatures. Winter precipitation has played a less significant role in affecting terminus fluctuations on an annual timescale.

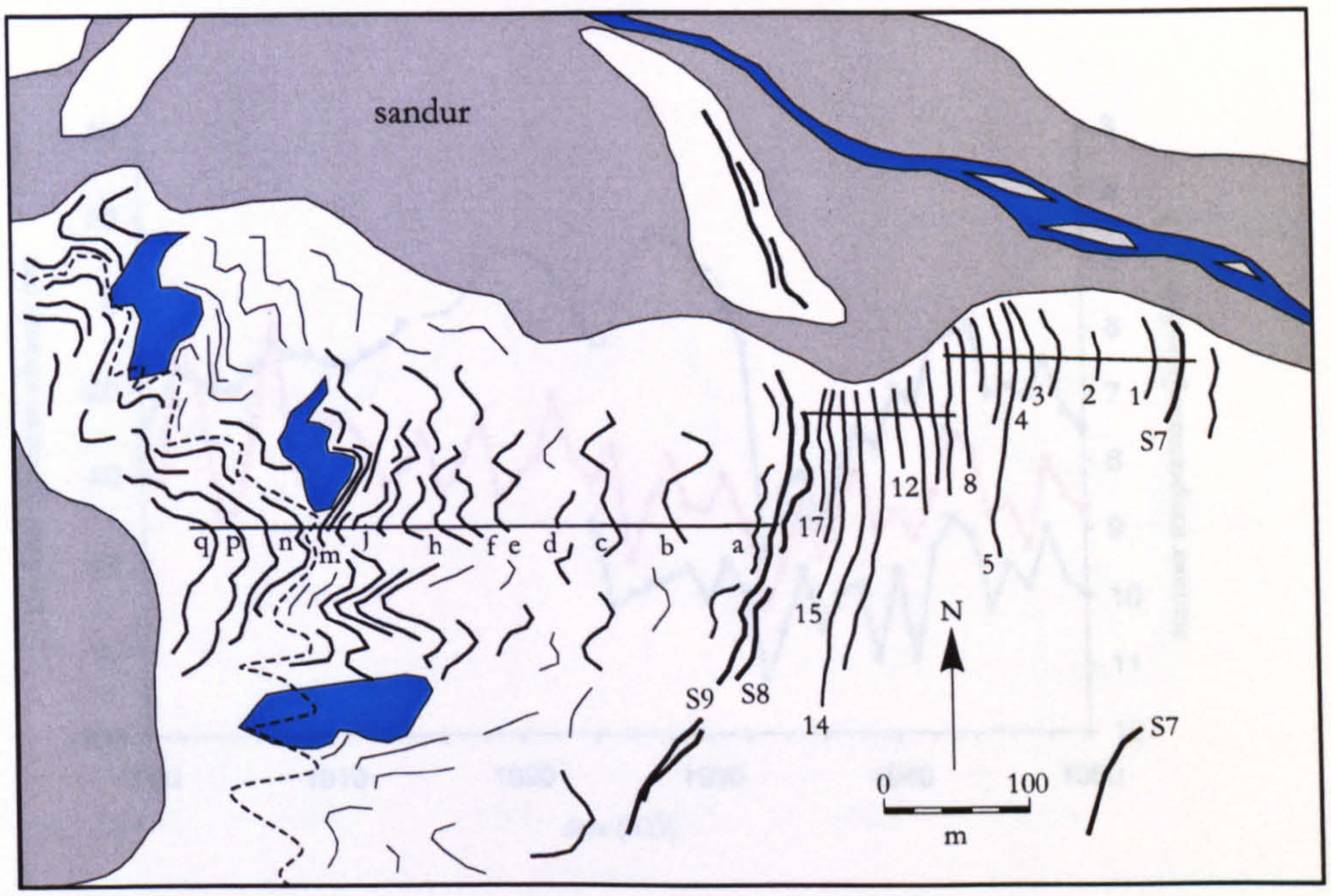


Figure 10.14 Map of moraines in Skyndidalur formed between c.1890 and 1950. The ice-margin position (dashed line) is traced from the September 1945 aerial photograph.

Lichenometric studies combined with geomorphological evidence have confirmed that ridges 1-17 and A-P are *annual moraines*. Accepting that the lichenometric dating technique has a precision of 2 or 3 years, there is a correpondence between the number of years elapsed and the number of minor moraines formed.

Formation dates for the moraines are given below. Dates underlined are based on lichenometric evidence. *Moraine m* is of known age. The third column states the moraine spacing (in metres) between successive ridges. This represents the amount of ice-front retreat during the year. For example, the glacier receded 20 m during 1898; *moraine 1* was formed during the spring of 1899 following the small winter re-advance or stillstand. The ice-margin subsequently retreated a total of 40 m between spring 1899 and spring the following year.

S7	1898	20	
1	1899	40	
2	1900	30	
3	1901	20	
4	1902	15	
5	1903	18	
6	1904	20	
7	1905	20	
8	1906	15	
9	1907	10	
10	1908	12	
11	1909	12	
12	1910	14	
13	1911	16	
14	1912	18	
15	1913	12	
16	1914	10	
17	1915	5	
S8	1922		
S9	1932	10	
a	1933	60	
b	1934	40	
c	1935	34	
d	1936	52	
e	1937	40	
f	1938	30	
g	1939	34	
h	1940	20	
I	1941	24	
j	1942	10	
k	1943	8	
l	1944	6	
m	**1945**	22	
n	1946	20	
o	1947	22	
p	1948	14	
q	1949	26	

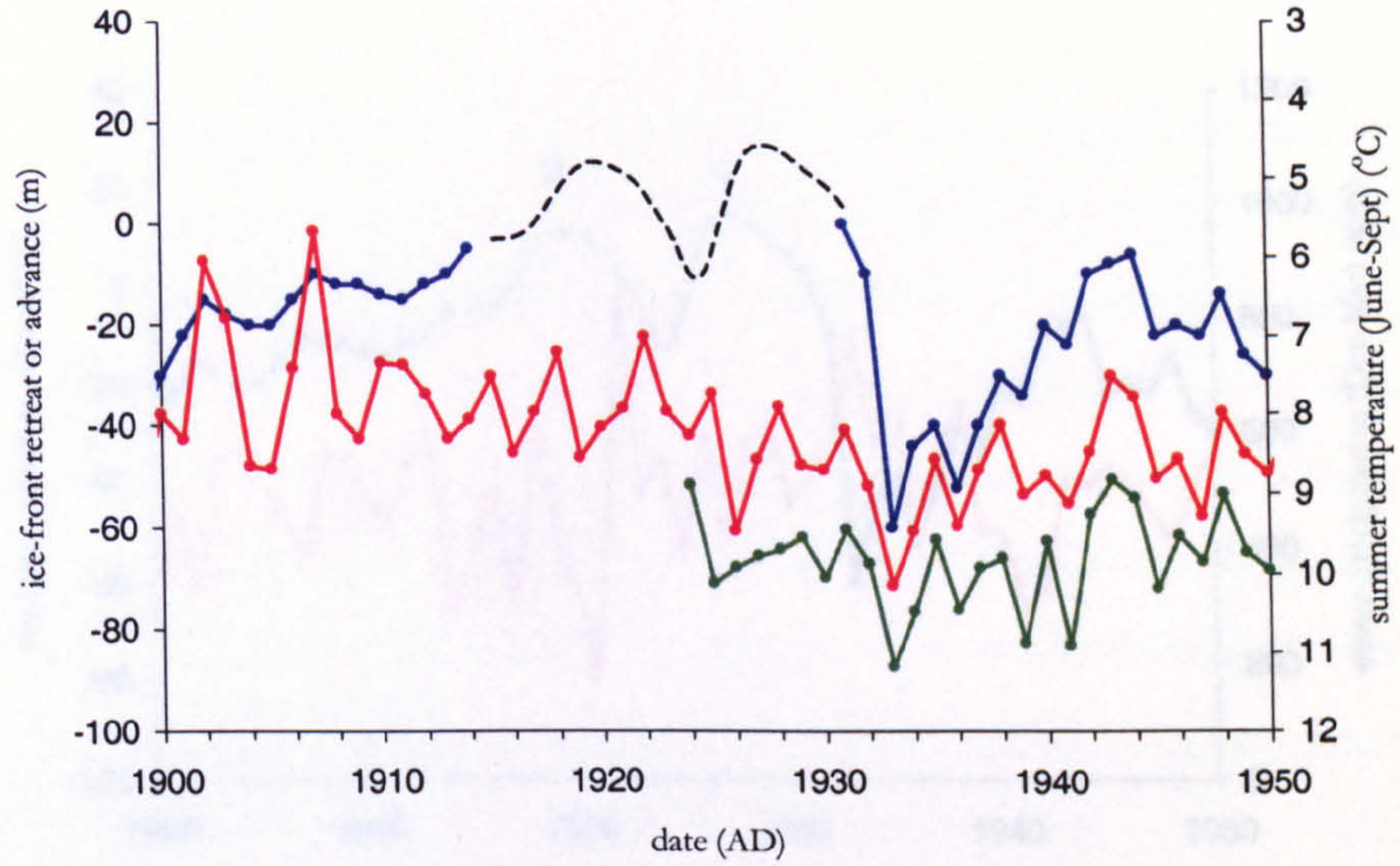


Figure 10.15 (a) Annual ice-front fluctuations of Lambatungnajokull between AD 1900 and 1950 compared with summer temperatures. The graph shows mean summer temperatures (June September) at Teigarhorn (red line) and Hólar í Hofn (green line). Data for Hólar only extend back to 1924. There is a good agreement between the degree of ice-front recession per year and summer warmth. Between 1900 and 1914 this relationship is less strong, however during the years 1930 to 1950 this relationship is striking.

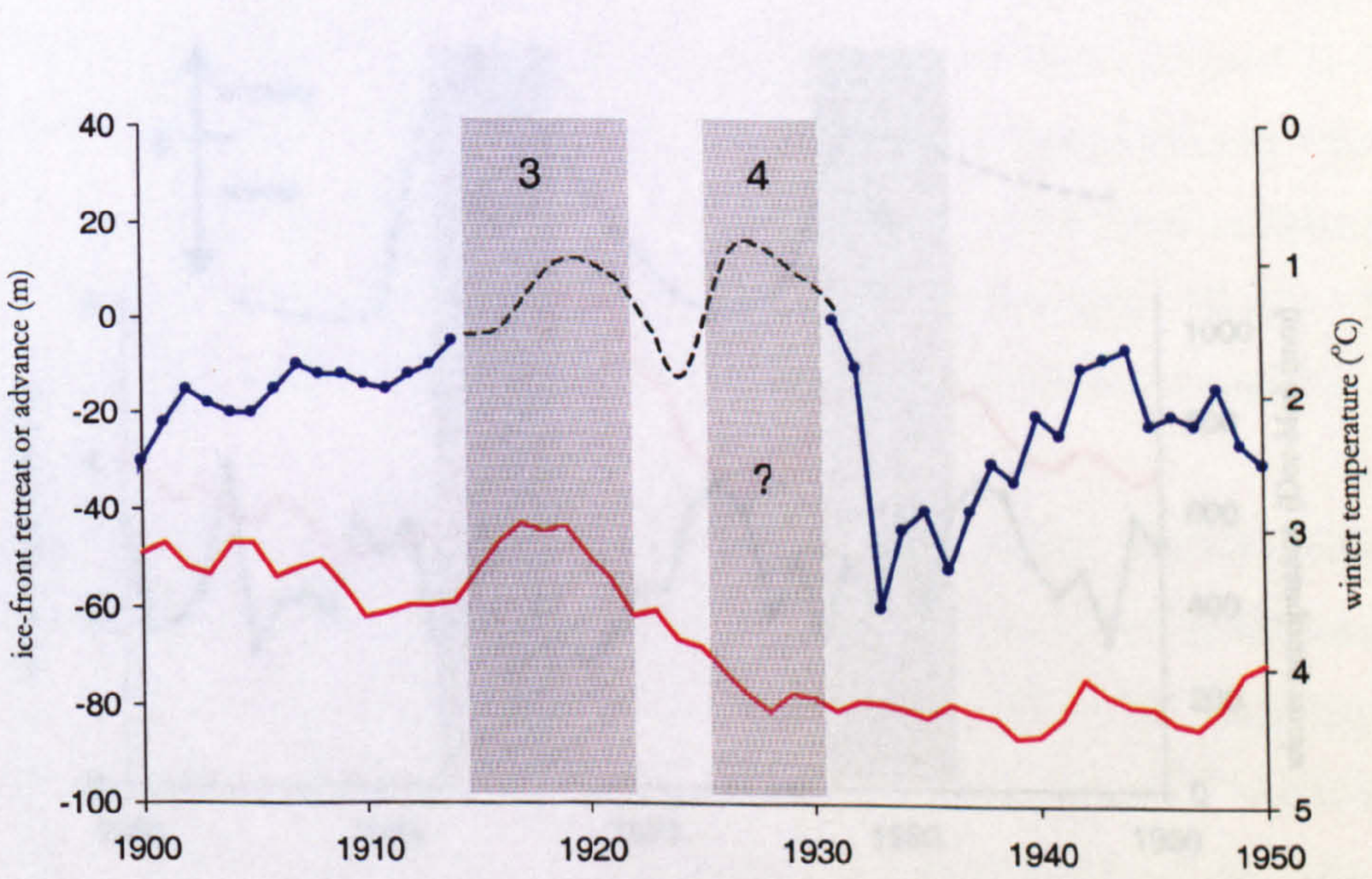


Figure 10.15 (b) Annual ice-front fluctuations of Lambatungnajökull between AD 1900 and 1950 compared with mean winter temperatures (December-March) at Teigarhorn. The cold period between 1915 and 1920 is emphasised when the data is smoothed using a 5-year running mean (red line). Glacier advance 3 is viewed as a direct result of this cooler spell of weather. Advance 4, however, appears to be unrelated to seasonal temperature variations.

[Note that temperature scales are inverted.]

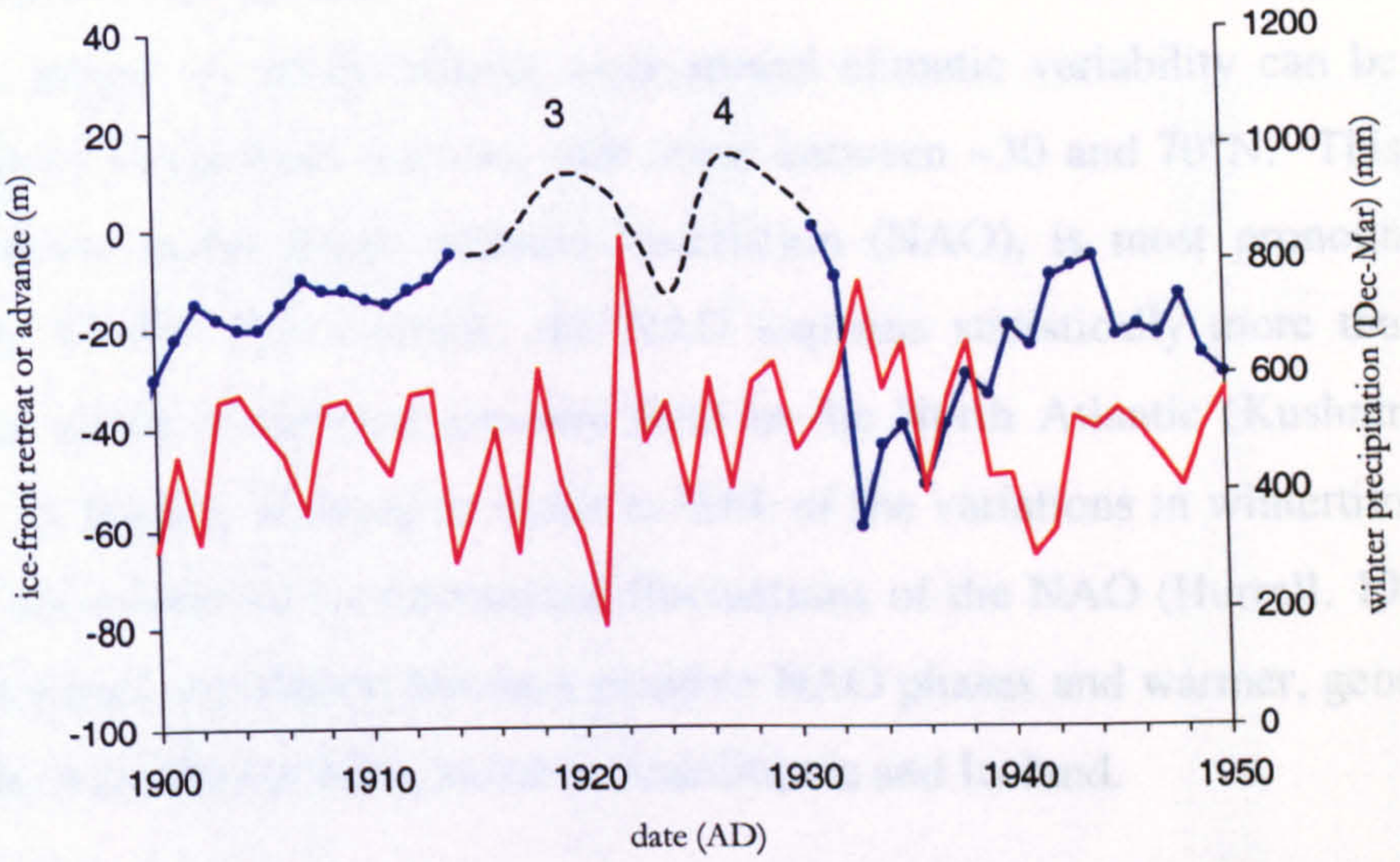


Figure 10.15 (c) Annual ice-front fluctuations of Lambatungnajökull between AD 1900 and 1950 compared with winter precipitation (December-March) measured at Teigarhorn (red line). Although advance 3 appears to be primarily as a result of lowered temperatures (see Fig. 10.15 (b)), the precipitation spike in 1922 may partly explain advance 4, allowing for a lag of 3-6 years.

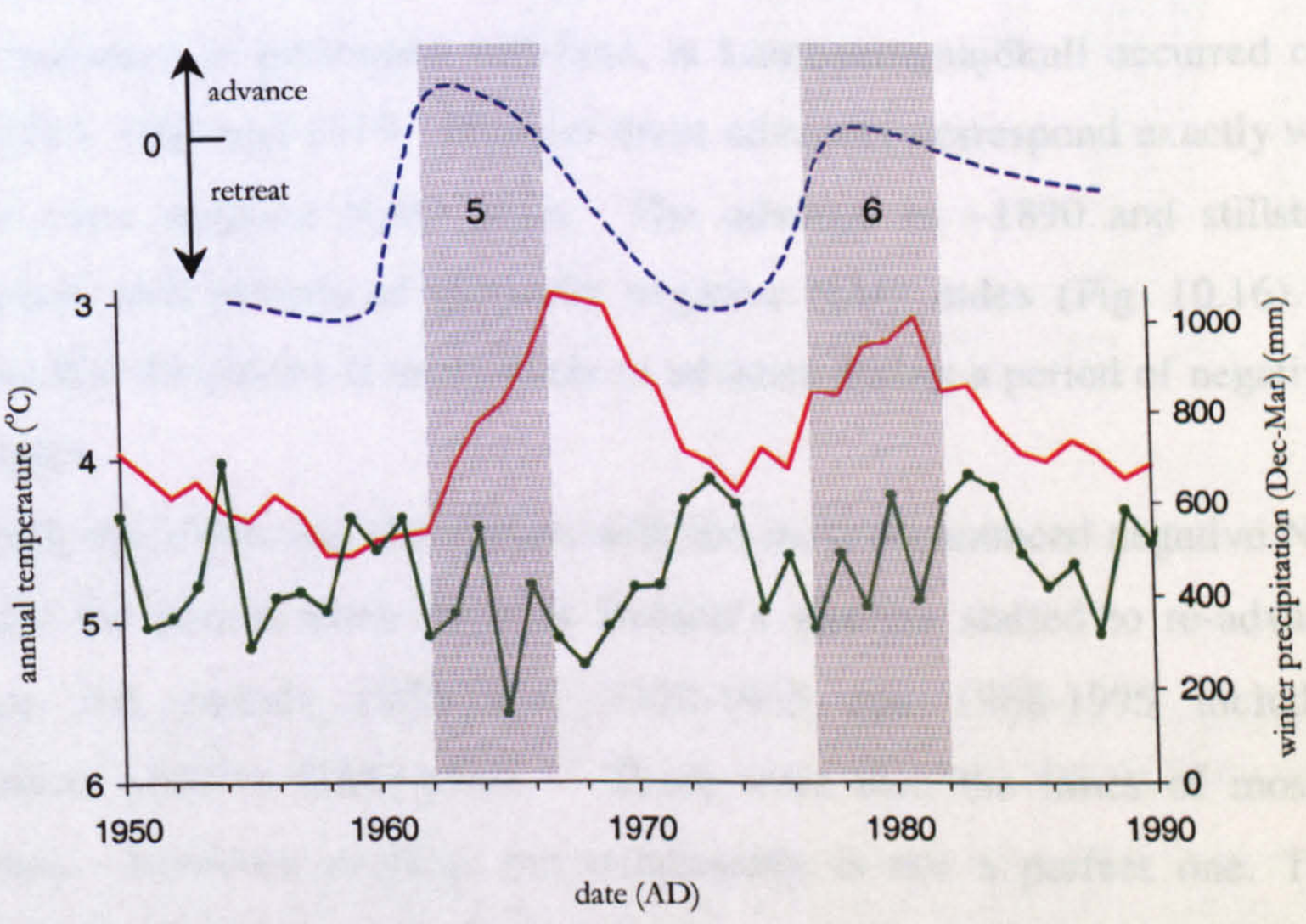


Figure 10.15 (d) Relative movements of the ice-front between AD 1950 and 1990 (blue line) compared with mean annual air temperature [5-yr running mean] (red line) and winter precipitation (green line) recorded at Teigarhorn. Retreat rates are estimates based on all the available moraine evidence and four aerial photographs. The advance in the mid-1960s coincides with a period of falling temperatures (5), as does the smaller stillstand around 1980 (6). Winter precipitation appears to play a less significant role in affecting small-scale fluctuations of the terminus.

North Atlantic Oscillation

A large degree of North Atlantic inter-annual climatic variability can be explained by fluctuations in sea-level pressure that occur between ~30 and 70°N. This pressure see-saw, known as the North Atlantic Oscillation (NAO), is most pronounced in winter (Rogers, 1984). For example, the NAO explains statistically more than 36% of the variance of the wintertime pressure field in the North Atlantic (Kushnir and Wallace, 1989). In Bergen, Norway as much as 60% of the variations in wintertime precipitation can be accounted for by interannual fluctuations of the NAO (Hurrell, 1995). Similarly there is a high correlation between positive NAO phases and warmer, generally stormier, weather in the British Isles, western Scandinavia and Iceland.

When the record of ice-front fluctuations from Lambatungnajökull is compared with the wintertime NAO index since AD 1865, an interesting pattern emerges (Fig. 10.16). Glacier advances generally coincide with periods of persistent negative NAO (Dec-Mar) index. Four or more consecutive winters with negative NAO indices have occurred only four times since 1864 – in 1870-1873, 1915-1919, 1962-1966 and 1968-1971. Periods of glacier advance, or prolonged stillstand, at Lambatungnajökull occurred c. 1870, 1890, 1918, 1925, 1965 and 1979. Three of these advances correspond exactly with periods of four or more negative NAO years. The advance in ~1890 and stillstand in ~1980 correspond with periods of generally negative NAO index (Fig. 10.16). This pattern suggests that the glacier is more likely to advance during a period of negative NAO (Dec-Mar) index.

Overall, the 1960s was the decade with the most pronounced negative NAO index and was also the period when most of Iceland's glaciers started to re-advance. In sharp contrast, the periods 1905-1915, 1920-1955 and 1988-1995 included numerous, successive, positive NAO years. These were also the times of most rapid glacier recession. However striking, the relationship is not a perfect one. The advance of Lambatungnajökull in ~1925 is anomalous as it took place during a period of predominantly positive NAO years. Furthermore, negative NAO phases around 1900 and 1940 do not appear to coincide with glacier advances.

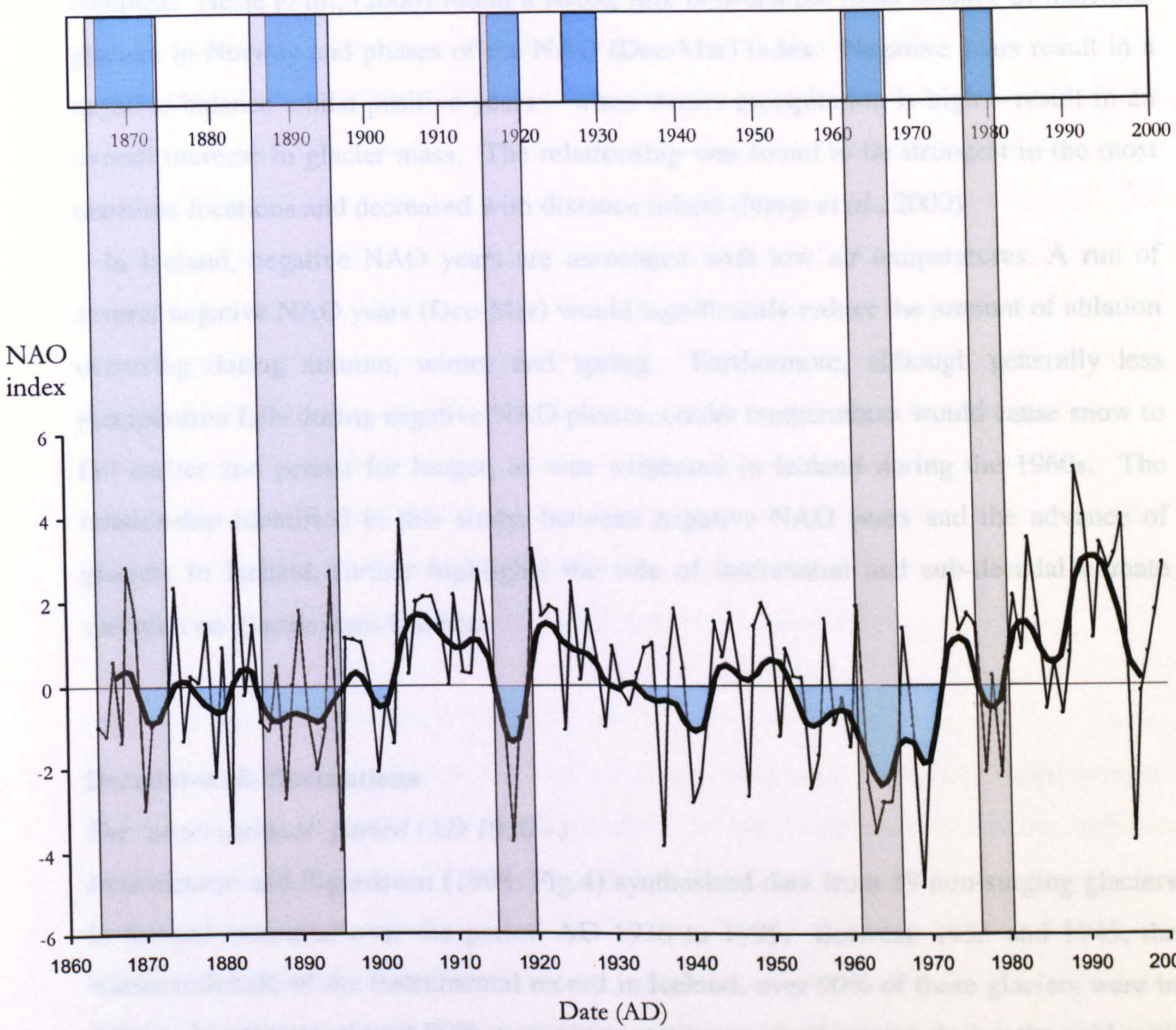


Figure 10.16 Comparison of the North Atlantic Oscillation index (winter) with glacier fluctuations at Lambatungnajökull.

The upper diagram shows the behaviour of Lambatungnajökull over the last 140 years (deduced in this study). Blue boxes represent the periods when the ice-front was advancing or stationary for a number of years.

The lower graph shows the index of the NAO (December-March) from 1864 - 2000, as used by Hurrell (1995). The index is based on the difference of normalised sea level pressure between Lisbon, Portugal and Stykkisholmur, Iceland. Sea level pressure data have been normalised relative to the period 1864-1983. The data have been smoothed (heavy line) using a weighted, seven-point low-pass filter.

Five of the six advances of Lambatungnajökull since ~1865 coincide with periods of negative NAO index (both shaded blue). Positive NAO values dominate the periods from 1905-1915, 1920-1930, 1945-1955 and the 1990s – Lambatungnajökull retreated most rapidly during these decades.

The relationship between the *North Atlantic Oscillation* and glacier fluctuations is highly complex. Nesje *et al.*, (2000) found a strong link between the mass balance of maritime glaciers in Norway and phases of the NAO (Dec-Mar) index. Negative years result in a negative balance whilst positive years – when winter precipitation is high – result in an overall increase in glacier mass. The relationship was found to be strongest in the most maritime locations and decreased with distance inland (Nesje *et al.*, 2000)

In Iceland, negative NAO years are associated with low air temperatures. A run of several negative NAO years (Dec-Mar) would significantly reduce the amount of ablation occurring during autumn, winter and spring. Furthermore, although generally less precipitation falls during negative NAO phases, cooler temperatures would cause snow to fall earlier and persist for longer, as was witnessed in Iceland during the 1960s. The relationship identified in this study, between negative NAO years and the advance of glaciers in Iceland, further highlights the role of interannual and sub-decadal climate variation on glacier mass balance.

Decadal-scale fluctuations

The 'observational' period (AD 1930 –)

Johannesson and Sigurdsson (1998: Fig.4) synthesised data from 19 non-surging glaciers in Iceland measured over the period AD 1930 to 1995. Between 1935 and 1945, the warmest decade of the instrumental record in Iceland, over 90% of these glaciers were in retreat. In contrast, almost 80% were either stationary or advancing during the cold spell centred around 1980.

When compared with Johannesson and Sigurdson's synthesis the reconstructed record from Lambatungnajökull since 1930 agrees well with the other 19 glaciers shown (Fig. 10.17). Ice-front recession dominated the 1930s, 1940s and 1950s; whilst advances occurred in the 1960s and 1980s. After ~1990, Lambatungnajökull, like many glaciers in Iceland, began to retreat again, but at a reduced rate. There is a strong relationship between the behaviour of Icelandic glaciers and fluctuations in air temperature (Fig. 10.17). Over the last 70 years glacier behaviour, on a decadal scale, seems to correspond more closely with mean annual air temperature than mean summer temperature

(measured at Teigarhorn). This is somewhat surprising as the degree of glacier terminus recession is normally controlled primarily by variations in summer air temperature (Johannesson, 1997; Johannesson and Sigurdsson, 1998). This relationship is illustrated well in the 1940s and 1950s. The tendency for some glaciers to advance around 1950 corresponds with the period of lowered annual temperatures between 1948 and 1953. However, this slight cooling is not apparent in the summer temperature record. The summers of 1948-1953 were as warm as those of the preceding six years 1942-1947, when most glaciers in Iceland receded strongly. This observation suggests that a period of cool springs and autumns may be responsible for effectively halting glacier recession or even leading to advance. Lambatungnajökull, like most other glaciers, continued to retreat during the 1950s. A few glaciers did not retreat during the warm 1930s and 1940s; whilst others continued retreating during the cool periods of the 1960s and 1980s. This demonstrates that the relationship between glaciers and climate on a decadal scale is complex and cannot be explained by temperature fluctuations alone.

The 'instrumental' period (AD 1846 –)

Comparisons between glacier fluctuations and directly-measured air temperature in Iceland are possible over the last 15 decades. Although sporadic meteorological data exist from the 18th century it was not until 1822 that temperature measurements were made with regularity (Jonsson and Garðasson, 2001). The longest continuous, air temperature series is from Stykkisholmur in western Iceland (65° 05' N; 22° 44' W). Mean monthly values exist for the period from November 1845 to the present day. This series has been recently extended using data from Reykjavík (1823-1945), although the first few years of observations are probably not reliable (pre-1830) (T. Jonsson, *pers. comm.*, 1999). Air temperature readings from southeastern Iceland date back to January 1873 at Teigarhorn. The correlation between the temperature records at Stykkisholmur and Teigarhorn is surprisingly good even though they are 400 km apart ($r^2 = 0.66$). It was decided to use mean annual temperatures at Stykkisholmur to reconstruct air temperatures at Teigarhorn between 1846 and 1872. Mean annual temperatures were compared over the 30-year period 1873-1903, and a linear regression analysis was performed. Mean annual air temperatures for Teigarhorn were reconstructed using the

regression equation $y = 1.005x - 0.17$. Generally, annual temperature values at Teigarhorn between 1873 and 1846 were 0.17°C cooler than at Stykkisholmur.

A comparison of annual temperatures at Teigarhorn with ice-front fluctuations of Lambatungnajökull is revealing (Figure 10.18). Six of the seven periods of glacier advance, or stillstand correspond with cold spells – when mean annual temperatures dipped below the average (AD 1873-1972) for several years. The coldest part of the last 150 years was between ~1855 and 1895, during which time Lambatungnajökull retreated little and advanced twice. The warmest decades, the 1930s and 1940s, were characterised by the greatest degree of frontal recession. However, a small advance of the ice-front between ~1925-1930 cannot be explained by air temperature variations alone (see Fig. 10.15).

The 'documentary' period (AD ~900 –)

Prior to the collection of reliable meteorological data in Iceland (ie. before AD ~1846) the most useful sources of climatic information are historical documents and written accounts. Those of most value relate to Arctic drift ice off the coasts of Iceland because a good positive correlation exists between air temperature at the coast and sea-ice conditions (Bergthorsson, 1969; Ogilvie, 1981). Several authors have reconstructed sea-ice indices for Iceland spanning the last 1000 years or so (Thorodssen, 1916-1917; Koch, 1945; Thorarinsson, 1956; Sigtryggsson, 1972; Ogilvie, 1984). The reliability of various annals and documentary sources used to compile these indices has been questioned (Vilmundarsson, 1972; Bell and Ogilvie, 1978). In particular, certain periods have yielded very little climatic information, for instance the 13th and 14th centuries. Thorarinsson (1943) cautions against the use of drift-ice reports dating from before the 17th century. However, reports from after ~1700 are deemed fairly reliable (Thorarinsson, 1956; Sigtryggsson, 1972).

It was decided to examine the relationship between glacier fluctuations in southeast Iceland and sea-ice conditions. The study draws on the sea-ice record put forward by Koch (1945) and later extended up to 1980. Ogilvie (1984) states that the data are based

mainly on work by Thoroddsen (1916-1917) and comprise annual sea-ice incidence (in weeks) for every year since 1600. Prior to 1600 the values are ten-year averages. The raw data are available in Lamb (1977). It should be noted that prior to ~1700 those years where no sea-ice was reported do not necessarily represent ice-free years. The absence of reported sea-ice sightings may be due to a range of factors and cannot be taken as an absolute indicator of clear sea conditions.

When compared with the record of glacier fluctuations of Lambatungnajökull since the 17th century, a clear relationship emerges. Lambatungnajökull was largest (ie. most advanced) during periods when sea ice was most common and persisted at the coast for longest. The heaviest sea-ice years occurred in the 1780s, coinciding with the time when Lambatungnajökull was at its greatest extent in recent centuries (Fig. 10.19). Ogilvie (1992: 114) has recently asserted that the 1780s stand out as the coldest decade of the period AD 1500-1800 and probably of the last 500 years in Iceland. A second sea-ice maximum occurred between AD 1860 and 1880, during which time Lambatungnajökull advanced to within 500 m of its late-18th-century position. Drift ice was rare in the 1920s and 1930s and coincided with a period of rapid glacier retreat. An increased frequency of sea ice in Icelandic waters during the 1960s brought unusually low temperatures to southern Iceland, prompting a small re-advance of Lambatungnajökull (Fig. 10.19).

Prior to c. AD 1650 the size of Lambatungnajökull is uncertain. Although difficult to substantiate, it is likely that glaciers were significantly smaller in Iceland, when the island was settled in the ninth century, than at any time since. Furthermore, during the last 1000 years, most glaciers have not exceeded those dimensions attained during the period AD ~1750-1900, when conditions in Iceland probably resembled most closely those of a *"Little Ice Age"*.

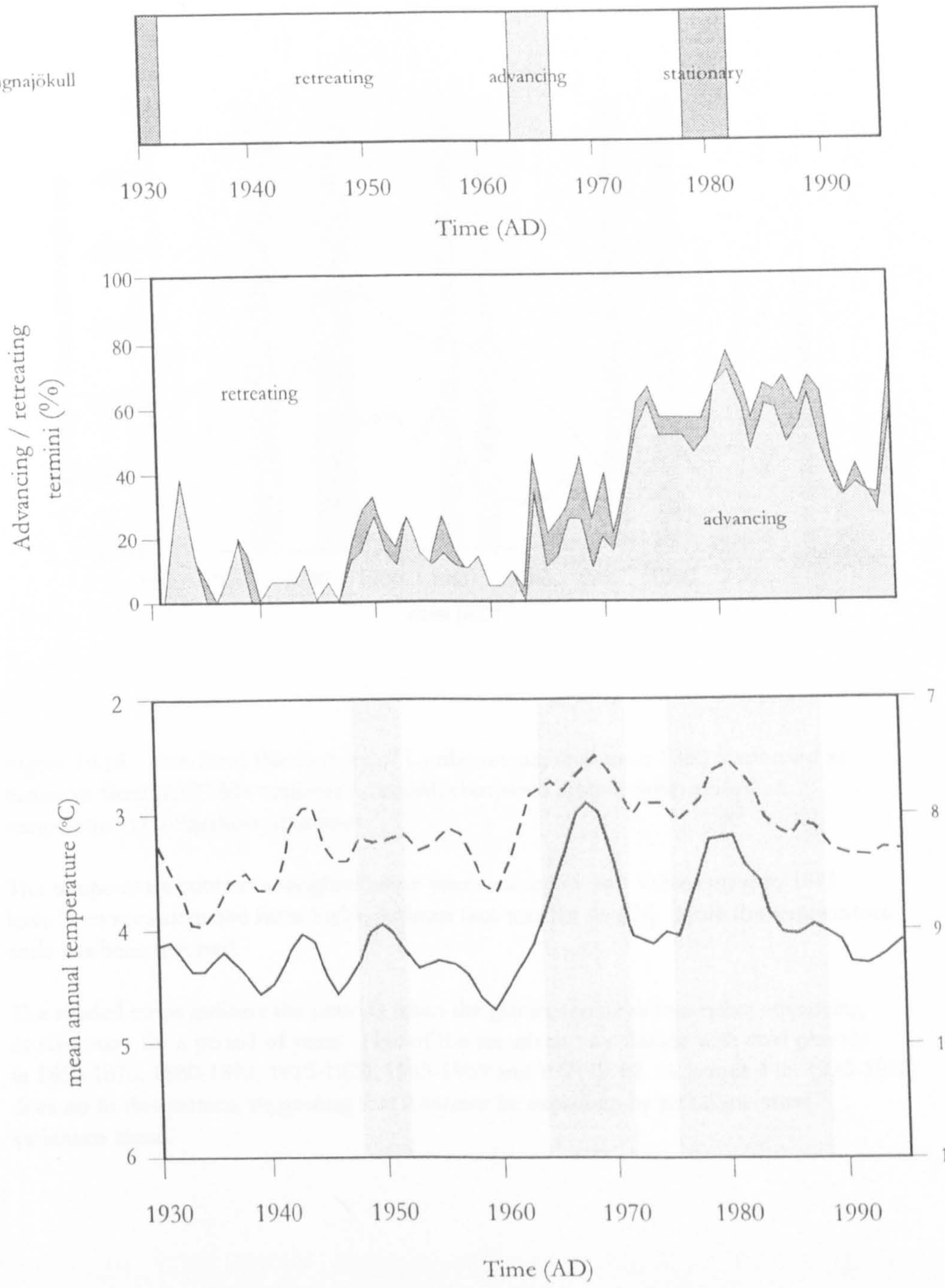


Figure 10.17 Percentage of advancing and retreating glacier termini in Iceland as a function of time (taken from Johannesson and Sigurdsson, 1998: Fig.4), compared against air temperature over the same period. Fluctuations of Lambatungnajökull deduced in this study are also shown for comparison (uppermost diagram).

Solid line – weighted five-year running mean of annual temperature at Teigarhorn.
Dashed line – weighted five-year running mean of summer temperature at Teigarhorn.
Note the scales are inverted.

A greater number of glaciers began to advance during the mid-1960s and between 1975-1990. The shaded strips highlight the times when glacier advances and cooler conditions coincide. Prior to 1935 the dataset comprises fewer glaciers. (See text for a more detailed discussion.)

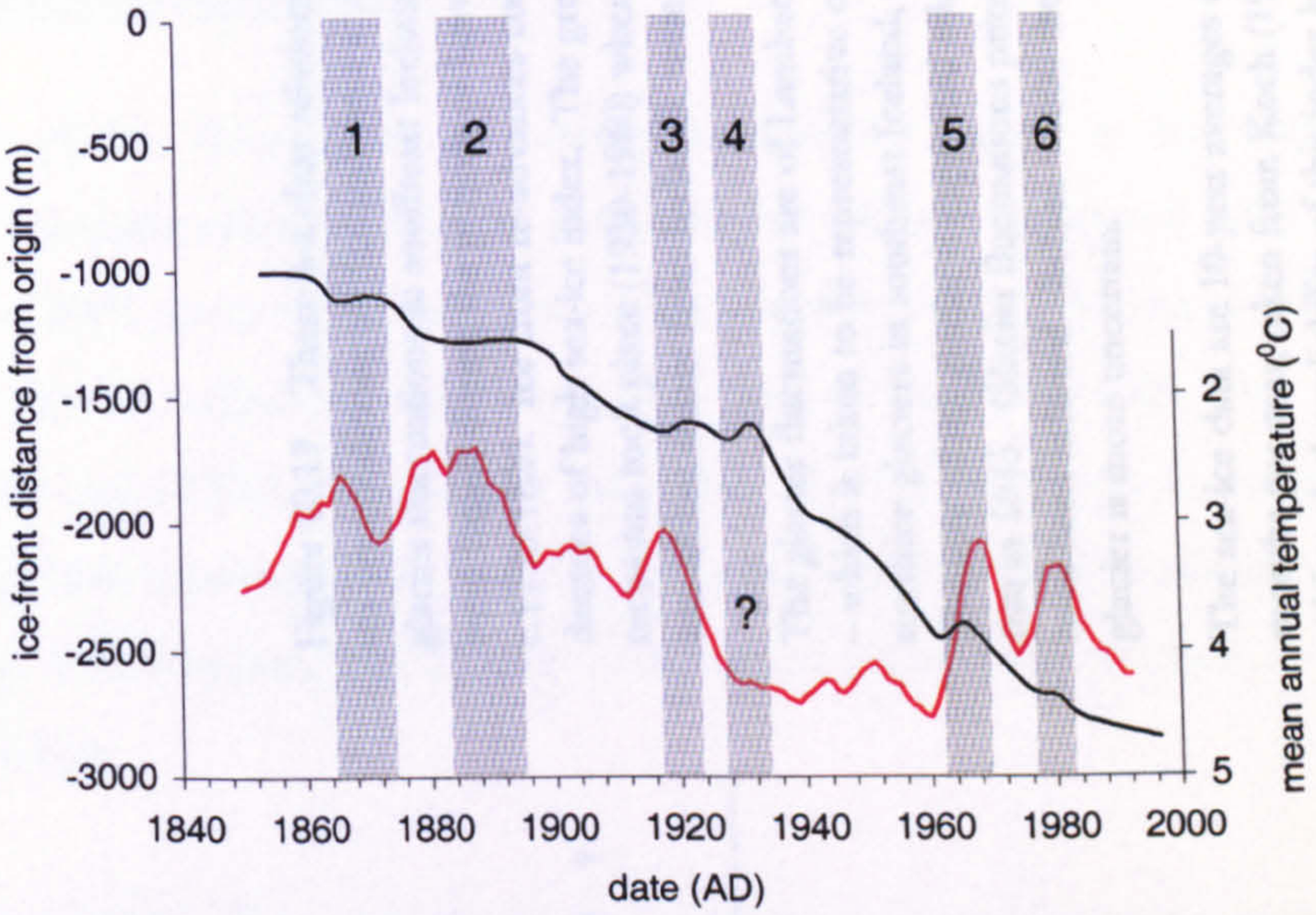


Figure 10.18 Ice-front fluctuations of Lambatungnajokull since 1850 (expressed as distances from the 1780s terminus position), compared against mean annual air temperature at Teigarhorn (red line).

The temperature curve is a weighted, nine-year running mean. Values prior to 1873 have been reconstructed from Stykkisholmur (see text for details). Note the temperature scale has been inverted.

The shaded strips indicate the periods when the glacier terminus was either advancing or stationary for a period of years. Five of the six advances coincide with cold periods in 1860-1870, 1880-1892, 1915-1920, 1963-1969 and 1977-1982. Advance 4 (c. 1925-1930) does no fit this pattern, suggesting that it cannot be explained by air temperature variations alone.

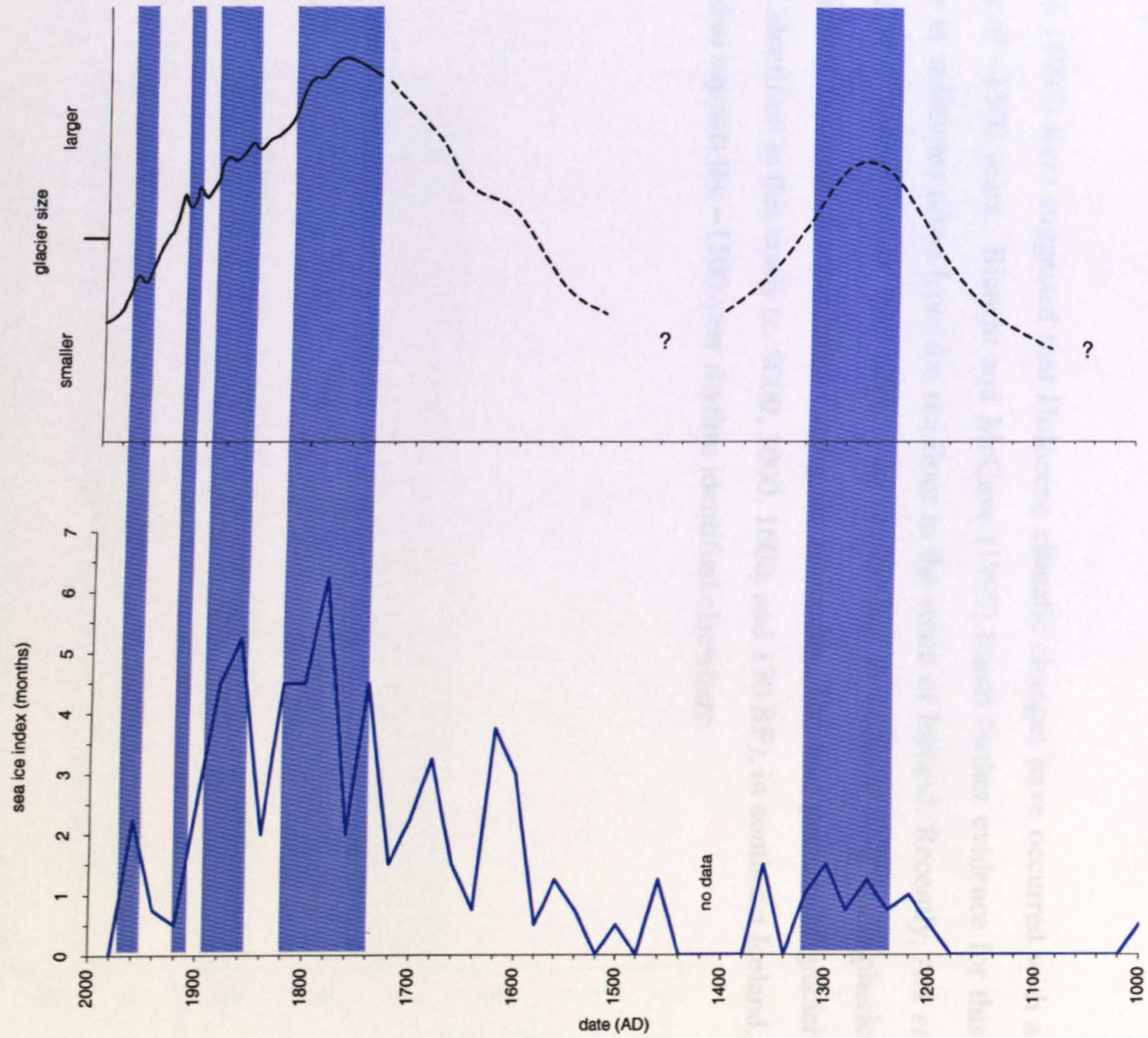


Figure 10.19 There is a clear relationship between sea-ice conditions off Iceland since c. AD 1000 and glacier fluctuations in southeast Iceland. Glaciers were largest during the periods of heaviest sea-ice – c.1750-1820. Ice front re-advances coincide with decades of high sea-ice index. The greatest recession took place (1930-1960) when drift ice was all but absent from Icelandic waters.

The glacier fluctuations are of Lambatungnajokull – which is taken to be representative of climatically-sensitive glaciers in southeast Iceland, generally. The size of the glacier is expressed relative to its size in 1945. Glacier fluctuations prior to c. 1780 have been inferred. Before ~1600 the size of the glacier is more uncertain.

The sea-ice data are 10-year averages expressed in months per year, taken from Koch (1945). Although the reliability of this index has been questioned (eg. Ogilvie 1984, 1990), it is seen broadly to represent temperature conditions in Iceland over the last 1000 years. (See text for more detailed discussion.)

Centennial-scale fluctuations

Holocene glacier fluctuations of Lambatungnajökull appear to be related to the wider record of climatic change over the same period. When compared with the $\delta\ ^{18}O$ record from the GISP2 ice-core, a proxy for air temperature, some striking similarities arise (Fig. 10.20). Glacier advances *Hoffellsdalur I* and *III* coincide with cool periods between 4500-5500 and 1200-1800 calendar years BP. Interestingly, the ice-core record suggests that the period between 1000 and 100 BP represents the coolest millennium of the entire Holocene, a finding that agrees well with the advanced state of Lambatungnajökull in the *Little Ice Age.*

Bond *et al.* (1997) have suggested that Holocene climatic changes have occurred with a periodicity of ~1500 years. Bianchi and McCave (1999) found further evidence for this periodicity in sediments taken from the sea floor to the south of Iceland. Recently, Arz *et al.* (2001) also demonstrated that large-scale variations in oceanic and atmospheric circulation have occurred several times during the Holocene. The four periods of glacier expansion identified in this study (c. 5000, 3000, 1600, and 170 BP), in southeast Iceland, would tend to support the ~1500-year rhythm identified elsewhere.

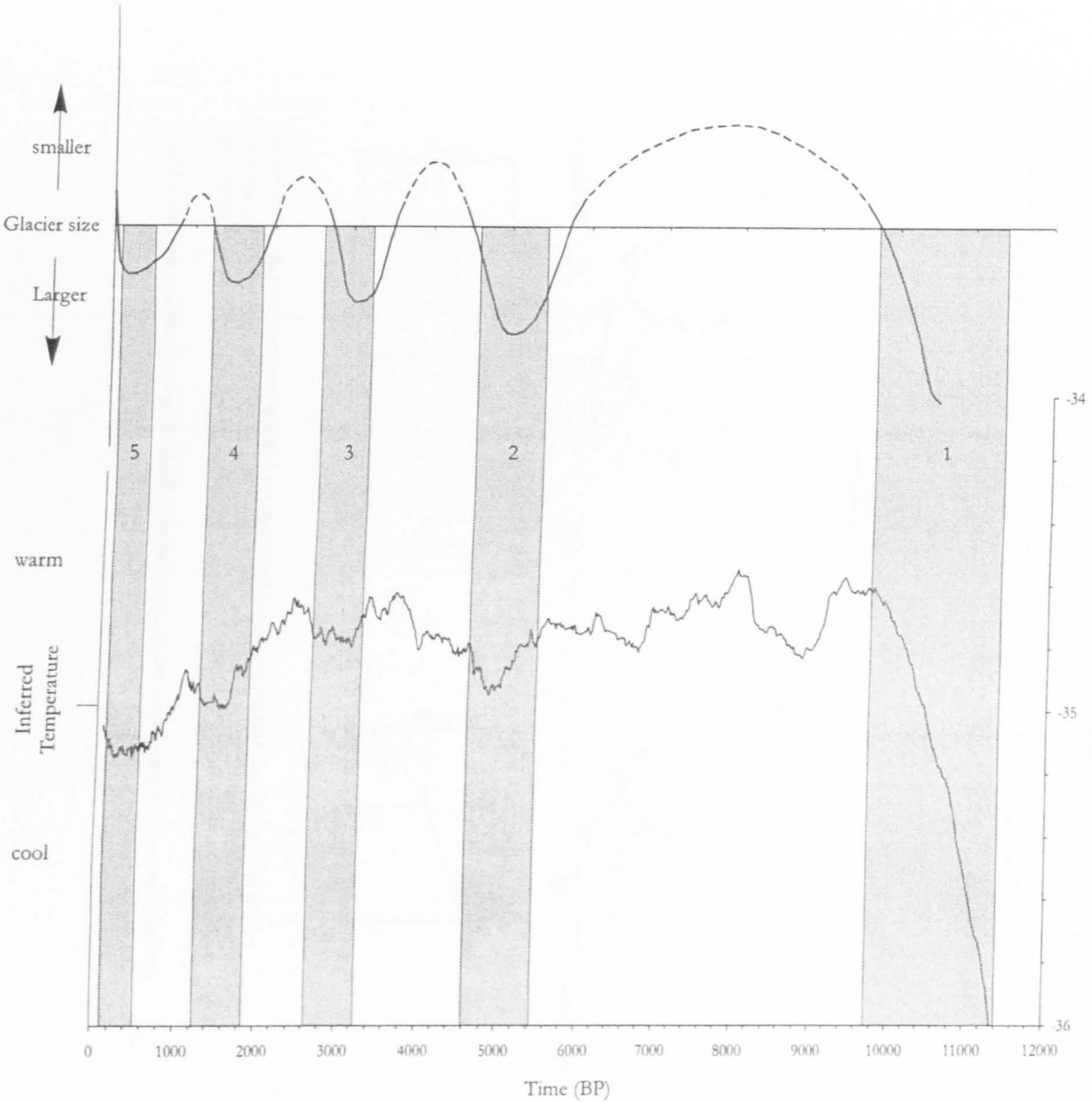


Figure10.20 Holocene temperatures, inferred from the GISP2 oxygen-isotope record, compared with fluctuations of Lambatungnajökull.

The phases of glacier expansion identified at Lambatungnajökull are as follows:

1 -	Dalsheiði stage	> 9000 BP
2 -	Hoffellsdalur I stage	c. 5000 BP
3 -	Hoffellsdalur II stage	c. 3000 BP
4 -	Hoffellsdalur III stage	c. 1600 BP
5 -	Hoffellsdalur IV stage	c. 170 BP ie. *Little Ice Age*

All five periods of glacier advance are seen to correspond with several centuries of below average temperature (shaded strips). The size of the glacier during the retreat phases is uncertain.

The oxygen-18 isotope curve has been smoothed to remove fluctuations with periods of less than 20 years.

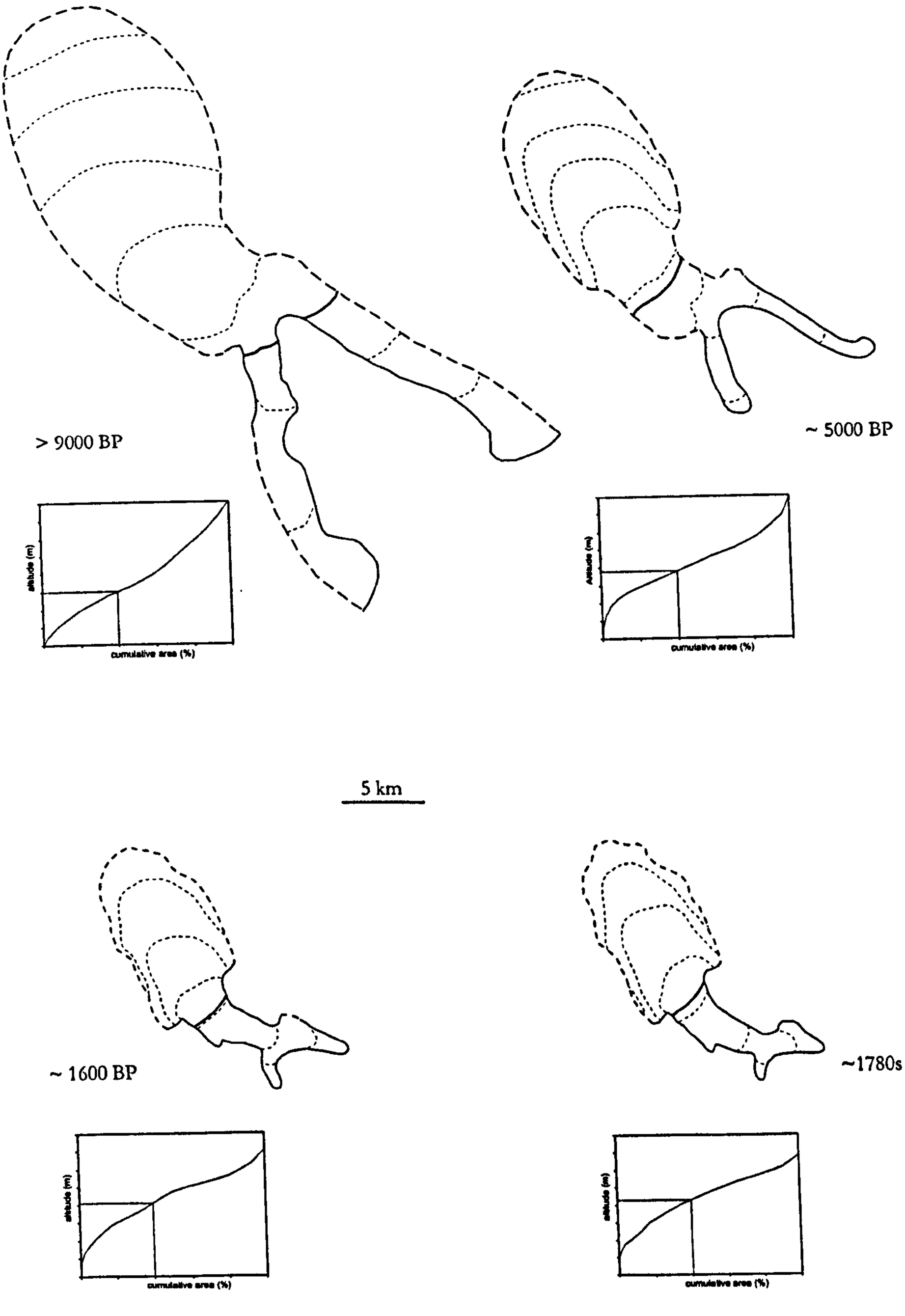


Figure 10.21 Reconstructions of Lambatungnajökull at various periods during the Holocene. All maps are to the same scale. Contours are at 200 m vertical intervals. Where the glacier comes into contact with another glacier the perimeter is dashed.

ELAs have been calculated using the AAR method (0.6) and are represented by a heavier solid line.

Quantifying the climatic changes involved

The former extent of Lambatungnajökull has been reconstructed based on geomorphological evidence. Based on this information, it is possible to constrain the amplitude of climatic change that has occurred over the last 10,000 years. The *equilibrium-line altitude* (ELA) defines a theoretical surface on the glacier where accumulation and ablation are equal. At the ELA summer temperature (ie. ablation) is closely related to winter precipitation (ie. accumulation). The nature of this relationship has been established based on data from maritime, Norwegian glaciers (Sutherland, 1984).

$$A = 0.915\, e^{0.339t}$$

A rise in winter precipitation (A), expressed in metres, should be balanced by a rise in mean summer temperature (t) if the ELA is to remain at the same elevation. To calculate changes in summer temperature, knowledge of the winter precipitation regime is required. The equation can then be re-arranged:

$$\delta T = 2.95[\log_e(1.093A)]$$

Working on Sólheimajökull in southern Iceland, a temperate glacier in a maritime climate similar to that of eastern Vatnajökull, Mackintosh (2000) developed a mass balance model capable of relating changes in ELA (ΔELA) to variations in both temperature (ΔT) and precipitation (ΔP). The model was tuned using mass balance measurements from Hofsjökull in central Iceland (Sigurdsson, 1998). The relationship is as follows:

$$\Delta ELA = \Delta T160 + 2.2\Delta P$$

It was decided to use the formula derived by Mackintosh (2000) to reconstruct the temperature changes associated with the fluctuations of Lambatungnajökull since the end of the Last Ice Age. Potential ELAs have been reconstructed, using the AAR method, for four stages of glacier advance:– *Dalsheiði*, *Hoffellsdalur I*, *Hoffellsdalur III*, and *Hoffellsdalur IV*, prior to 9000 BP, at c. 5000 BP, c. 1600 BP and AD 1780s, respectively

(Fig. 10.21). All palaeo-ELAs are compared to the modern ELA (which has been designated as the 1982 position of 980 m asl). The resulting temperature differences are expressed as deviations from the 1961-1990 mean at Hólar in Hornafjörður. Furthermore, for simplicity, the model is based on the assumption that precipitation levels have not changed markedly from those of today and that other meteorological variables have a negligible effect on glacier mass balance. The results are presented in Table 8.3, below.

An alternative climate-ELA model has been proposed by Caseldine and Stötter (1993) based on work in the more arid climate of the Tröllaskagi highlands, northern Iceland. By assuming an average vertical temperature gradient (*lapse rate*), it is possible to estimate the changes in summer temperature that correspond to reconstructed paleao-ELA values. As before, Caseldine and Stotter's model assumes constant precipitation over time, hence:

$$\delta T = \delta ELA \times (\delta T / \delta A)$$

The environmental lapse rate, $\delta T / \delta A$, was measured at Lambatungnajökull during 1999 and found to be 0.0069°C/m (Fig.4.20).

Table 8.3. Reconstructed ELA-climatic data for Lambatungnajökull

Date	Reconstructed potential ELA (m)	ELA difference from 1982 (m)	Temperature difference (°C) (Mackintosh, 2000)	Temperature difference (°C) (Caseldine and Stotter, 1993)
> 9000 BP	600	-380	-2.38	-2.62
~5000 BP	750	-230	-1.44	-1.59
~1600 BP	820	-160	-1.00	-1.10
AD 1780s	850	-130	-0.81	-0.90
AD 1982	980	0	0	0

Both sets of results are in good agreement. An ELA rise of 380 m since ~9000 years BP represents a 2.4°C warming according to Mackintosh's (2000) model and 2.6°C according to Caseldine and Stotter's (1993) model. Holocene *neoglacial* advances c. 5000 and 1600 years BP represent temperature depressions of approximately 1.0°C and

1.5°C from the present-day mean. A rise in ELA of 130 m over the last ~200 years indicates a temperature increase of between 0.80 and 0.90°C, according to both models.

These results can be partly checked against the measured temperature rise since the mid-19th century. A thermal minimum in southeast Iceland occurred between AD 1860 and 1890 according to instrumental records (Fig.10.18). Mean annual temperatures during this period are thought to have been very similar to those experienced during the late 18th century – a period of severe cold in Iceland. The degree of cooling represented by an ELA lowering of 130 m (-0.9°C) is in close agreement with that seen in the instrumental record (c. -1.0°C). These initial results are encouraging and show that the former extent of glaciers can be used to infer former climatic conditions in Iceland.

Wider significance

When placed in a wider context, these new results from southeast Iceland allow important conclusions to be drawn regarding the climate of the Northeast Atlantic over the Holocene. Studies in Scandinavia and Greenland indicate that there is a similarity between those glacier fluctuations identified in Iceland and elsewhere in the Northern Hemisphere (eg. Grove, 1988; Boulton *et al.*, 1997; Geirsdottir *et al.*, 2000; Matthews *et al.*, 2000; etc).

Arz *et al.* (2001) argue that the Holocene climate has been as unstable as the last glacial period, although they concede that the amplitude of the oscillations has been lower. The evidence for a pervasive 1-2000 year cycle is mounting, with ice-cores, ocean cores, and glacier fluctuations all detecting a similar signal (Meese *et al.*, 1994; Bond et al., 1997; Matthews *et al.*, 2000). Early Holocene warming around the North Atlantic coincided with the invasion of Atlantic waters into the Norwegian Sea (Koç *et al.*, 1993). Duplessy *et al.*, (2001) have shown that during the temperature optimum, between 6900-7900 yrs BP, warm Atlantic waters reached as far north as the Barents Sea and Svalbard (80°N). Glaciers in Iceland are thought to have retreated strongly during this period. The smaller

ice-caps may have disappeared altogether (Björnsson, 1979; Kaldal and Vikingsson, 1991).

Neoglaciation appears to occur synchronously around the Northern Hemisphere at c. 5-6000 yrs BP, although the number of studies is still limited. Following the *Holocene Optimum,* cooler conditions returned to the North Atlantic (Knudsen, *et al.*, 1998; Bianchi and McCave, 1999; Duplessy *et al.*, 2001). Imbrie *et al.* (1992) suggest that this cooling was rapid and occurred as a lagged response to the summer insolation minimum in the Northern Hemisphere. Arz *et al.*, (2001) associate this cooling with large-scale variations in the Atlantic thermohaline circulation. Ultimately, both these factors could have contributed to the re-introduction of cool conditions to the North Atlantic region. However, oxygen isotope data from sea-floor sediments indicate that polar waters migrated southwards between ~5800-6300 ^{14}C yrs BP (Duplessy *et al.*, 2001) and glacier advances culminated several centuries later (~5000 BP) (Dugmore *et al.*, 1989; Boulton *et al.*, 1997; Gudmundsson, 1998a; and *this study*).

Climatic variability continued into the late Holocene. Glaciers advanced at c. 3000-3500 and ~1500-2000 BP in Iceland. Duplessy *et al.* (2001) report a southward incursion of Polar water and an increase in sea ice in the Barents Sea during these periods. The *Little Ice Age* of recent centuries may be a continuation of the ~1500 year cycle. There is growing evidence that this, most recent, cooling event was a hemisphere-wide event (Grove, 1988; Bradley and Jones, 1993), although the exact cause remains elusive.

Chapter 11 : Conclusions

The new data presented in this thesis have established several important conclusions – both philosophical and methodological. These will be summarised under the following headings, reflecting the title of this thesis:

Lichenometry

Lichens have been used as a dating tool in this research. The following conclusions, regarding the growth history of yellow-green *Rhizocarpon* lichens and their applicability for dating studies, have been reached.

- In southeast Iceland, the relationship between lichen age and lichen size in Section *Rhizocarpon* over the last 270 years is non-linear.

- Lichen growth rates decrease with time, with larger (older) *Rhizocarpon* lichens growing more slowly than smaller ones.

- On older surfaces, the *gradient* of lichen size-frequency (log_{10}) distributions is more representative of surface age than the single *largest-lichen.*

- Lichen growth rates are related to climate. Specifically, the growth rate of Section *Rhizocarpon* over the last 100 years reflects the *oceanicity* of the climate ($r^2 = 0.95$). This relationship could be used to estimate lichen growth rates in areas where calibration surfaces are absent. In Iceland, for example, lichen growth rates in mild, maritime, south coast locations are likely to be three or four times greater than those in parts of the arid central highlands.

- In southeast Iceland, yellow-green *Rhizocarpon* lichens can be used to date surfaces exposed within the last three to four centuries. In parts of central and northern Iceland – where growth is considerably slower – lichens, where present, could be used to date surfaces over 1000 years old.

- Finally, it is highly likely that environmental change over the last two centuries in Iceland has accounted for secular variations in lichen growth rates. In maritime locations, slow-growing lichens, such as *Rhizocarpon*, probably grow more rapidly now – since the climatic amelioration of the 1920s and 30s – than they did in the cooler and drier periods of the 18th and 19th centuries.

Glacier fluctuations

Glaciers in southeast Iceland have changed in size considerably over the last 10,000 years. A long history of glacier fluctuations has been identified in the vicinity of Lambatungnajökull – a glacier flowing from the eastern margin of Vatnajökull. Moraine evidence suggests that the glacier was previously 300-400 m thicker than at present, depositing lateral moraines on the *Dalsheiði* plateau. Tephra layers in the soil deposited since this time show that this stage of glaciation must have occurred before c. 9000 BP. Other terminal moraine fragments, relating to smaller, no-longer existent, glaciers in the Hornafjörður region, have been tentatively correlated with the *Dalsheiði stage* glaciation. All these moraines formed in the *Pre-Boreal* or possibly *Younger Dryas* chronozone when a large ice sheet covered almost all of Iceland. At this time the ice-sheet margin was probably located close to the present-day coastline, ~20 km from the late 20th-century glacier terminus. However, the location and distribution of moraines in Hornafjörður indicate that not all the relief was inundated, opening the possibility that there were extensive ice-free areas in southeast Iceland during Lateglacial times.

Further glacier fluctuations, relating to the Holocene interglacial, have also been identified. Geomorphological evidence indicates that a tongue of glacier ice extended down the valley of Hoffellsdalur and terminated c. 7 km from the present-day ice margin. Tephrastratigraphy places this period of *neoglaciation* at c. 4500-6000 years BP. Younger moraines, representing subsequent re-advances of Lambatungnajökull occurred at c.3000 BP and 1600 BP. This pattern of mid- to late-Holocene glacier fluctuations is similar to those identified by Dugmore (1987, 1989), Stötter (1994) and Gudmundsson (1998) elsewhere in Iceland.

The timing of the most recent glacier advances in Iceland is a point of some contention. Previous workers, including Gordon and Sharp (1983), Thompson and Jones (1986) and Evans *et al.* (1999), have proposed that glaciers in southern Iceland, principally those of Vatnajökull, reached their maximum 'historical' extent between AD 1850 and 1900. However, this was found not to be the case. Two different lichenometric techniques were used to date moraines deposited within the last few centuries. The results indicate that Lambatungnajökull was considerably more extensive during the late 18th century, between AD ~1780 and 1796, than between AD 1850-1900. Since this time, the glacier has been in overall retreat. Smaller re-advances or stillstands occurred in the 1810s, 1850s, c.1870, c.1890 and in the 1920s. Recession was greatest during the 1930s and 1940s when ice-front retreat averaged 20 m per year.

The findings presented in this thesis have important implications for the extent of Lateglacial – Early Holocene glaciation and the timing of subsequent *neoglaciation* in Iceland.

Climatic change

Iceland has experienced considerable climatic change over the last 10,000 years. In the earliest Holocene, during the *Dalsheiði stage* glaciation, temperatures were probably c. 2.5°C lower than at present (AD 1961-1990). Between ~9-6000 yrs BP glaciers in Hornafjörður retreated into the main valleys as the climate warmed. The extent of Lambatungnajökull in ~5000 BP indicates that temperatures were ~1.5°C cooler than at present. A temperature optimum may have occurred between 9-6000 BP accompanied by an invasion of warm Atlantic waters into the Norwegian and Barents Seas. The periods centred around 3000 and 1600 BP were also at least ~1.0°C cooler than the late-20th century average.

Over the last 1000 years, since the Norse settlement of Iceland, the climate has changed. By the late 18th century, Lambtaungnajökull was more extensive than at any time in the 'historical' period (ie. since AD ~900). Temperatures in the coldest

decades of the *Little Ice Age* were ~1.0° below those of today. A change in climate was accompanied by a migration of the oceanic Polar Front. Waters of Polar origin flowed further south during much of the 18th and 19th centuries and sea-ice was common in the waters off Iceland.

Broadly speaking, the pattern of glacier fluctuations over the last 150 years, deduced in this study, reflects the temperature oscillations recorded at nearby meteorological stations on an annual and decadal scale. Much of the climatic variation experienced in Iceland over the last two centuries can be explained by secular changes in the North Atlantic Oscillation. A shift to more zonal atmospheric circulation and a weaker Icelandic Low – resulting in a greater frequency of negative NAO anomalies – may have been responsible for the cooling and associated glacier advances of the 18th and 19th centuries.

The warming of recent times should be viewed in an historical context. The temperature rise experienced in Iceland and much of Northern Europe over the last 100 years or so, marks the termination of several centuries of predominantly cold and inclement '*Little Ice Age*' conditions. The 20th century could be seen as a transition from one climatic oscillation to another. Atmospheric and oceanic re-organizations on this timescale have been a feature of much of the last 10,000 years.

Reference List

Acharius, E. 1798 *Lichenographiae Suecicae Prodomus*. Linkoping.

Ahlmann, H.W. 1938 *The Land of Ice and Fire*. London.

Ahlmann, H.W. 1948 *Glaciological research on the North Atlantic coasts*. In: RGS research series No.1, Royal Geographical Society, London, pp83.

Ahlmann, H.W. and Thorarinsson, S. 1937 Vatnajökull. Scientific results of the Swedish-Icelandic investigations. *Geografiska Annaler*, **19**: 146-231.

Ahmadjian, V. 1967 *The Lichen Symbiosis*. Blaisdell Publishing Company, 152pp

Ahmadjian, V. 1970 Adpatations of Antarctic terrestrial plants. - *In:* M. W. Holdgate (ed.): *Antarctic Ecology, Volume 2,* Academic Press, London & New York pp. 801-811

Anda, E. Orheim, O and Mangerud, J. 1985 Late Holocene glacier variations and climate at Jan Mayen. *Polar Research,* 3(2): 129-145.

Andersen, J.L. and Søllid, J.L. 1971 Glacial chronology and glacial geomorphology in the marginal zones of the glaciers, Midtdalsbreen and Nigardsbreen, south Norway. *Norsk Geografisk Tidsskrift*, **25**: 1-38.

Andersson, J.G. 1906 Solifluction, a component of subaerial denudation. *Journal of Geology*, **14**: 91-112.

André, M.F. 1985 Lichénométrie et vitesses d'evolution des versants arctiques pendant l'holocène (Région de la Baie du Roi, Spitsberg, 79°N). *Revue de Géomorphologie Dynamique*, **34**:49-72.

André, M.F. 1986 Dating slope deposits and estimating rates of rock wall retreat in Northwest Spitsbergen by lichenometry. *Geografiska Annaler*, **68(A)**:65-75.

Andrews, J.T. 1972 Glacier power, mass balances, velocities and erosion potential. *Zeitschrift für Geomorphologie*. Suppl Bd **13**: 1-17.

Andrews, J.T. 1975 *Glacial systems: An approach to glaciers and their environments*. Duxbury Press, North Scituate, Mass.

Andrews, J.T. and Barnett, D.M. 1979 Holocene (Neoglacial) moraine and periglacial lake chronology, Barnes Ice Cap, North West Territories, Canada. *Boreas*, **8**: 341-358.

Andrews, J.T. and Webber, P.J. 1964 A lichenometrical study of the northwestern margin of the Barnes Ice Cap: A geomorphological technique. *Geography Bulletin*, **22**: 80-104.

Andrews, J.T., Davies, P.T. and Wright, C. 1976 Little Ice Age permanent snow cover in the eastern Canadian Arctic: extent mapped from landsat-1 satellite imagery. *Geografiska Annaler,* **58(A)**:71-81.

Aplin, P.S. and Hill, D.J. 1979 Growth analysis of circular lichen thalli. *Journal of Theoretical Biology,* **78**: 347-363.

Appenzeller, C., Stocker, T.F. and Anklin, M. 1998 North Atlantic oscillation dynamics recorded in Greenland ice cores. *Science,* **282**: 446-449.

Armstrong, R.A. 1973 Seasonal growth and growth rate-colony size relationships in six species of saxicolous lichens. *New Phytologist*, **72**: 1023-1030.

Armstrong, R.A. 1974 Growth phases in the life of a lichen thallus. *New Phytologist*, **73**: 913-918.

Armstrong, R.A. 1975 Influence of aspect on pattern of seasonal growth in the lichen *Parmelia glabratula* ssp. *fuliginosa* (Fr. ex Duby) Laund. *New Phytologist*, **75**: 245-251.

Armstrong, R.A. 1976 The influence of the frequency of wetting and drying on the radial growth of three saxicolous lichens in the field. *New Phytologist*, **77**: 719-724.

Armstrong, R.A. 1977 The response of lichen growth to transplantation to rock surfaces of different aspect. *New Phytologist*, **78**: 473-478.

Armstrong, R.A. 1979 Growth response of lichen thalli with central portions artificially removed. *Environmental and Experimental Botany*, **19**: 175-178.

Armstrong, R.A. 1983 Growth curve of the lichen *Rhizocarcarpon geographicum*. *New Phytologist*, **73**: 913-918.

Armstrong, R.A. 1984 The influence of bird droppings and uric acid on the radial growth of five species of saxicolous lichens. *Environmental and Experimental Botany*, **24**: 95-99.

Armstrong, R.A. 1993a The growth of six saxicolous lichens transplanted to lime-rich and lime-poor substrates in south Gwynedd, Wales. *Symbiosis*, **15**: 257-267.

Armstrong, R.A. 1993b Seasonal growth of foliose lichens in successive years in South Gwynedd, Wales. *Environmental and Experimental Botany*, **33(2)**: 225-232.

Armstrong, R.A. 1994 The influence of bird droppings on the growth of lichen fragments transplanted to slate and cement substrates. *Symbiosis*, **17(1)**: 75-86.

Armstrong, R.A. 2000 The relationship between thallus size and growth rates in four species of crustose lichens. *Unpublished manuscript.*

Armstrong, R.A. and Bradwell, T. *in press* Variation in hypothallus width and the radial growth of the lichen *Rhizocarpon* Ram. Em.Th.Fr. Subgen. *Rhizocarpon*. *Symbiosis.*

Armstrong, R.A. and Smith, S.N. 1996 Experimental studies of hypothallus growth in the lichen *Rhizocarpon geographicum*. *New Phytologist*, **132**: 123-126.

Armstrong, R.A. and Smith, S.N. 1997 Factors associated with degeneration of the thallus centre in foliose lichens. *Symbiosis*, **22**: 293-302.

Arz, H.W., Gerhardt, S., Pätzold, J. and Röhl, U. 2001 Millennial-scale changes of surface-and deep-water flow in the western tropical Atlantic linked to the Northern Hemisphere high-latitude climate during the Holocene. *Geology*, **29**: 239-242.

Bacon, S. 1998 Decadal variability in the outflow from the Nordic Seas to the deep Atlantic Ocean. *Nature*, **394**: 871-874.

Bacon, S. and Carter, D.J.T. 1993 A connection between mean wave height and atmospheric-pressure gradient in the North Atlantic. *International Journal of Climatology*, **13(4)**: 423-436.

Bailey, R.H. 1975 Lichen propagule dispersal and establishment In: Brown, D.H., Hawksworth, D.L. and Bailey, R.H.(eds) *Lichenology: progress and problems*. Academic Press, London.

Ballantyne, C.K. 1987 The present-day periglaciation of upland Britain. In: Boardman, J. (ed) *Periglacial processes and landforms in Britain and Ireland*. Cambridge University Press: 113-126.

Ballantyne, C.K. 1989 The Loch Lomond readvance on the Isle of Skye, Scotland: glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science*, **4**: 95-108.

Ballantyne, C.K. 1990 The Holocene glacial history of Lyngshalvøya, northern Norway: chronology and climatic implications. *Boreas*, **19: 93-117.**

Ballantyne, C.K. and Harris, C. 1994 *The periglaciation of Great Britain*. Cambridge University Press.

Ballantyne, C.K. and Matthews, J.A. 1982 The development of sorted circles on recently deglaciated terrain, Jotunheimen, Norway. *Arctic and Alpine Research*, **14**: 341-354.

Bardarsson, G.G. 1934 *Islands Gletscher. Beiträge zur Kenntnis der Gletscherbewegungen und Schwankungen auf Grund alter Quellenschriften und neuster Forschung.* Reykjavik, Visindafélag Islendinga (Societas Scientarum Islandica) **XVIII.**

Barkman, J.J. 1958 *Phytosociology and ecology of cryptogamic epiphytes.* Van Gorcum, Assen.

Barlow, L.K., Sadler, J.P., Ogilvie, A.E.J., Buckland, P.C., Amorosi, T., Ingimundarson, J.H., Skidmore, P., Dugmore, A.J. and McGovern, T.H. 1997 Interdisciplinary investigations of the end of the Norse Western settlement in Greenland. *The Holocene*, **7(4)**: 489-501.

Barry, R.G. 1995 *Mountain weather and climate*. Routledge.

Barry, R.G. and Chorley, R.J. 1992 *Atmosphere, Weather and Climate*. 6th Edition, Routledge, London.

Bell, W.T. and Ogilvie, A.E.J. 1978 Weather compilations as a source of data for the reconstruction of European climate during the medieval period. *Climatic Change*, **1**: 331-348.

Benedict, J.B. 1967 Recent glacial history of an Alpine area in the Colorado Front Range, USA. 1: Establishing a lichen growth curve. *Journal of Glaciology*, **6**: 817-832.

Benedict, J.B. 1976 Frost creep and gelifluction features: a review. *Quaternary Research*, **6**: 55-76.

Benedict, J.B. 1985 Arapaho Pass: Glacial geology and archaeology at the crest of the Colorado Front Range. *Center for Mountain Archaeology, Ward, Colorado, Research Reports*, **3**:1-197.

Benedict, J.B. 1988 Techniques in lichenometry: identifying the yellow Rhizocarpons. *Arctic and Alpine Research*, **20**: 285-291.

Benedict, J.B. 1990a Experiments on lichen growth. I. Seasonal patterns and environmental controls. *Arctic and Alpine Research*, **22(3)**: 244-254.

Benedict, J.B. 1990 b Lichen mortality due to late lying snow: Results of a transplant study. *Arctic and Alpine Research* **22(1)**: 81-89.

Benedict, J.B. 1991 Experiments on lichen growth. II. Effects of a seasonal snow cover. *Arctic and Alpine Research*, **23(2)**: 189-199.

Benedict, J.B. 1993 A 2000-year lichen-snowkill chronology for the Colorado Front Range, USA. *The Holocene*, **3(1)**: 27-33.

Benedict, J.B. 1999 Effects of changing climate on game-animal and human use of the Colorado High Country (U.S.A) since 1000BC. *Arctic, Antarctic and Alpine Research,* **31(1)**: 1-15.

Bergthorsson, P. 1969 An estimate of drift ice and temperature in Iceland in 1000 years. *Jökull.* **19**: 94-101.

Beschel, R.E. 1950 Flechten als Altersmassstab rezenter Moränen. *Zeitschrift fur Gletscherkunde und Glazialgeologie,* **1(2)**: 152-161.

Beschel, R.E. 1954 Eine flechte als Niederschlagsmesser. *Wetter und Leben,* **6**: 56-60.

Beschel, R.E. 1956 *Lichenometrie im Gletschervorfeld.* Jahrbuch 1957, Ver. zum Schutze der Alpenpflanzen und Tiere, Munchen, **22**: 164-185.

Beschel, R.E. 1957 A project to use lichens as indicators of climate and time. *Arctic,* **10 (1)**: 60.

Beschel, R.E. 1958 Lichenometrical studies in West Greenland. *Arctic,* **11**: 254

Beschel, R.E. 1961 Dating rock surfaces by lichen growth and its application to glaciology and physiography (lichenometry). - *In:* G. O. Raasch (ed.): *Geology of the Arctic (Proceeding of the First International Symposium on Arctic Geology), Vol. 2.* University of Toronto Press, Toronto: 1044-1062.

Beschel, R.E. 1973 Lichens as a measure of the age of recent moraines. *Arctic and Alpine Research,* **5**: 303-309. (Translation, by William Barr, of Beschel, R.E. 1950).

Beschel, R.E. and Weidick, A. 1973 Geobotantical and geomorphological reconnaissance in West Greenland. *Arctic and Alpine Research,* **5**(4): 311-319.

Bianchi, G. and McCave, N. 1999 Holocene periodicity in North Atlantic climate and deep-ocean flow south of Iceland. *Nature,* **397**: 515-517.

Bickerton, R.W. and Matthews, J.A. 1992 On the accuracy of lichenometric dates: An assessment based on the 'Little Ice Age' moraine sequence at Nigardsbreen, southern Norway. *The Holocene,* **2**: 227-237.

Birkeland, P.W. 1973 The use of relative dating methods in a stratigraphic study of rock glacier deposits, Mt Sopris, Colorado. *Arctic and Alpine Research,* **5**: 401-416.

Birkenmajer, K. 1981 Lichenometric dating of raised marine beaches at Admiralty Bay, King George Island (South Shetland Islands, West Antartica). *Bulletin de l'Académie Polonaise des Sciences, Série des Sciences de la Terre,* **27**: 119-127.

Björnsson, H. 1971 Bægisárjökull, north Iceland: Results of glaciological investigations 1967-1968. Part 1: Mass balance and general meteorology. *Jökull,* **21**: 1-23.

Björnsson, H. 1975 Subglacial water reservoirs, jökulhlaups and volcanic eruptions. *Jökull.* **26**:40-50.

Björnsson, H. 1979 Glaciers in Iceland. *Jökull,* **29**: 74-80.

Björnsson, H. 1988 *Hydrology of Ice Caps in Volcanic Regions.* Reykjavik, Visindafélag Íslendinga. 139 pp.

Björnsson, H., Palsson, F., Gudmundsson, M, and Haraldsson, H. 1998 Mass balance of western and northern Vatnajökull, Iceland, 1991-1995. *Jökull,* **45**: 35-57.

Black, T.A. 1990 *The Holocene fluctuations of the Kviárjökull Glacier, southeastern Iceland.* Unpublished MSc Thesis, University of Colorado.

Bliss, L.C. 1956 A comparison of plant development in microenvironments of arctic and alpine tundras. *Ecological Monographs*, **26**: 303-337.

Bliss, L.C. and Hadley, E.B. 1964 Photosynthesis and respiration of alpine lichens. *American Journal of Botany*, **51**: 870-874.

Bond, G., Showers, W., Chesby, M., Lotti, M., Almasi, P., deMenocal, P., Priore, P., Cullen, H., Hajdas, I. and Bonani, G. 1997 A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science*, **278**: 1257-1266.

Bornefeldt, F. and Österborg, M. 1958 Lavarter som hjälpmedem för Datering av Ändmoräner vid Norska Glaciärer. *Stokholms Högskola, Geografiska Proseminariet*, **403**:1-37. (Unpublished seminar paper.)

Boulton, G.S. 1986 Push moraines and glacier contact fans in marine and terrestrial environments. *Sedimentology*, **33**: 677-698.

Boulton, G.S. and Clark, C.D. 1990 A highly mobile Laurentide ice-sheet revealed by satellite images of glacial lineations. *Nature*. **346**: 813-817.

Boulton, G.S., Mason, P., Ballantyne, C., Karlén, W., Matthews, J. and Nesje 1997 Holocene glacier variations in Scandinavia. In (Frenzel, B., Boulton, G., Glaser, B and Huckriede, U. (eds.) *Glacier Fluctuations during the Holocene*. Gustav Fischer Verlag, Stuttgart, ESF Project Special Issue: 5-33.

Bradley, R.S. 2000 Past global changes and their significance for the future. *Quaternary Science Reviews*, **19**: 391-402.

Bradley, R.S. and Jones, P.D. 1993 'Little Ice Age' summer temperature variations: their nature and relevance to recent global warming trends. *The Holocene*, **3**: 367-376.

Bradley, R.S., Kelly, P.M., Jones, P.D., Diaz, H.F. and Goodess, C. 1985 A climatic data bank for the Northern Hemisphere land area 1851-1950. *Technical Report no. TRO17 U.S Department of Energy Carbon Dioxide Research Division, Washington D.C.* 355pp.

Bradwell, T. 1998 Climate induced twentieth-century glacier fluctuations in southeast Iceland. *Fróðskaparrit*, **46**: 319-332.

Bradwell, T. *in press*. A new lichenometric dating curve for southeast Iceland. *Geografiska Annaler*.

Brázdil, R. 1988 Variation of air temperature and atmospheric precipitation in the region of Svalbard and of Jan Mayen. In: Gregory, S. (ed) *Recent climatic change – a regional approach*. Belhaven Press, London.

Broadbent, N.D. and Berqvist, 1986 Lichenometric chronology and archaeological features on raised beaches: preliminary results from the Swedish north Bothnian coastal region. *Arctic and Alpine Research*, **18**: 297-306.

Brodo, I.M. 1965 Studies on growth rates of corticolous lichens on Long Island, New York. *The Bryologist*, **68**(4): 451-456.

Broecker, W.S. 1990 Salinity history of the Northern Atlantic during the last deglaciation. *Paleoceanography*, **5**(4): 459-467.

Broecker, W.S. and Kennett, J.P. 1989 The routing of Laurentide ice-sheet meltwater during the Younger Dryas cold event. *Nature*, **341**: 318.

Broecker, W.S., Peteet, D.M. and Rind, D. 1985 Does the ocean-atmosphere system have more than one stable mode of operation? *Nature*, **315**: 21-26.

Bryan, F.O. 1986 High latitude salinity effects and interhemispheric thermohaline circulation. *Nature*. **323**: 301.

Bryant, E. 1997 *Climate Process and Change*. Cambridge University Press.

Buckland, P.I. 1994 Lichenometry and soil erosion in northwest Iceland. *In*: Stötter, J. and Wilhelm, F. *Environmental Change in Iceland*. Münchener Geographische Abhandlungen, Reihe B, **B12**: 31-40.

Bull, W.B. 1996 Dating San Andreas fault earthquakes with lichenometry. *Geology*, **24**: 111-114.

Bull, W.B. and Brandon, M.T. 1998 Lichen dating of earthquake-generated regional rockfall events, Southern Alps, New Zealand. *GSA Bulletin*, **110**: 60-84.

Bull, W.B., King, J., Kong, F., Moutoux, T. and Phillips, W.M. 1994 Lichen dating of coseismic landslide hazards in alpine mountains. *Geomorphology*, **10**: 253-264.

Bull, W.B., Schlyter, P. and Brogaard, S. 1995 Lichenometric analysis of the Kärkerieppe slush-avalanche fan, Kärkevagge, Sweden. *Geografiska Annaler*, **77A**: 231-240.

Burroughs, W.J. 1994 *Weather Cycles : real or imaginary?* Cambridge University Press.

Burrows, C.J. 1975 Late Pleistocene and Holocene moraines of the Cameron Valley, Arrowsmith Range, Canterbury, New Zealand. *Arctic and Alpine Research*, **7**: 125-140.

Burrows, C.J. and Lucas, J. 1967 Variations in two New Zealand glaciers during the past 800 years. *Nature*, **216**: 467-468.

Butin, H. 1954 Physiologisch-ökologische Untersuchungen uber den Wasser-haushalt und die Photosynthese bei Flechten. *Biol. Zentr.*, **73**: 459-502.

Caine, N. 1963 The origin of sorted stripes in the Lake District, northern England. *Geografiska Annaler*, **45A**: 172-179.

Calkin, P.E. and Ellis, J.M. 1980 A lichenometric dating curve: its application to Holocene glacier studies in Central Brooks Range, Alaska. *Arctic and Alpine Research*, **12**: 245-264.

Carrara, P.E. and Andrews, J.T. 1973 Problems and application of lichenometry to geomorphic studies, San Juan Mountains, Colorado. *Arctic and Alpine Research*, **5**: 373-384.

Caseldine, C.J. 1983 Resurvey of the margins of Gljúfurárjökull and the chronology of recent deglaciation. *Jökull*, **33**: 11-22.

Caseldine, C.J. 1985 The extent of some glaciers in Northern Iceland during the Little Ice Age and the nature of recent deglaciation. *The Geographical Journal*, **151(2)**: 215-227.

Caseldine, C.J. 1987 Neoglacial glacier variations in Northern Iceland: Examples from the Eyjafjörður area. *Arctic and Alpine Research*, **19(3)**: 296-304.

Caseldine, C.J. 1990 A review of dating methods and their application in the development of a chronology of Holocene glacier variations in northern Iceland. *In*: Caseldine, C., Häberle, T., Kugelmann, O., Münzer, U., Stötter, J. and Wilhelm, F. *Gletscher und Landschaftsgeschichtliche Untersuchungen in Nordisland.* Münchener Geographische Abhandlungen, **B8**: 59-82.

Caseldine, C.J. 1991 Lichenometric dating, lichen population studies and Holocene glacial history in Tröllaskagi, Northern Iceland. *In*: Maizels, J.K. and Caseldine, C.J. (Eds) *Environmental Change in Iceland: Past and Present*: 219-233. Kluwer, Dordrecht.

Caseldine, C. and Baker, A. 1998 Frequency distribution of *Rhizocarpon geographicum* s.l., modeling, and climate variation in Tröllaskagi, northern Iceland. *Arctic and Alpine Research*, **30(2)**: 175-183.

Caseldine, C. and Stötter, J. 1993 'Little Ice Age' glaciation of Tröllaskagi peninsula, northern Iceland: Climatic implications for reconstructed equilibrium-line altitudes (ELAs). *The Holocene*, **3**: 357-366.

Cernohrsky, Z. 1977 *Rhizocarpon* Ram. Em. Th.Fr. *In* Poelt, J. and Vezda, A., (eds.) *Bestimmungschlüssel europäischer Flechten*, Erganzungsheft 1, Vaduz: J.Cramer: 217-231.

Chapman, W.L. and Walsh, J.E. 1993 Recent variations of sea-ice and air temperature in high latitudes. *Bulletin American Meteorological Society*, **74**: 33-47.

Childress, S. and Keller, J.B. 1980 Lichen growth. *Journal of Theoretical Biology*, **82**: 157-65.

Clark, B.M., Mangelson, N.F., St.Clair, L.L., Bench, G.S. and Southon, J.R. 2000 Measurement of age and growth rate in the crustose saxicolous lichen *Caloplaca trachyphylla* using ^{14}C accelerator mass spectrometry. *Lichenologist*, **32**: 399-403.

Conrad, V. 1946 Usual formulas of continentality and their limits of validity. *Transactions of the American Geophysical Union*, **27**: 663-664.

Cook, E.R., D'Arrigo, D. and Briffa, K.R. 1998 A reconstruction of the North Atlantic Oscillation using tree ring chronologies from North America and Europe. *The Holocene*, **8**: 9-17.

Cook-Talbot, J.D. 1991 Sorted circles, relative-age dating and palaeoenvironmental reconstruction in an alpine periglacial environment, eastern Jotunheimen, Norway: lichenometric and weathering-based approaches. *The Holocene*, **1(2)**: 128-141.

Coxson, D.S. and Kershaw, K.A. 1983a The ecology of *Rhizocarpon superficiale*. I. The rock surface boundary-layer microclimate. *Canadian Journal of Botany*, **61**: 3009-3018.

Coxson, D.S. and Kershaw, K.A. 1983b The ecology of *Rhizocarpon superficiale*. II. The seasonal response of net photosynthesis and respiration to temperature, moisture, and light. *Canadian Journal of Botany*, **61**: 3019-3030.

Coxson, D.S., Brown, D. and Kershaw, K.A. 1983 The interaction between CO_2 diffusion and the degree of thallus hydration in lichens: some further comments. *New Phytologist*, **93**: 247-260.

Crittenden, P.D. 1998 Nutrient exchange in an Antarctic macrolichen during summer snowfall-snow melt events. *New Phytologist*, **139(4)**: 697-707.

Crittenden, P.D. and Kershaw, K.A. 1979 Studies on lichen-dominated systems. XXII. The environmental control of nitrogenase activity in *Stereocaulon paschale* in spruce-lichen woodland. *Canadian Journal of Botany*, **57**: 236-254.

Curry, W.B. and Oppo, D.W. 1997 Synchronous, high frequency oscillations in tropical sea surface temperatures and North Atlantic Deep Water production during the last glacial cycle. *Paleoceanography*, **12**: 1-14.

D'Arrigo, R.D., Cook, E.R., Jacoby, G.C. and Briffa, K.R. 1993 NAO and sea-surface temperature signatures in tree-ring records from the north-Atlantic sector. *Quaternary Science Reviews*, **12**: 431-440.

Dansgaard, W., Clausen, H.B., Gundestrup, N., Hammer, C.U., Johnsen, S.F., Kristinsdottir, P.M. and Reeh, N. 1982 A new Greenland deep ice core. *Science*, **218**: 1273-1277.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N., Hammer, C.U. and Oeschger, H., 1984 North Atlantic climatic oscillations revealed by deep Greenland ice cores. In: Hansen, J.E. and Takahashi, T. (eds) *Climate processes and sensitivity*, **29**, Geophysical Monograph series A.M. Geophysical Union, Washington D.C.: 288-298.

Darbishire, O.V. 1905 The lichens of the South Orkneys. *Transactions and Proceedings of the Botanical Society Edinburgh*, **23**: 1-6.

Dawson, A.G., Matthews, J.A. and Shakesby, R.A. 1986 A catastrophic landslide (Sturzstrom) in Verkilsdalen, Rondane National Park, southern Norway. *Geografiska Annaler*, **68**(A): 77-87.

Degelius, G. 1964 Biological studies of epiphytic vegetation on twigs of *Fraxinus excelsior*. *Acta Horti Gotoburg*, **27**: 11-55.

Delworth, T., Manabe, S. and Stouffer, R.J. 1993 Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model. *Journal of Climate*, **6**: 1993-2011.

Denton, G.H. and Karlén, W. 1973 Lichenometry: its application to Holocene moraine studies in southern Alaska and Swedish Lappland. *Arctic and Alpine Research*, **5**: 347-372.

Deser, C. and Blackmon, M.L. 1993 Surface climate variations over the North Atlantic Ocean during winter: 1900-1989. *Journal of Climate*, **6**: 1743-1753.

Dickson, R., Lazier, J., Meincke, J., Rhines, P. and Swift, J. 1996 Long-term co-ordinated changes in the convective activity of the North Atlantic. *Progress in Oceanography*, **38**: 241-295.

Dickson, R.R. and Brown, J. 1994 The production of North Atlantic Deep Water: Sources, rates and pathways. *Journal of Geophysical Research*, **99**(C6): 12,319-12,341.

Donner, J., Eronen, M. and Jungner, H. 1977 The dating of Holocene relative sea level changes in Finnmark, North Norway. *Norsk Geografisk Tidsskrift*, **31**: 103-108.

Douglas, T.D. and Harrison, S. 1996 Turf-banked terraces in Öræfi, southeast Iceland: Morphometry, rates of movement and environmental controls. *Arctic and Alpine Research*, **28**: 228-236.

Dugmore, A.J. 1987 *Holocene Glacier Fluctuations around Eyjafjallajökull, south Iceland: A tephrochronological study.* Unpublished Ph.D. Thesis, Department of Geography, University of Aberdeen, Scotland.

Dugmore, A.J. 1989 Tephrochronological studies of Holocene glacier fluctuations in south Iceland. *In*: Oerlemans, J. (Ed.) *Glacier Fluctuations and Climatic Change*. Kluwer Academic Publishers, Dordrecht: 37-55.

Dugmore, A.J. and Buckland, P.C. 1991 Tephrochronology and late Holocene soil erosion. In: Maizels, J.K. and Caseldine, C. (eds) *Environmental change in Iceland: Past and present*. Dordrecht: 37-55.

Dugmore, A.J. and Erskine, C.C. 1994 Local and regional pattern of soil erosion in southern Iceland. In: Stötter, J. and Wilhelm, F. (eds) *Environmental change in Iceland. Münchener Geographische Abhandlungen,*. **B12**: 41-63.

Dugmore, A.J. and Kirkbride, M.K. *submitted* Holocene glacier fluctuations of Hrútfell, Central Iceland. *Journal of Quaternary Science.*

Dugmore, A.J. and Sugden, D.E. 1991 Do the anomalous fluctuations of Sólheimajökull reflect ice-divide migration? *Boreas*, 20: 105-113.

Dugmore, A.J., Cook, G.T., Shore, J.S., Newton, A.J., Edwards, K.J. and Larsen, G. 1995 Radiocarbon dating tephra layers in Britain and Iceland. *Radiocarbon*, **37**: 379-388.

Duplessy,J.-C., Shackleton, N.J., Fairbanks, R.G., Labeyrie, L., Oppo, D., Kallel., N. 1988. Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation.

Duplessy, J-C., Ivanova, E., Murdmaa, I., Paterne, M. and Labeyrie, L. 2001 Holocene paleoceanography of the northern Barents Sea and variations of the northward heat transport by the Atlantic Ocean. *Boreas*, **30**: 2-16.

Einarsson, M.A. 1991 Temperature conditions in Iceland 1901-1990. *Jökull*, **41**: 1-42

Einarsson, Th. 1960 Geologie von Hellisheiði. *Sönderöffentlichungen des Geologischen Institutes der Universitat Köln.*, 5. 55pp.

Einarsson, Th. 1963 Pollen-analytical studies on vegetation and climate history of Iceland in late and post-glacial times IN: Löve, A. and Löve, D. (eds) *North Atlantic biota and their history*. Pergamon: 255-365.

Einarsson, Th. 1964 Aldursákvardanir á fornskeljum. *Náttúrufraeðingurinn*, **34**: 127-136.

Einarsson, Th. 1967 Zu der Ausdehung der weichselzeitlichen verisung Nordislands. *Sönderöffentlichungen des Geologischen Institutes der Universitat Köln*, **1**: 167-173.

Einarsson, Th. 1968 *Jarðfræði*. Mál og Menning. Reykjavik.

Einarsson, Th. 1979 The deglaciation of Iceland. *Norsk Geologisk Forening Medlemsblad*, **13** (18).

Einarsson, Th. 1994 *Jarðfræði. Myndun og mótun land.* Mál og Menning. Reykjavik.

Ensgraber, A. 1954 Über den Einfluss der Antrocknung auf die Assimilation und Atmung von Moosen und Flechten. *Flora (Jena)*, **141**: 432-475.

Eriksson, H.H. 1932 *Observations and measurements of some glaciers in Auster-Skaftafellssysla, in the summer of 1930.* Visindafelej Islandinga.

Erikstad, L. and Sølid, J.L. 1986 Neoglaciation in South Norway using lichenometric methods. *Norsk Geografisk Tidsskrift*, **40**: 85-103.

Ertl, L. 1951 Über die Lichtverhaltnisse in Laubflechten. *Planta*, **39(3)**: 245-270.

Evans, D.J.A., Archer, S. and Wilson, D.J.H. 1999 A comparison of the lichenometric and Schmidt hammer dating techniques based on data from the proglacial areas of some Icelandic glaciers. *Quaternary Science Reviews*, **18**: 13-41.

Eyles, N. 1978 Rock glaciers in Esjufjöll Nunatak area, southeast Iceland. *Jökull*, **28**: 53-60.

Eythorsson, J. 1935 On the variation of glaciers in Iceland. *Geografiska Annaler*, **1-2**: 121-137.

Eythorsson, J. 1942 Breytingar islenzkra jökla á timabilinu 1930-1941 (i metrum). *Náttúrufræðingurinn.* **12**: 125-128.

Eythorsson, J. 1963 Variations of Icelandic glaciers 1931-1960. *Jökull*, **13**: 31-33.

Eythorsson, J. and Sigtrysson, H. 1971 The climate and weather of Iceland. *The Zoology of Iceland*, **1**: 1-62.

Farrar, J.F. 1976 Ecological physiology of the lichen *Hypogymnia physodes*. I. Some effects of constant water saturation. *New Phytologist*, **77**: 93-103.

Farrar, J.F. and Smith, D.C. 1976 Ecological physiology of the lichen *Hypogymnia physodes*. III. The importance of the rewetting phase. *New Phytologist*, **77**: 127-134.

Ferry, B. W., Baddeley M. S. and Hawksworth, D.L. (eds.) 1973 *Air Pollution and Lichens*. University of Toronto Press, Toronto.

Fisher, P.J. and Proctor, M.C.F. 1978 Observations on the season's growth in *Parmelia caperata* and *Parmelia sulcata* in South Devon. *Lichenologist*, **10**: 81-89.

Flock, J.W. 1974 Lichen-bryophyte distribution along a snow-cover–soil-moisture gradient, Niwot Ridge, Colorado. *Arctic and Alpine Research*, 10: 31-47.

Follmann, G. 1961 Estudios liquenometricos en los monumentos prehistoricos de la Isla de Pascua. *Revista Universitaria – Anales de la Academia Chilena de Ciencias Naturales*, **24**: 149-154.

Folland, C.K., Karl, T.R. and Vinnikov, K.Ya. 1990 Observed climate variations and change. In: Houghton, J.T., Jenkins, G.J. and Ephraums, J.J. (eds) *Climate change: The IPCC scientific assessment*. Cambridge University Press: 195-238.

Frey, E. 1959 Flechtenflora und -vegetation des Nationalparks im Unterengadin. II. Teil: Die Entwicklung der Flechtenvegetation aud photogrammetrisch kontrollierten Dauerflachen. *Ergebnisse der wissenschaftlichen Untersuchungen des schweizerischen Nationalparks*, **6(41)**: 237-319.

Fristrup, B. 1951 Wind erosion in the Arctic deserts. *Geografisk Tidsskrift*, **52**: 51-65.

Furbish, D.J. and Andrews, J.T. 1984 The use of hyposometry to indicate long term stability and response of valley glaciers to changes in mass transfer. *Journal of Glaciology*, **30**: 199-211.

Galun, M. (ed.) 1988a *CRC Handbook of Lichenology. Volume I*. CRC Press, Inc., Boca Raton. 297 pp.

Galun, M. (ed.) 1988b *CRC Handbook of Lichenology. Volume II*. CRC Press Inc., Boca Raton. 181 pp.

Galun, M. (ed.) 1988c *CRC Handbook of Lichenology. Volume III*. CRC Press, Inc., Boca Raton. 147 pp

Gams, H. 1932 Die Klimatische Bergenzung von Pflanzenarealen und die Verteilung der hygrischen Kontinentalität in den Alpen. *Zeitschrift der Gesellschaft f ür Erdkunde zu Berlin, 1931/32.*

Geirsdóttir, A., Harðardóttir, J. and Andrews, J.T. 2000 Late-Holocene terrestrial glacial history of Miki and I.C. Jacobsen fjords, East Greenland. *The Holocene*. **10(1)**: 123-134.

Gellatly, A.F. 1982 Lichenometry as a relative-age dating method in Mount Cook National Park, New Zealand. *New Zealand Journal of Botany*. **20**: 343-353.

Gilbert, O.L. 1984 The lichens of Choire Garbh. *New Scientist*, **101(1398)**: 42-43.

Gilbert, O.L., Fryday, A.M., Giavarini, V.J. and Coppins, B.J. 1992 The lichen vegetation of high ground in the Ben Nevis Range, Scotland. *Lichenologist*, **24(1)**: 43-56.

Godske, L.L. 1944 The geographical distribution in Norway of certain indices of humidity and oceanicity. *Bergens Museums Arbok, Naturvitenskapelig*, **8**.

Goldthwaite, R.P. 1976 Frost-sorted patterned ground: a review. *Quaternary Research*, **6**: 27-35.

Gorcyznski, W. 1920 Sur la calcul du degré de continent et on application dans la climatologie. *Geografiska Annaler*, **2**: 234-331.

Gordon, J.E. and Sharp, M. 1983 Lichenometry in dating recent glacial landforms and deposits, southeast Iceland. *Boreas*, **12**: 191-200.

Gribbon, P.W.F. 1970 Frontal recession of Sermikavsak, West Greenland. *Journal of Glaciology*, **9**: 279-282.

Griffey, N.J. 1978 Lichen growth on supraglacial debris and its implications for lichenometric studies. *Journal of Glaciology*, **20(82)**: 163-172.

Grönvold, K., Óskarsson, N., Johnsen, S.J., Clausen, H.B., Hammer, C.U., Bond, G. and Bard, E. 1995 Ash layers from Iceland in the Greenland Grip ice core correlated with oceanic and land sediments. *Earth and Planetary Science Letters*, **135**: 149-155.

Grove, J.M. 1988 *The Little Ice Age*. Methuen, London.

Gudmundsson, A. 1995 Berghlaup eða urdarjöklar? *Náttúrufræðingurinn*, **64**: 177-186.

Gudmundsson, A. 2000 The relict periglacial face of the Héraðsflói area - east Iceland. *Iceland 2000: Modern processes and past environments*. International Conference, Keele University (abstract volume): 44-47.

Gudmundsson, H. J. 1997 A review of the Holocene environmental history of Iceland. *Quaternary Science Reviews*, **16**: 81-92.

Gudmundsson, H.J. 1998a *Holocene Glacier Fluctuations and Tephrochronology of the Öræfi district, Iceland*. Unpublished Ph.D. Thesis, University of Edinburgh.

Gudmundsson, H.J. 1998b Holocene glacier fluctuations of the Eiríksjökull ice cap, west central Iceland. *Jökull*, **46**: 17-28.

Häberle, T. 1991 Holocene glacial history of the Hörgárdalur area, Tröllaskagi, northern Iceland. *In*: Maizels, J.K. and Caseldine, C. (Eds.) *Environmental Change in Iceland: Past and Present*. Kluwer, Dordrecht: 193-202.

Häberle, T. 1994 Glacial, Late Glacial and Holocene history of the Hörgárdalur area, Tröllaskagi, northern Iceland. *In*: Stötter, J and Wilhelm, F. (Eds.) *Environmental Change in Iceland. Münchener Geographische Abhandlungen*, **B12**: 41-63.

Hacquebord, L 1984 The history of early Dutch whaling: a study from the ecological angle. In: Van Marken, A. (ed) *Arctic whaling*. Groningen, University of Groningen: 135-148.

Haeberli, W., King, L. and Flotron, A. 1979 Surface movement and lichen-cover studies at the active rock glacier near the Grubengletscher, Wallis, Swiss Alps. *Arctic and Alpine Research*, **11**: 421-441.

Haflidason, H., Eirikson, J. and Van Kreveld, S. 2000 The tephrochronology of Iceland and the North Atlantic region during the Middle and Late Quaternary : a review. *Journal of Quaternary Science*, **15**: 3-22.

Haines-Young, R.H. 1983 Size variation of *Rhizocarpon* on moraine slopes in southern Norway. *Arctic and Alpine Research*, **15**: 295-305.

Hakulinen, R. 1966 Uber die Wachtumsgeschwindigkeit einiger Laubflechten. *Annales Botanici Fennici,* **3**: 167-179.

Hale, M.E. 1954 First report on lichen growth rate and succession at Aton Forest, Connecticut. *The Bryologist,* **57(3)**: 244-247.

Hale, M.E. 1959 Studies on lichen growth rate and succession. *Bulletin of the Torrey Botanical Club,* **86(2)**: 126-129.

Hale, M.E. 1967 *The Biology of Lichens.* Edward Arnold, London, pp176.

Hale, M.E. 1970 Single-lobe growth-rate patterns in the lichen *Parmelia caperata. The Bryologist,* **73**: 72-81.

Hale, M.E. 1974 Growth. *In*: Ahmadjian, V. and Hale, M.E., (Eds.) *The Lichens.* London, Academic Press: 473-492.

Harvey, A.M., Alexander, R.W. and James, P.A. 1984 Lichens, soil development and the age of Holocene valley floor landforms, Howgill Fells, Cumbria. *Geografiska Annaler,* **66(A)**: 353-366.

Hausmann, E.H. 1948 Measurements of the annual growth rate of two species of rock lichens. *Bulletin of the Torrey Botanical Club*, **75**: 116-117.

Hawkins, F. 1985 Equilibrium-line altitudes and palaeoenvironments in the Merchants Bay area, Baffin Island, NWT, Canada. *Journal of Glaciology*, **31**: 205-213.

Haworth, L.A., Calkin, P.E. and Ellis, J.M. 1986 Direct measurement of lichen growth in the central Brooks Range, Alaska, U.S.A., and its application to lichenometric dating. *Arctic and Alpine Research,* **18**: 289-296.

Henderson, E. 1819 *Iceland, or the journal of a residence in that island during the years 1814-1815.* Edinburgh, 376pp.

Hill, D.J. and Smith, D.C. 1972. Lichen physiology XII. The inhibition technique. *New Phytologist*, **71**: 15-30.

Hill, D.J. 1981 The growth of lichens with special reference to the modelling of circular thalli. *Lichenologist,* **13**: 265-287.

Hjartarson, A., Sigurdsson, F. and Hafstad, Th.H. 1981 *Vatnabúskapur Austeurlands III* (English summary). Mimeographed Report to the English Authority of Iceland, OS 810006/VOD 04.

Hjartarson, Á. and Ingolfsson, Ó. 1988 Preboreal glaciation of southeastern Iceland. *Jökull*, **38**: 1-16.

Hjört, C., Ingolfsson, Ó. and Norðdahl, H. 1985 Late Quaternary geology and glacial history of Hornstrandir, Northwest Iceland. *Jokull*, **35**: 9-29.

Hoinkes, H.C. 1968 Glacier variation and weather. *Journal of Glaciology*, **7**: 3-19.

Hooker, T.N. 1980 Factors affecting the growth of Antarctic crustose lichens. *British Antarctic Survey Bulletin*, **50**: 1-19.

Hooker, T.N. and Brown, D.H. 1977 A photographic method for accurately measuring the growth of crustose and foliose saxicolous lichens. *Lichenologist*, **9**: 65-75.

Hoppe, G. 1968 Grimsey and the maximum extent of the last glaciation in Iceland. *Geografiska Annaler*, **50 (A)**: 16-24.

Hoppe, G. 1982 The extent of the last inland ice sheet of Iceland. *Jökull*, **32**: 3-11.

Houghton, J.T., Meira Filho, L.G., Callander, B.A., Harris, N., Kattenberg, A., Maskell, K. (eds) 2000 *Climate Change 2000*. Cambridge University Press, Cambridge.

Howell, F.W.W. 1893 *Icelandic pictures*. Religious Tract Society, London.

Huang, J., Higuchi, K. and Shabbar, A. 1998 The relationship between the North Atlantic Oscillation and El Niño-Southern Oscillation. *Geophysical Research Letters*, **25(14)**: 2707-2710.

Humlum, O. 1998a Climatic forcing of Late Holocene geomorphic activity, Faeroe Islands. *Froðskaparrit*, **46**: 333-352.

Humlum, O. 1998b Rock glaciers on the Faeroe Islands, the North Atlantic. *Journal of Quaternary Science*, **13(4)**: 293-307.

Hurrell, J.W. 1995 Decadal trends in the North Atlantic Oscillation regional temperatures and precipitation. *Science*, **269**: 676-679.

Hurrell, J.W. and van Loon, H. 1997 Decadal variations in climate associated with the North Atlantic Oscillation. *Climate Change*, **36**: 301-326.

Imbrie, J. 1982 Astronomical theory of the Pleistocene ice-ages – a brief historical review. *Icarus*, **50**: 408-422.

Imbrie, J. and 17 others. 1992 On the structure and origin of major glaciation cycles. 1. linear reponses to Milankovitch forcing. *Paleoceanography*, **7**: 701-738.

Ingolfsson, Ó. 1985 Late Weichselian glacial geology of the lower Borgarfjörður region, western Iceland: A preliminary report. *Arctic*, **38**: 210-213.

Ingolfsson, Ó. 1987 Investigation of the Late Weichselian glacial history of the lower Borgarfjörður region, western Iceland. *Lundqua Thesis* **19**, Department of Quaternary Geology, University of Lund.

Ingolfsson, Ó. 1988 Glacial history of the lower Borgarfjörður area, western Iceland. *Geologiska Föreningens i Stockholm Förhandlingar*, **110**: 293-309.

Ingolfsson, Ó. and Norðdahl, H. 1994 A review of the environmental history of Iceland, 13000-9000 yr BP. *Journal of Quaternary Science*, **9(2)**: 147-150.

Ingolfsson,Ó., Björck, S., Haflidason, H. and Rundgren, M. 1997 Glacial and climatic events in Iceland reflecting regional North Atlantic climatic shifts during the Pleistocene-Holocene transition. *Quaternary Science Reviews*, **16**: 1135-1144.

Innes, J.L. 1982 Lichenometric use of an aggregated *Rhizocarpon* 'species'. *Boreas*, **11**: 53-57.

Innes, J.L. 1983a Use of an aggregated *Rhizocarpon* 'species' in lichenometry: an assessment. *Boreas*, **12**: 183-190.

Innes, J.L. 1983b Size frequency distributions as a lichenometric technique: an assessment. *Arctic and Alpine Research*, **15**: 285-294.

Innes, J.L. 1983c The development of lichenometric dating curves for Highland Scotland. *Transactions of the Royal Society of Edinburgh Earth Sciences,* **74**: 23-32.

Innes, J.L. 1984 The optimal sample size in lichenometric studies. *Arctic and Alpine Research,* **16**: 233-244.

Innes, J.L. 1985 Lichenometry. *Progress in Physical Geography,* **9**: 187-254.

Innes, J.L. 1986 The size-frequency distributions of the lichens *Sporastatia testudinea* and *Rhizocarpon alpicola* through time at Storbreen, south-west Norway. *Journal of Biogeography,* **13**: 283-291.

Innes, J.L. 1988 The use of lichens in dating. *In:* M. Galun (ed.) *CRC Handbook of Lichenology. Volume III.* CRC Press, Inc., Boca Raton: 75-91.

Ives, J. 1956 *Öræfi, Southeast Iceland: An essay in regional geomorphology.* Unpublished Ph.D. Thesis, McGill University, Montreal, Canada.

Ives, J.D. 1962 Indications of recent extensive glacierization in north-central Baffin Island, N.W.T. *Journal of Glaciology,* **4(32)**: 197-205.

Ives, J.D. 1996. Glacier and climate reconstructions in southeast Iceland during the last two millennia. *Jahrbuch der Geographische Gesellschaft,* 59: 123-132.

Ives, J. and King, C.A.M. 1954 Geological observation on Morsàrjökull. Part 1: The Ogive banding. *Journal of Glaciology,* **2**: 423-428.

Ives, J. and King, C.A.M. 1955 Glaciological observations on Morsárjökull, southwest Vatnajökull. Part II: Regime of the glacier, present and past. *Journal of Glaciology,* **2(17)**: 477-482.

Jaksch, K. 1970 Beobachtungen in den Gletschervorfeldern des Sólheima- und Siðujökull im Sommer 1970. *Jökull,* **20**: 45-48.

Jaksch, K. 1975 Das Gletschervorfeld des Sólheimajökull. *Jökull,* **25**: 34-37.

Jennings, A.E. and Weiner, N.J. 1996 Environmental change in eastern Greenland during the last 1300 years: evidence from foraminifera lithofacies in Nansen Fjord, 68°N. *The Holocene,* **6**: 179-191.

Jennings, J.N. 1939 The glaciers of Jan Mayen, Appendix II In: King, A. The Imperial College of Science expedition to Jan Mayan Island. *Geographical Journal,* **94**: 128-131.

Jennings, J.N. 1951 Glacier retreat in Jan Mayen. *Journal of Glaciology,* **1**: 167-181.

Johannesson, T. 1985 On the ages of the two recent lava flows in Eyjafjöll and the late-glacial terminal moraines in south Iceland. *Jokull,* **35**: 95.

Johannesson, T. 1997 The response of two Icelandic glaciers to climatic warming computed with degree-day mass balance model coupled to a dynamic glacier model. *Journal of Glaciology,* **43**: 321-327.

Jóhannesson, T. and Sigurðsson, O. 1998 Interpretation of glacier variations in Iceland 1930-1995. *Jökull,* **45**: 27-33.

John, B.S. 1975 Moraine stages in the Reykjafjörður outlet glacier trough. *Durham University Vestfirðir Project, 1975 Fieldwork Report and Research Notes*: 46-54.

John, B.S. and Sugden, D.E., 1962 The morphology of Kaldalon, a recently deglaciated valley in Iceland. *Geografiska Annaler.* **XLIV**: 347-365.

Jonasson, C. 1991 Holocene slope processes of periglacial mountain areas in Scandanavia and Poland. *Uppsala Universitet, Naturgeografisk Institutuonen (UNIG) Report* **79**: 1-156.

Jonasson, C., Kot, M. and Kotarba, A. 1991 Lichenometrical studies and dating of debris flow deposits in the High Tatra Mountains, Poland. *Geografiska Annaler*, **73**(A): 141-146.

Jones, J.M. and Platt, R.B. 1969 Effects of ionizing radiation, climate, and nutrition on growth and structure of a lichen *Parmelia conspersa* (Ach.) Ach. *In:* Nelson, D. L. and Evans, F. C. (eds.) *Symposium on Radioecology*. Second National Symposium, Ann Arbor, Michigan, May 15-17, 1967: 111-119.

Jones, P.D., Raper, S.C.B., Bradley, R.S., Diaz, H.F., Kelly, P.M. and Wigley, T.M.L. 1986 Northern Hemisphere surface air temperature variations: 1851-1984. *Journal of Climate and Applied Meteorology*, **25**: 161-179.

Jones, P.D., Jonsson, T. and Wheeler, D. 1997 Extension to the North Atlantic Oscillation using early instrumental pressure observations from Gibraltar and southwest Iceland. *International Journal of Climatology*, **15**: 1433-50.

Jonsson, T. and Garðarsson, H. 2001 Early instrumental meteorological observations from Iceland. *Climatic Change*, **48**: 44-53.

Juckes, L.M. 1969 Weathering hollows in charnockite at Mannefaulknausane, Dronning Maud Land. *Bulletin of the British Antarctic Survey*, **22**: 97-98.

Kaldal, I. and Víkingsson, S. 1991 Early Holocene deglaciation in central Iceland. *Jökull*, **40**: 51-66.

Kamb, B. 1987 Glacier surge mechanism based on linked cavity configuration of the basal water conduit system. *Journal of Geophysical Research*, **92**: 9083-9100.

Kappen, L. 1973 Response to extreme environments. *In:* Ahmadjian, V. and Hale, M. E. (eds.) *The Lichens*. Academic Press, New York and London: 311-380.

Kappen, L. 1988 Ecophysiological relationships in different climatic regions. *In:* Galun, M. (ed.) *CRC Handbook of Lichenology. Volume II*. CRC Press, Inc., Boca Raton: 37-100.

Kappen, L. 1993 Plant activity under snow and ice, with particular reference to lichens. *Arctic*, **46**(4): 297-302.

Kappen, L. and Breuer, M. 1991 Ecological and physiological investigations in continental Antarctic cryptogams. II. Moisture relations and photosynthesis of lichens near Casey Station, Wilkes Land. *Antarctic Science*, **3**(3): 273-278.

Kappen, L., Schroeter, B., Scheidegger, C., Sommerkorn, M. and Hestmark, G. 1996 Cold resistance and metabolic activity of lichens below 0°C. *Advances in Space Research*, **18**(12): 119-128.

Karenlampi, L. 1971 Studies on the relative growth rate of some fruticose lichens. *Report of the Kevo Subarctic Research Station*, **7**: 33-39.

Karlén, W. and Kuylenstierna, J. 1996 On solar forcing of Holocene climate: evidence from Scandinavia. *The Holocene*, **6**: 359-366.

Karlén, W. 1979 Glacier fluctuations in the Svartisen area, northern Norway. *Geografiska Annaler*, **61**(A), 11-28.

Karlén, W. and Denton, G.H. 1975 Holocene glacial variations in Sarek National Park, northern Sweden. *Boreas,* **5**, 25-56.

Kelly, P., Goodess, C. and Cherry, B. 1987 The interpretation of the Icelandic sea ice record. *Journal of Geophysical Research*, **92**: 10843-19853.

Kerr, A. 1993 Topography, climate and ice masses: a review. *Terra Nova*, **5**: 332-342.

Kershaw, K.A. 1972 The relationship between moisture content and net assimilation rate of lichen thalli and its ecological significance. *Canadian Journal of Botany*, **50**: 543-555.

Kershaw, K.A. 1983 The thermal operating-environment of a lichen. *Lichenologist*, **15**: 191-207.

Kershaw, K.A. 1985 *Physiological ecology of lichens.* Cambridge University Press. 293pp.

King, R.B. 1971 Boulder polygons and stripes in the Cairngorm Mountains, Scotland. *Journal of Glaciology*, **10**: 375-86.

Kirkbride, M.P. and Dugmore, A.J. 2001a Timing and significance of mid-Holocene glacier advances in northern and central Iceland. *Journal of Quaternary Science*, **16**: 145-154.

Kirkbride, M.P. and Dugmore, A.J. 2001b Can lichenometry be used to date the 'Little Ice Age' glacial maximum in Iceland? *Climatic Change*, **48**: 151-167.

Kirkpatrick, A.H. and Rushton, B.S. 1990 The oceanicity/continentality of the climate of the north of Ireland. *Weather*, **45(9)**: 322-326.

Kjartanssan, G. 1943 *Árnesingasaga.* Árnesingafélagid I Reykjavik.

Kjartanssan, G. 1958 Jarðmyndanir i Holtum og nágrenni. *Rit landbúnadardeildar, B-flokkur*, **11**. Atvinnudeild Háskólans, Reykjavik: 1-22.

Kjartansson, G. 1964 Islaldalok og eldfjöll a Kili. *Náttúrufraeðingurinn.* **34**: 9-38.

Kjartansson, G. 1966 Nokkrar nyjar ^{14}C-aldursákvardanir. *Náttúrufraeðingurinn.* **36**: 126-141.

Knudsen., K.L., Henriksen, P., Rytter, F., Jiang, H. and Eiriksson, J. 1998. Palaeoclimatic shifts on the North Atlantic Shelf since 14,000 BP: Biostratigraphical results. In, Wilson, J and Plesner, S (eds.), *23rd Nordic geological Winter Meeting* (Abstract Volume). University of Arhus.

Koç, N., Jansen, E. and Haflidason, H. 1993 Palaeoceanographic reconstrictions of surface ocean conditions in the Greenland, Iceland and Norwegian Seas through the last 14 ka based on diatoms. *Quaternary Science Reviews*, **12**: 115-140.

Koch, L. 1945 *The East Greenland Ice.* Meddelelser om Grönland.

Koerner, R.M. 1980 The problem of lichen-free zones in Arctic Canada. *Arctic and Alpine Research,* **12**: 87-94.

Koppen, W. 1923 *Die Klimat der Erde.* Berlin.

Krauss, W. 1986 The North Atlantic current. *Journal of Geophysical Research*, **91**: 5061-5074.

Kruger, J. 1995 The origin and climatological significance of push moraines, Myrdalsjökull, Iceland. *The Holocene*, **5**: 420-427.

Kugelmann, O. 1990 Datierung Neuzeitlicher Gletschervorstosse im Svarfaðardalur/ Skiðadalur (Nordisland) mit Einer Neu Erstallen Flechtenwachstumskurve. *Münchener Geographische Abhandlungen Reche B*, **8**: 36-58.

Kugelmann, O. 1991 Dating recent glacier advances in the Svarfaðardalur–Skiðadalur area of northern Iceland by means of a new lichen curve. *In*: Maizels, J.K. and Caseldine, C. (Eds.) *Environmental Change in Iceland: Past and Present*: 203-217. Kluwer Academic Publishers.

Kuhle, M. 1988 Topography as a fundamental element of glacial systems. *Geojournal*, **17**: 545-568.

Kushnir, Y. and Wallace, J.M. 1989 Low frequency variability in the Northern-Hemisphere winter – Geographical-distribution, structure and time-scale. *Journal of Atmospheric Science*, **46**: 3122-3142.

Lamb, H.H. 1977 *Climate: Past, Present and Future. Volume 2 - Climatic History and the Future.* Methuen: London.

Lamb, H.H. 1979 Climatic variation and changes in the wind and ocean circulation: the Little Ice Age in the Northeast Atlantic. *Quaternary Research*, **11**: 1-20.

Lamb, H.H., Probert-Jones, J.R. and Sheard, J.W. 1963 A new advance of the Jan Mayen glaciers and a remarkable increase of precipitation. *Journal of Glaciology*, 9: 355-365.

Lange, O.L. 1953 Hitze- und Trockenresistenz der Flechten in Beziehung zu ihrer Verbreitung. *Flora.* **140**: 39-97.

Lange, O.L. 1965 Der CO_2-Gaswechsel von Flechten bei tiefen Temperaturen. *Planta.* **64**: 1-19.

Lange, O.L. 1969 Experimentell-ökologische Untersuchungen an Flechten der Negev-Wuste. I. CO_2-Gaswechsel von *Ramalina maciformis* (Del.) Bort unter kontrollierten Bedingungen im Laboratorium. *Flora, Abt. B* **158**: 324-359.

Lange, O.L. 1990 Twenty-three years of growth measurements on the crustose lichen *Caloplaca aurantia* in the central Negev Desert. *Israel Journal of Botany.* **39**: 383-394.

Lange, O.L., Meyer, A., Zellner, H., Ullmann, I. and Wessels, D.C.J. 1990 Eight days in the life of a desert lichen: water relations and photosynthesis of *Teloschistes capensis* in the coastal fog zone of the Namib Desert. *Madoqua*, **17(1)**: 17-30.

Lange, O.L., Meyer, A., Ullmann, I. and Zellner, H. 1991 Mikroklima, Wassergehalt und Photosynthesis von Flechten in der küstennahen Nebelzone der Namib-Wüste: Messungen während der herbstlichen Witterungsperiode. *Flora*, **185(4)**: 233-266.

Larsen, G. 1984 Recent volcanic history of the Veiðivötn fissure swarm, southern Iceland - an approach to volcanic risk assessment. *Journal of Volcanology and Geothermal Research*, **22**: 33-58.

Larsen, G. and Thorarinsson, S. 1977 H4 and other acidic Hekla tephra layers. *Jökull*, 27: 28-46.

Larson, D.W. 1978 Patterns of lichen photosynthesis and respiration following prolonged frozen storage. *Canadian Journal of Botany*, **56**: 2119-2123.

Larson, D.W. 1983 Environment stress and *Umbilicaria* lichen thalli. *American Journal of Botany*, **73**: 214-223.

Larson, D.W. and Kershaw, K.A. 1975 Measurement of CO_2 exchange in lichens: a new method. *Canadian Journal of Botany*, **73**: 1535-1541.

Laundon, J.R. 1980 The use of lichens for dating walls in Bradgate Park, Leicestershire. *Transactions of the Leicester Literary and Philosophical Society*, **74**: 11-30.

Lawrey, J.D. and Hale, M.E. 1977 Natural history of Plummers Island, Maryland XXIII. Studies on lichen growth rate at Plummers Island, Maryland. *Proceedings of the Biological Society of Washington*, **90**: 698-725.

Lean, J., Beer, J. and Bradley, R. 1995 Reconstruction of solar irradiance since 1610 – implications for climate-change. *Geophysical Research Letters*, **22**: 3195-3198.

Lindsay, D.C. 1973 Estimates of lichen growth rates in the maritime Antarctic. *Arctic and Alpine Research*, **5**: 341-346.

Lindsay, D.C. 1977 Lichens of cold deserts. *In:* Seaward, M.R.D. (ed.) *Lichen Ecology*. Academic Press, London: 183-209.

Link, S.O., Driscoll, M.F. and Nash, T.H. 1985 CO_2 exchange in lichens: towards a mechanistic model. *In:* D. H. Brown (ed.) *Lichen Physiology and Cell Biology*. Plenum Press, New York and London: 77-91.

Linkola, K. 1918 *Meddn Societe Fauna Fennici*, **44**: 153.

Llano, G.A. 1959 Antarctic plant life. *IGY Bulletin (National Academy of Sciences)*, **24**: 10-13.

Locke, C.W. and Locke, W.W. 1977 Little Ice Age snow-cover extent and palaeoglaciation thresholds: north-central Baffin Island, N.W.T., Canada. *Arctic and Alpine Research*, **9**: 291-300.

Locke, W.W., Andrews, J.T. and Webber, P.J. 1979 *A manual for lichenometry*. British Geomorphological Research Group Technical Bulletin 26.

Lowe, D.J. 1988 Late Quaternary volcanism in New Zealand: towards an integrated record using distal airfall tephras in lakes and bogs. *Journal of Quaternary Science*, **3**: 111-120.

Luckman, B.H. 1977 Lichenometric dating of Holocene moraines at Mount Edith Cavell, Jasper, Alberta. *Canadian Journal of Earth Sciences*, **14**: 1809-1822.

Lynge, B. 1937 Lichens from West Greenland, collected chiefly by Th. M. Fries. *Meddelelser om Grönland*, **118**: 1-225.

MacFarlane, J.D. and Kershaw, K.A. 1980 Physiological-environmental interactions in lichens. XI. Snowcover and nitrogenase activity. *New Phytologist*, **84**: 703-710.

Mackintosh, A. 2000 *Glacier fluctuations and Climatic Change in Iceland*. Unpublished Ph.D. University of Edinburgh.

Macklin, M.G. 1986 Channel and Floodplain metamorphosis in the River Nent, Cumberland. *In*: Macklin, M.G. and Rose, J. (Eds.) *Quaternary River Landforms and sediments in the Northern Pennines, England: Field Guide*: 19-33. British Geomorphological Research Group / Quaternary Research Association, London.

Macklin, M.G., Rumsby, B.T. and Heap, T. 1993 Flood alluviation and entrenchment: Holocene valley floor development and transformation in the British Uplands. *Geological Society of America Bulletin.*

Maizels, J.K. and Dugmore, A.J. 1985 Lichenometric dating and tephrochronology of sandur deposits, Sólheimajökull area, southern Iceland. *Jökull*, **35**: 69-77.

Maizels, J.K. and Petch, J.R. 1985 Age determination of inter-moraine areas, Austerdalen, southern Norway. *Boreas*, **14**: 51-65.

Maksimov, I.V. and nine others 1967 Nutational migration of the Iceland low pressure. *Doklady* **177**: 88-91. Moskow

Manabe, S and Stuoffer, R.J. 1995 Simulation of abrupt climate-change induced by freshwater input to the North Atlantic Ocean. *Nature*, **378**: 165-167.

Manabe, S and Stuoffer, R.J. 2000 Study of abrupt climate change by a coupled ocean-atmosphere model. *Quaternary Science Reviews*, **19**: 285-299.

Manabe, S. and Stouffer, R. 1988 Two stable equilibria of a coupled ocean-atmosphere model. *Journal of Climate*. **1**: 841

Mangerud, J., Furnes, H. and Johansen, J. 1986 A 9000 year old ash bed on the Faroe Islands. *Quaternary Research*, **26**: 262-265.

Manley, G. 1959 The late-glacial climate of north-west England. *Manchester and Liverpool Geological Journal*, **2**: 188-215.

Mann, M.E., Bradley, R.S., Hughes, M.K. 1999 Northern hemisphere temperatures during the past millenium: inferences, uncertainties and limitations. *Geophysical Research Letters*, **26**: 759-762.

Martin, H.E. and Whalley, W.B. 1987 A glacier ice-covered rock glacier in Tröllaskagi, northern Iceland. *Jökull*, **37**: 49-55.

Martin, H.E., Whalley, W.B., Orr, J. and Caseldine, C. 1994 Dating and interpretation of rock glaciers using lichenometry, south Tröllaskagi, north Iceland. In: Stötter, J and Wilhelm, F. *Environmental change in Iceland*. Münchener Geographische Abhandlungen **B12**: 205-224.

Martyn, D. 1991 *Climates of the World*. Elsevier, London: 435pp.

Matthews, J.A. 1974 Families of lichenometric dating curves from the Storbreen gletschervorfeld, Jotunheimen, Norway. *Norsk geografisk Tidsskrift*, **28**: 215-235.

Matthews, J.A. 1975 Experiments on the reproducibility and reliability of lichenometric dates, Storbreen gletschervorfeld, Jotunheimen, Norway. *Norsk Geografisk Tidsskrift*, **29**: 97-109.

Matthews, J.A. 1977 A lichenometric test of the 1750 end-moraine hypothesis: Storbreen gletschervorfeld, southern Norway. *Norsk geografisk Tidsskrift*, **31**: 129-136.

Matthews, J.A. 1987 A comment on "Neoglaciation in south Norway using lichenometric methods". *Norsk Geografisk Tidsskrift*, **41**: 61-67.

Matthews, J.A. 1994 Lichenometric dating: A review with particular reference to 'Little Ice Age' moraines in southern Norway. *In*: Beck, C (ed.) *Dating in Surface Contexts*. New Mexico University Press: 185-212.

Matthews, J.A., Cornish, R. and Shakesby, R.A. 1979 "Saw-tooth" moraines in front of Bødalsbreen, southern Norway. *Journal of Glaciology*, **22** (88): 535-546.

Matthews, J.A., Dawson, A.G. and Shakesby, R.A. 1986 Lake shoreline development, frost weathering and rock platform erosion in an alpine periglacial environment, Jotunheimen, Norway. *Boreas*, **13**: 333-400.

Matthews, J.A., Dahl, S.O., Nesje, A., Berrisford, M. and Andersson, C. 2000 Holocene glacier variations in central Jotunheim, southern Norway based on distal glaciolacustrine sediment cores. *Quaternary Science Reviews*, **19**: 1625-1647.

McBoyle, G.R. and Steiner, D. 1972 A factor-analysis approach to the problem of continentality. *Geografiska Annaler*, **54**: 12-27

McCarthy, D.P. 1997 Habitat selection and ecology of *Xanthoria elegans* (Link) Th. Fr. in glacier forefields: implications for lichenometry. *Journal of Biogeography*, **24**: 363-373.

McCarthy, D.P. 1999 A biological basis for lichenometry? *Journal of Biogeography*, **26**: 379-386.

Meese, D.A., Gow, A.J., Grootes, P., Mayewski, P.A., Ram, M., Stuiver, M., Taylor, K.C., Waddington, E.D. and Zielinski, G.A. 1994 The accumulation record from the GISP 2 core as an indicator of climate-change throughout the Holocene. *Science*. **266**: 1680-1682.

Meier, M.F. and Post, A.S. 1962 Recent variations in mass net budgets of glaciers in western North America. *IASH* **58**: 63-77.

Mercer, J.H. 1961 The response of fjord glaciers to changes in the firn limit. *Journal of Glaciology*, **10**: 850-858.

Miller, C.D. 1969 Chronology of neoglacial moraines in the Dome Peak area, North Cascade Range, Washington. *Arctic and Alpine Research*, **1(1)**: 49-66.

Miller, C. D. 1973 Chronology of Neoglacial deposits in the Northern Sawatch Range, Colorado. *Arctic and Alpine Research*, **5**: 385-400.

Miller, G.H. and Andrews, J.T. 1972 Quaternary history of northern Cumberland Peninsula, East Baffin Island, Northwest Territory, Canada. Part VI: Preliminary lichen growth curve. *Geological Society of America Bulletin*, **83**: 1133-1138.

Milne, J.A. 1982 River channel change in the Harthorpe Valley, Northumberland, since 1897. *University of Newcastle-Upon-Tyne, Department of Geography, Research Series*, **13**: 1-39.

Mottershead, D.N. and White, I.D. 1972 The lichenometric dating of glacier recession, Tunbergsdal, southern Norway. *Geografiska Annaler*, **54A**: 47-52.

Murray, D.R. and Locke, W.W., 1989 Dynamics of the late Pleistocene Big Timber Glacier, Crazy Mountains, Montana, USA. *Journal of Glaciology*, **35**: 183-190.

Mysak, L.A. and Power, S.B. 1996 Sea-ice anomalies in the western Arctic and Greenland-Iceland Sea and their relation to an interdecadal climate cycle. *Climatological Bulletin*, **26**: 147-176.

Mysak, L., Ingram, R., Wang, J. and van der Baaren, A. 1996 The anomalous sea-ice extent in Hudson Bay, Baffin Bay and the Labrador Sea during three simultaneous NAO and ENSO episodes. *Atmosphere Ocean*, **34**: 313-343.

Nash, T.H. (ed.) 1996 *Lichen Biology*. Cambridge University Press. 303 pp

Nash, T.H. and Gries, C. 1991 Lichens as indicators of air pollution. *In:* Hutzinger, O. (Ed.) *The Handbook of Environmental Chemistry*. Springer-Verlag, Heidelberg, New York: 1-29.

Nash, T.H., Kappen, L., Losch, R. Matthes-Sears, U. Larson, D.W. 1987 Cold resistance of lichens. *In:* E. Peveling (Ed.) *Progress and Problems in Lichenology in the Eighties*. Bibliotheca Lichenologica, No. 25, J.Cramer, Berlin-Stuttgart: 313-317.

Nesje, A., Kvamme, M. and Rye, N. 1989 Neoglacial gelifluction in the Jostedalsbreen region, western Norway – evidence from dated buried paleopodsols. *Earth Surface Processes*, **14**: 259-270.

Nesje, A., Johannessen, T. and Birks, H.J.B. 1995 Briksdalsbreen, western Norway: climatic effects on the terminal response of a temperate glacier between AD 1901 and 1994. *The Holocene*, **5**: 343-347.

Nesje, A., Lie, Ø. and Dahl, S.O. 2000 Is the North Atlantic Oscillation reflected in Scandinavian glacier mass balance records? *Journal of Quaternary Science*, **15(6)**: 587-601.

Newton, A. 1999 *Ocean –transported pumice in the North Atlantic.* Unpublished Ph.D. thesis, Geography Department, University of Edinburgh.

Nienburg, W. 1919 Studien zur Biologie der Flechten, I. II. III. *Zetischrift Botanica*, **11**: 1-38.

Nikonov, A.A. and Shebalina, T.Y. 1979 Lichenometry and earthquake age determination in Central Asia. *Nature*, **280**: 675-677.

Norðdahl, H. 1979 *The last glaciation in Flateyardalur central north Iceland, a preliminary report.* University of Lund, Department of Quaternary Geology, Report 18, Lund.

Norðdahl , H. 1981 A prediction of the minimum age of the Weichselian maximum glaciation in North Iceland. *Boreas*, 10: 471-476.

Norðdahl, H. 1983 Late Quaternary Stratigraphy of Fnjóskadalur Central North Iceland: A Study of Sediments, Ice-Lake Strandlines, Glacial Isostasy and Ice-Free Areas. *Lundqua Thesis* **12.**

Norðdahl, H. 1991 Late Weichselian and early Holocene deglaciation history of Iceland. *Jökull*, **40**: 27-50.

Norðdahl, H. and Einarsson, T. 1988 Late Weichselian deglaciation in east and southeast Iceland *Náttúrufræðingurinn*, 58 (2) 59-80.

Nyberg, R. 1985 Debris flow and slush avalanches in Swedish Lappland, Northern Scandinavia. *Meddelanden från Lunds Universitets Geografiska Institution, Avhandlingar*, **97**: 1-222.

Nye, J. 1961 The influence of climatic variations on glaciers. *International Association of Science and Hydrology*, **54**: 397-404.

Oerlemans, J. 1989 On the response of valley glaciers to climatic change. In: Oerlemans, J. (ed.) *Glacier Fluctuation and Climatic Change*. Kluwer Academic Publishers, Dordrecht: 353-372.

Oerlemans. J. 1998 (ed) *Climate sensitivity of glacier mass balance: the effect of topographic barriers.* Second annual report.

Oerlemans, J., Björnsson, H., Kuhn, M., Obleitner, F., Palsson, F., Smeets, P., Vugts, H.F. and de Wolde, J. 1999 Glacio-meteorological investigations on Vatnajökull, Iceland, Summer 1996: An overview. *Boundary-Layer Meteorology*. **92**: 3-26.

Ogilvie, A.E.J. 1981 *Climate and Society in Iceland from the Medieval Period to the Late Eighteenth Century*. Unpublished Ph.D. dissertation, University of East Anglia, Norwich, U.K.

Ogilvie, A.E.J. 1984 The past climate and sea-ice record from Iceland, part 1: data to A.D. 1780. *Climatic Change*. **6**: 131-152.

Ogilvie, A.E.J. 1990 The climate of Iceland, 1701-1784. *Jökull*, **40**: 77-80.

Ogilvie, A.E.J. 1992 Documentary evidence for changes in the climate AD 1500-1800. *In*: Bradley, R. and Jones, P. (eds.) *Climate since AD 1500*. Routledge, London: 92-117.

Oliver, J.E. 1970 A genetic approach to climate classification. *Annals of the Association of American Geographers*, **56**: 508-515.

Orombelli, G. and Porter, S.C. 1983 Lichen growth curves for the southern flank of the Mont Blanc Massif, western Alps. *Arctic and Alpine Research*, **15**: 193-200.

Orwin, J. 1970 Lichen succession on recently deposited rock surfaces. *New Zealand Journal of Botany*, **8(4)**: 452-477.

Orwin, J. 1972 The effect of environment on assemblages of lichens growing on rock surfaces. *New Zealand Journal of Botany*, **10**: 37-47.

Osborn, G. and Luckmann, B.H. 1988 Holocene glacier fluctuations in the Canadian Cordillera (Alberta and British Columbia). *Quaternary Science Reviews*, **7**: 116-128.

Østrem, G. 1966 The height of the glaciation limit in southern British Columbia and Alberta. *Geografiska Annaler*. **48A**: 126-138.

Palais, J.M., Taylor, K., Mayweski, P.A. and Grootes, P. 1991 Volcanic ash from the 1362 AD. Öræfajökull eruption (Iceland) in the Greenland ice sheet. *Geophysical Research Letters*, **18**: 1241-1244.

Paterson, D.R. 1951 Lichens growing around a snowpatch margin. *Journal of Botany*, **45**: 200-212.

Paterson, D.R., Paterson, E.W. and Kenworthy, J.B. 1983 Physiological studies on temperate lichen species I. A mathematical model to predict assimilation in the field, based on laboratory responses. *New Phytologist*, **94**: 605-618.

Pearson, L.C. 1969 Influence of temperature and humidity on distribution of lichens in a Minnesota bog. *Ecology*, **50**: 740-746.

Pentecost, A. 1980 The lichens and bryophytes of rhyolite and pumice tuff rock outcrops in Snowdonia and some factors affecting their distribution. *Journal of Ecology*, **68**: 251-267.

Perlwitz, J. and Graf, H.F. 1995 The statistical connection between tropospheric and stratospheric circulation of the Northern Hemisphere in winter. *Journal of Climate*, **8**: 2281-2295.

Pétursson, H.G. 1986 *Kvarteraergeologiske undersökelser på Vest-Melrakkaslétta, Nordöst Island*. Cand. Real Thesis, University of Tromsö.

Pétursson, H.G. 1991 The Weichselian glacial history of West Melrakkaslétta, northeastern Iceland. *In*: Caseldine, C.J. and Maizels, J. (eds.) *Environmental Changes in Iceland: Past and Present*. Kluwer, Dordrecht: 49-65.

Péwé, T.L. and Reger, R.D. 1972 Modern and Wisconsinan snowlines in Alaska. *Proceedings of the 24th International Geologic Congress, Section 12*: 187-197.

Phillips, H.C. 1963 Growth rate of *Parmelia isidiosa* (Mull. Arg.) Hale. *Journal of the Tennessee Academy of Science*, **38(3)**: 95-96.

Phillips, H.C. 1969 Annual growth rates of three species of foliose lichens determined photographically. *Bulletin of the Torrey Botanical Club*, **96**: 202-206.

Pickett, R.L. and Athey, G.L. 1968 Iceland sea-surface temperature survey, April 1968. Naval Oceanographic Office: IR 68-46. Washington D.C.

Pitman, G.T.K. 1973 A lichenometrical study of snow patch variation in the Frederikshåb District, South-West Greenland, and its implications for studies of climatic and glacial fluctuations. *Meddeleser om Grønland*, **192**: 1-31.

Pjeturss 1910. Island: Handbuck der Regionalen. *Geologie*, 4: 1-22.

Platt, R.B. and Amsler, F.P.A. 1955 A basic method for the immediate study of lichen growth rates and succession. *Journal of the Tennessee Academy of Science*, **30(2)**: 177-183.

Poore, M.E.D. and McVean, D.N. 1957 A new approach to Scottish mountain vegetation. *Journal of Ecology*, **45**: 401-439.

Porter, S.C. 1975 Equilibrium line altitudes of late Quaternary glaciers in the Southern Alps, New Zealand. *Quaternary Research*, **5**: 27-47.

Porter, S.C. 1977 Present and past glaciation threshold in the Cascade Range, Washington, USA: Topographic and climatic controls and palaeoclimatic implications. *Journal of Glaciology*, **18**: 101-116.

Porter, S.C. 1981 Lichenometric studies in the Cascade Range of Washington: Establishment of *Rhizocarpon geographicum* growth curves at Mount Rainier. *Arctic and Alpine Research*, **13**: 11-23.

Porter, S.C. and Orombelli, G. 1981 Alpine rockfall hazards. *American Scientist*, **69**: 67-75.

Price, R.J. 1970 Moraines at Fjallsjökull, Iceland. *Arctic and Alpine Research*, **2(1)**: 27-42.

Proctor, M.C.F. 1977 The growth curve of the crustose lichen *Buellia canescens* (Dicks.) De Not. *New Phytologist*, **79**: 659-663.

Proctor, M.C.F. 1983 Sizes and growth-rates of thalli of the lichen *Rhizocarpon geographicum* on the moraines of the Glacier de Valsorey, Valais, Switzerland. *The Lichenologist*, **15**: 249-261.

Rapp, A. and Nyberg, R. 1981 Alpine debris flows in Northern Scandinavia. *Geografiska Annaler*, **63(A)**: 183-196.

Redon, J. and Lange, O.L. 1983 Epiphytische Flechten im Bereich einer chilenischen "*Nebeloase*" (Fray Jorge). I. Vegetationskündliche Gliederung und Standortsbedingungen. *Flora*, **174**: 213-243.

Reger, R.D. and Péwé, T.L. 1969 Lichenometric dating in the Central Alaska Range. *In:* Péwé, T.L. *The Periglacial Environment*: 223-247. McGill-Queen's University Press, Montreal.

Ried, A. 1960 Thallusbau und Assimilationshaushalt von Laub- und Krunstenflechten. *Biol. Zentralbl.*, **79(2)**: 129-151.

Roberts, S.J. 2001. *Quaternary tephrochronology in Iceland: dating principles and environmental applications*. Unpublished PhD thesis, Department of Geography, University of Edinburgh.

Rodbell, D.T. 1990 Soil-age relationships on Late Quaternary moraines, Arrowsmith Range, Southern Alps, New Zealand. *Arctic and Alpine Research*, **22**: 355-365.

Roemmich, D. and Wunsch, C. 1985 Two transatlantic sections: Meridional circulation and heat flux in the subtropical North Atlantic Ocean. *Deep Sea Research*, **32**: 619.

Rogers, J.C. 1984 The association between the NAO and the SO in the Northern Hemisphere. *Monthly Weather Review*, **112**: 1999-2015.

Rogerson, R.J., Evans, D.J.A. and McCoy, W.D. 1986 Five-year growth of rock lichens in a low-arctic mountain environment, Northern Labrador. *Géographie physique et Quaternaire*, **XL(1)**: 85-91.

Rooth, C. 1982 Hydrology and ocean circulation. *Progress in Oceanography*, **11**: 131-140.

Rosqvist, G. and Ostrem, G. 1989 The sensitivity of a small ice-cap to climatic fluctuations. *Geografiska Annaler*, **71A**: 99-103.

Rossby, C.-G. 1949 On the nature of the general circulation of the lower atmosphere. In: Kuiper, G.P. (ed) *The Atmosphere of the earth and planets*. University of Chicago Press: 16-48.

Rossby, T. 1996 The North Atlantic current and surrounding waters: at the crossroads. *Review of Geophysics*, **34(4)**: 463-81.

Röthlisberger, F. 1986 *10,000 Jahre Gletschergeschichte der Erde*. Sauerländer, Aarau.

Ruddiman, W.F. and McIntyre, A. 1979 Warmth of the subpolar North Atlantic Ocean during Northern Hemisphere ice-sheet growth. *Science*, **204**: 173-175.

Rundgren, M. and Ingolfsson, Ó. 1999 Plant survival in Iceland during periods of glaciation? *Journal of Biogeography*, **26**: 387-396.

Runemark, H. 1956 Studies in *Rhizocarpon* II. Distribution and ecology of the yellow species in Europe. *Opera Botanica*, **2**: 1-150.

Russell, A., Knudsen, Ó., Maizels, J., Marren, P.M. 1999 Channel cross-section area changes and peak discharge calculation on the Gigjukvisl during the November 1996 jökulhlaup, Skeiðarársandur, Iceland. *Jökull*, **47**: 45-58.

Rydzak, J. 1956 A method of studying growth in lichens. *Annales Universitatis Mariae Curie-Sklodowska, Section. C [Lublin]*, **10(4)**: 87-91.

Rydzak, J. 1958 The influence of small towns on the lichen vegetation, Part VII. Discussion and general conclusions. *Annales Universitatis Mariae Curie-Sklodowska, Section. C [Lublin]*, **13(6)**: 275-323.

Rydzak, J. 1961 Investigations on the growth rate of lichens. *Annales Universitatis Mariae Curie-Sklodowska, Section. C [Lublin]*,**16(1)**: 1-15.

Rydzak, J. 1966 Badania nad szybkoscia wzrostu porostow. Czesc II. Investigations on the growth rate lichens. Part II. *Annales Universitatis Mariae Curie-Sklodowska, Section. C [Lublin]*,**21**: 167-182.

Sachs, J. 1882 *Textbook of botany, morphological and physical*. Oxford University Press.

Schell, I.I. 1961 The ice off Iceland and the climates during the last 1200 years approximately. *Geografiska Annaler*, **43**: 354-362

Schmitz, W.J. and McCartney, M.S. 1993 On the North-Atlantic circulation. *Reviews of Geophysics*, **31**: 29-49.

Schneider, A. 1897 *A text-book of general lichenology*. Clute, Binghampton, New York.

Seaward, M.R.D. 1977 *Lichen Ecology*. Academic Press, London.

Self, S. and Sparks, R.S.J. 1981 (eds) *Tephra Studies*. 1-12, Reidel, Dordrecht.

Shakesby, R.A., Dawson, A.G., Matthews, J.A. 1987 Rock glaciers, protalus ramparts and related phenomena, Rondane, Norway: A continuum of large-scale talus-derived landforms. *Boreas*, **16**: 305-317.

Sharp, M. 1984 Annual moraine ridges at Skálafellsjökull, south-east Iceland. *Journal of Glaciology*, **30(104)**: 82-93.

Sharp, M. and Dugmore, A.J. 1985 Holocene glacier fluctuations in eastern Iceland. *Zeitschrift für Gletscherkunde und Glazialgeologie*, **21**: 341-349.

Sigbjarnarson, G. 1982 Alpajöklar og öldubrjótar. In: Thórarinsdóttir, H., Óskarsson, O.H., Steinthórsson, S. and Einarsson, Th. (eds) *Eldur er I nordri*. Sögufélag, Reykjavik: 79-89.

Sigtryggson, H. 1972 An outline of sea ice conditions in the vicinity of Iceland. *Jökull*, **22**: 1-11.

Sigurdsson, H. 1982 Útbreidsla islanskra gjóskulaga á botni Atlantshaf. In Thorarinsdaottir, H., Oskarsson., O.H., Steinthorsson, S., and Einarsson, Th. (eds.) *Eldur er i nordri*. Reykjavik: 119-127.

Sigurdsson, O. 1998 Glacier variations in Iceland 1930-1995: From the database of the Iceland Glaciological Society. *Jökull*, **45**: 3-25.

Sigurvinsson, J.R. 1983 Weichselian glacial lake deposit in the highlands of north-western Iceland. *Jökull*, **33**: 99-109.

Sigurvinsson, J.R. 1998 *Unpublished data*. Isafjörður, Iceland.

Smirnova, T.Tu. and Nikonov, A.A. 1990 A revised lichenometric method and its application: dating great past earthquakes. *Arctic and Alpine Research*, **22**:375-388.

Smith, D.C. 1962 The biology of lichen thalli. *Biological Review* **37**: 537-570

Smith, D.J. 1988 Rates and controls of soil movements on a solifluction slope in the Mount Rae area, Canadian Rocky Mountains. *Zeitschrift für Geomorphologie*, supp. **71**: 25-44.

Smith, D.J., McCarthy, D.P. and Colenutt, M.E. 1995 Little Ice Age glacial activity in Peter Lougheed and Elk Lakes Provincial Parks, Canadian Rocky Mountains. *Canadian Journal of Earth Sciences*, **32**: 579-589.

Spence, J.R. and Mahaney, W.C. 1988 Growth and ecology of *Rhizocarpon* section *Rhizocarpon* on Mount Kenya, East Africa. *Arctic and Alpine Research*, 20: 237-242.

Steffensen, E. 1982 The climate of Norwegian-Arctic stations. *Klima-Norwegian Meteorologiske Instsitut*. (5): 44pp

Stötter, J. 1990 Beobachtungen und Üntersuchungen zur postglazialen Landschaftsgeschichte Islands am Beispiel des Svarfaðardalur-Skiðadalsdalur. *Münchener Geographische Abhandlungen.*, **B8**: 83-104. München.

Stötter, J. 1991 New observations on the postglacial glacier history of Tröllaskagi, northern Iceland. *In*: Maizels, J.K. and Caseldine, C. (Eds.) *Environmental Change in Iceland: Past and Present*. Kluwer, Dordrecht: 181-192.

Stötter, J. 1994 Changing the Holocene record – A call for international interdisciplinary co-operation. *Münchener Geographische Abhandlungen*, **B12**: 257-273.

Stötter, J., Wastl, M., Caseldine, C. and Häberle, T. 1999 Holocene palaeoclimatic reconstruction in northern Iceland: approaches and results. *Quaternary Science Reviews*, **18**: 457-474.

Sturm, M., Hall, D., Benson, C. and Field, O. 1991 Non-climatic control of glacier fluctuations in the Wrangell and Chugach Mountains, Alaska, USA. *Journal of Glaciology*, **37**: 348-356.

Sutherland, D.G. 1984 Modern glacier characteristics as a basis for inferring former climates with particular reference to the Loch Lomond Stadial. *Quaternary Science Reviews*, **3**: 291-309.

Sutton, R.T. and Allen, M.R. 1997 Decadal predictability of North Atlantic sea surface temperatures and climate. *Nature*, **388**: 563-567.

Ten Brink, N.W. 1973 Lichen growth rates in West Greenland. *Arctic and Alpine Research*, **5**: 323-331.

Thompson, A. 1988 Historical development of the proglacial landforms of Svinafellsjökull and Skaftafellsjökull, southeast Iceland. *Jökull*, **38**: 17-31.

Thompson, A. and Jones, A. 1986 Rates and causes of proglacial river terrace formation in southwest Iceland: An application of lichenometric dating techniques. *Boreas*, **15**: 231-246.

Thorarinsson, S. 1936 Vatnajökull. Chapter III. *Geografiska Annaler*, **18**: 189-195.

Thorarinsson, S. 1937 The main geological and topographical features of Iceland. *Geografiska Annaler*, **19**: 161-174.

Thorarinsson, S. 1943 Vatnajökull: Scientific results of the Swedish-Icelandic investigations 1936-37-38. Chapter XI: Oscillations of the Iceland glaciers in the last 250 years. *Geografiska Annaler*, **1-2**: 1-54.

Thorarinsson, S. 1944 Tefrokronologiska studier på Island. *Geografiska Annaler*, **26A**: 1-217.

Thorarinsson, S. 1946 *Lesbók Morgunblaðsins*. Reykjavik: 21pp.

Thorarinsson, S. 1949 Some tephrachronological contributions to the volcanology and glaciology of Iceland. *Geografiska Annaler*, **21**: 239-256.

Thorarinsson, S. 1951 Laxárgljúfur and Laxárhraun, a tephrochronological study. *Geografiska Annaler*, **33**: 1-90.

Thorarinsson, S. 1956 On the variations of Svinafellsjökull, Skaftafellsjökull and Kviárjökull in Öræfi. *Jökull*, **6**: 1-15.

Thorarinsson, S. 1958 The Öræfajökull eruption of 1362. *Acta Nat. Isl.* II(2). Reykjavik, Náttúrufræðistofnun Islands.

Thorarinsson, S. 1960 Glaciological knowledge in Iceland before 1800: A historical outline. *Jökull*, **10**: 1-18.

Thorarinsson, S. 1967 Fortvitnilegir jokulgarðar við jaðar Skeiðarájokuls. *Jökull*. **17**, 311-312.

Thorarinsson, S. 1974 Vötnin strið. Saga Skeiðarárhlaupa og Grimsvatnagosa, Reykavik, Bókaútgáfa Menningarsjóðs.

Thorarinsson, S. 1975 Katla og annáll Kötlgosa. *Árbók Ferðafélags Islands*. Ferðafélag Islands: 125-149.

Thorarinsson, S. 1981 The application of tephrochronology in Iceland. In: Self, S. and Sparks, R.S.J. (eds) *Tephra Studies*. 1-12, Reidel, Dordrecht.

Thornthwaite, C.W. 1933 The climates of the earth. *Geographical Review*, **23**: 433-440.

Thoroddsen, Th. 1905-1906 Island. Gundrisz der Geographie und Geologie. Ergänzungsheft 152, 153 zu Peterm. Mitteilungen. Gotha.

Thoroddsen, Th. 1913-1915 *Ferdabók, I-IV*. Kaupmannahöfn.

Thorodssen, T. 1916-1917 *Arferdi a Islandi I Thusund* ar 1-2.

Timmis, R.J. 1986 *Glacier changes in South Georgia and their relationship to climatic trends.* Unpublished PhD thesis, University of East Anglia.

Todtmann, E.M. 1952 Im Gletscherruckzugsgebiet des Vatnajökull auf Island, 1951. *Neues Jahrbuch Geologie und Palaeontologie*, 1952: 401-411.

Topham, P.B. 1977 Colonization, growth and lichen taxonomy. *In*: Seaward, M.R.D. (ed.) *Lichen Ecology*. Academic Press, London.

Torsnes, I., Rye, N. and Nesje, A. 1993 Modern and Little Ice-Age equilibrium line altitudes on outlet valley glaciers from Jostedalsbreen, western Norway: an evalutation of different approaches to their calculation. *Arctic and Alpine Research*, **25**: 106-116.

Tournefort, J.P de. 1694 *Elemens de Botanique*. Paris.

Trewartha, G.T. 1954 *An introduction to Climate*. 3rd Edition. McGraw-Hill, New York. 402pp

Trinci, A.P.J. 1971 Influence of the width of the peripheral growth zone on the radial growth rate of fungal colonies on solid media. *Journal of Genetic Microbiology*, **67**: 325-344.

Vallot, J. 1896 Sur la vitesse de la croissance d'un lichen saxicole. *Review of General Botany*, **8**: 201-202.

Van Geel, B., Heusser, C.J., Renssen, H. ET AL 2000 Climatic change in Chile at around 2700 BP and global evidence for solar forcing: A hypothesis. *Holocene*. **10**: 659-664.

Van Loon, H. and Rogers, J.C. 1978 The see-saw in winter temperature between Greenland and Northern Europe. *Monthly Weather Review*, **106**: 296-310

Van Steijn, H. 1988 Debris flow involved in the development of Pleistocene stratified slope deposits. *Zeitschrift für Geomorphologie*, **71**: 45-58.

Vere, D. and Matthews, J.A. 1985 Rock glacier formation from a lateral moraine at Bukkeholsbreen, Jotunheimen, Norway: A sedimentological approach. *Zeitschrift für geomorphologie N.F.* **29**: 397-415.

Veum, T., Jansen, E., Arnold, M., Beyer, I. and Duplessy, J.C. 1992 Water mass exchange between the North Atlantic and Norwegian Sea during the past 28,000 years. *Nature*, **365**: 783-785.

Vidal, L., Labeyrie, L., Cortijo, E. ET AL 1997 Evidence for changes in North Atlantic Deep Water linked to meltwater surges during Heinrich events. *Earth Science Planetary Letters*, **146**: *13-27.*

Vilmundarson, 1972 Evaluation of historical sources on sea ice near Iceland. In: Einarsson, M.A. (ed) *Sea Ice*, Proceedings of an international conference: 159-169. Reykjavik: National Research council.

Von Boldva, A.B. 1886 Aufnahme und Beschreibung der Insel Jan Mayen. Beobachtungen über Gletscherbewegung. *Die Oesterreichische Polarstation Jan Mayen: Beobachtungs-Ergebnisse*, Wien. **1**: 27-50.

Wallace, J. 2000 North Atlantic Oscillation/Annular Mode: Two paradigms, one phenomenon. *Quarterly Journal of the Royal Meteorological Society.*

Washburn, A.L. 1956 Classification of patterned ground and review of suggested origins. *Geological Society of America Bulletin*, **67**: 823-865.

Washburn, A.L. 1969 Weathering, frost action and patterned ground in the Mesters Vig district, northeast Greenland. *Meddelesler om Grønland*, 176, 303pp.

Washburn, A.L. 1980 Permafrost features as evidence of climatic change. *Earth Science Reviews*, **15**: 327-402.

Webber, P.J. and Andrews, J.T. 1973 Lichenometry: a commentary. *Arctic and Alpine Research*, 5 (4), 295-302.

Werner, A. 1990 Lichen growth rates for the northwest coast of Spitsbergen, Svalbard. *Arctic and Alpine Research*. **22**: 129-140.

Westgate , J.A., Walter, W.C., Pearce, G.W. and Gorton, M.P. 1985 Distribution, stratigraphy, petrochemistry and palaeomagnetism of the Late-Pleistocene Old Crow tephra in Alaska and the Yukon. *Canadian Journal of Earth Sciences,* **45**: 232-249.

Whaley, W.G. 1961 *Growth as a general process.* In: Ruhland, W. (ed.) Encyclopedia of Plant Physiology. Vol. 14: pp71-112. Springer, Berlin.

Whalley, W.B. 1974 Rock glaciers and their formation as part of a glacier debris-transport system. *Geographical Papers*. University of Reading, Department of Geography, 24pp.

Wiesner, J. 1895 Untersuchungen über Lichtgenuss der Pflanzen, etc. Sitzungsb.K.Akab. Wiss. Wien., *Math-Naturw*. **104**: 606-711.

Williams, P.J. 1959 An investigation into processes occurring in solifluction. *American Journal of Science*, **257**: 42-58.

Winchester, V. 1984 A proposal for a new approach to lichenometry. *British Geomorphological Research Group, Technical Bulletin*. **33**:3-20.

Winchester, V. 1988 An assessment of lichenometry as a method for dating recent stone movements in two stone circles in Cumbria and Oxfordshire. *Botanical Journal of the Linnean Society*. **96**: 57-68.

Winchester and Harrison S. 1994 A development of the lichenometric method applied to dating of glacially influenced debris flows in southern Chile. *Earth Surface Processes and Landforms*, **19**: 137-151.

Winkler, S. 2000 The 'Little Ice Age' maximum in the Southern Alps, New Zealand: preliminary results at Mueller Glacier. *The Holocene*. **10(5)**: 643-647.

Winton, M. 1997 The effect of cold climate upon North Atlantic Deep Water formation in a simple ocean-atmosphere model. *Journal of Climate*, **10**: 37-51.

Wirth, V. 1972 Die Silikatflechten-Gemeinschaften im ausseralpinen Zentraleuropa. *Dissertation Botanische Lehre*, **17**: 1-306.

Wise, K.A.J. and Gressitt, I.L. 1965 Far southern animals and plants. *Nature*, **207**: 101.

Worthington, L.V. 1970 The Norwegian Sea as a Mediterranean basin. *Deep Sea Research*, **17**: 77-84.

Zielinski, G., *and nine others*. 1994 Record of explosive volcanism since 7000 B.C. from the GISP2 Greenland ice-core and implications for the volcano-climatic system. *Science*, **264**: 948-952.

Appendix I

Major element geochemistry of a silicic tephra layer, thought to be from the Hekla-5 eruption. Single grain analyses of selected glass shards from a fine-grained, grey-white tephra layer at 1.60 m in soil pit DH03, Dalsheiði, southeast Iceland, are presented below. The samples were analysed with a Cambridge Instruments Microscan V using a standard WDS technique, an accelerating voltage of 20kV and a beam current of 15 nA.

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
74.66	0.14	12.44	1.55	0.13	0.10	1.33	4.58	2.75	97.68
74.72	0.15	12.55	1.65	0.12	0.02	1.19	4.60	2.86	97.86
73.95	0.12	12.26	1.56	0.13	0.05	1.16	4.56	3.07	96.86
74.12	0.11	12.12	1.56	0.15	0.08	1.24	4.55	2.74	96.67
74.88	0.13	12.35	1.54	0.10	0.10	1.34	4.56	2.88	97.88
75.10	0.11	12.46	1.67	0.12	0.02	1.25	4.08	2.78	97.59
75.02	0.13	12.77	1.48	0.14	0.02	1.36	4.52	3.03	98.47
74.56	0.17	12.75	1.52	0.17	0.06	1.20	3.98	3.51	97.62

The stratigraphic relationship and micromorphology of the thin grey-white tephra layer suggest that it is H-5. The major element analysis (above) would tend to confirm this assertion. When mean values are compared, the K_2O and CaO contents are in close agreement with those given in Sigurdsson (1982). However, there is still some uncertainty surrounding the status of the Hekla-5 eruption (Newton, 2000 *pers. comm.)*, with relatively few published geochemical studies in existence.

Data for Hekla-5 tephra from Sigurdsson (1982: Table 2)

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Total
74.80	0.12	12.62	1.37	0.12	0.00	1.29	4.54	2.53	97.39

Appendix II

The choice of a suitable sample size for lichenometric studies

Method

A similar, smaller, study to that carried out by Innes (1984) was conducted as part of lichenometric research in Iceland during 1998-99. Three moraine ridges were chosen on the basis of their age, size and degree of lichen covering.

1. Moraine S9, Lambatungnajökull (dating from c. AD 1920-30)
2. Moraine HR2, Hólárjökull (dating from AD 1890-91)
3. Moraine H5, Lambatungnajökull (dating from before 1890)

The longest axis of the largest *Rhizocarpon* Section *Rhizocarpon* lichen was measured on the ice-proximal flank of each moraine within quadrats of increasing area. Quadrats ranged from 10 m^2 to 200 m^2 in area. Measurements were made with a clear, flexible ruler and were recorded to the nearest millimetre. Irregular-shaped, elongate and partially fragmented thalli were included.

By way of comparison, the entire surface of each ridge (proximal flank only) was examined and the largest Section *Rhizocarpon* lichen was also permanently recorded. The surface area of each moraine was calculated using measuring tape or approximated from aerial photographs.

Results

The results of the lichenometric surveys are presented below:

moraine	date of surface exposure	quadrat area (m^2)	largest lichen (mm)	size factor (%) lichen[1]	area[2]
Lambatungnajökull *S9*	1920-30	10	18	43.9	0.8
		30	30	73.2	2.4
		60	34	82.9	4.4
		100	35	85.4	8.0
		200	36	87.8	16.0
		1250	41	100	100
Hólárjökull *HR2*	1890-91	10	21	31.3	2.1
		30	52	77.6	6.3
		100	60	89.6	20.9
		480	67	100	100
Lambatungnajökull *H16*	pre-1890	10	42	40.0	2.9
		30	80	78.1	8.6
		60	80	78.1	17.1
		200	92	87.6	57.1
		350	105	100	100

Notes

1 - the size of the largest lichen in the quadrat expressed as a percentage of the size of the largest lichen on the entire surface

2 - the area of the quadrat expressed as a percentage of the area of the entire surface

Discussion

It can be seen clearly that the size of the 'largest-lichen' on a surface is ultimately a function of the area searched. As the search area increases, the size of the 'largest lichen' increases. This investigation confirms the findings of Innes (1984), and indicates that the largest lichen growing on a surface of limited area is likely to be significantly smaller than the largest lichen on a large surface of the same age, if all other factors are considered to be equal. These results suggest that although substrate age is important in determining the size of the largest lichen growing on a surface, sample area may be an equally important factor as stressed by Innes (1984).

On three moraine ridges of different age and size in southeast Iceland, the largest lichen in a 30 m^2 quadrat is 22-27% smaller than the largest lichen on the entire surface. Whereas the area involved in the 30 m^2-quadrat search is less than 10% of the entire surface area. The difference between the *actual* largest lichen growing on a surface and the largest lichen measured within a smaller quadrat, decreases with increasing quadrat size. Innes (1984) has suggested that this decrease is logarithmic, possibly due to the nature of lichen population distributions.

In order to promote consistency amongst lichenometric studies and minimise the 'area effect', surveys will be conducted within a fixed area. This investigation has found that a fixed area of 30 m^2 is a suitable, yet efficient, quadrat size for lichenometric studies in southeast Iceland. However, it is recognised that results obtained from fixed-area studies are not directly comparable with surveys that use the entire substrate.

Appendix III

Table AIII-1 Measurements of crustose lichens in southeast Iceland in 1996 and 1999.

Thallus	Lichen family	Maximum diameter (mm)	
		1996	1999
VW-RM-01	section *Rhizocarpon*	21	23.2
VW-RM-02	section *Rhizocarpon*	6	6.3
VW-RM-03	section *Rhizocarpon*	12	12.6
VW-RM-04	section *Rhizocarpon*	16	17.4
VW-RM-05	section *Rhizocarpon*	15	17.1
VW-RM-06	section *Rhizocarpon*	4	5.5
VW-RM-07	section *Rhizocarpon*	10	na
VW-RM-08	section *Rhizocarpon*	21	21.6
VW-RM-09	section *Rhizocarpon*	13	13.0
VW-RM-10	section *Rhizocarpon*	22	22.2
VW-RM-11	section *Rhizocarpon*	14	15.0
VW-RM-12	section *Rhizocarpon*	17	18.0
VW-RM-13	section *Rhizocarpon*	10	12.8
VW-RM-14	section *Rhizocarpon*	14	13.7
VW-RM-15	section *Rhizocarpon*	4	na
VW-RM-16	section *Rhizocarpon*	14	15.3
VW-RM-17	section *Rhizocarpon*	17	16.6
VW-RM-18	Buellia canescens	37	na
VW-RM-19	Buellia canescens	46	46.5
VW-RM-20	Buellia canescens	22	24.0
VW-RM-21	Buellia canescens	37	39.0
VW-RM-22	Buellia canescens	30	34.5

Table AIII-2 Measurements of *Rhizocarpon* Section *Rhizocarpon* thalli made in the summer of 1999. All are growing on porphyritic basaltic boulders on glacier forelands in southeast Iceland. The code given to each thallus reflects the location and the date on which the growth observations commenced. It is hoped that these thalli will be revisited in several years' time in order to determine the radial growth rate of this lichen family directly. Other observations regarding the role of age on thallus growth rates could also be investigated. Photographs and detailed descriptions of each thallus can be obtained from the author on request.

Thallus	Maximum diameter (mm)	Distance from fixed mark (mm)			
		a	b	c	d
V01-200799	41.0	6.0	6.0	7.0	6.2
V02-200799	31.0	7.0	7.0	4.5	
V03-200799	45.0	4.0	6.0	6.5	
V04-200799	16.0	5.0	5.0		
V05-200799	60.5	4.0	6.0	6.5	4.5
V06-200799	21.0	4.0	4.5	4.2	5.2
V07-200799	77.3	4.5	4.0	5.0	
V08-210799	47.0	4.8	4.0	4.0	
V09-210799	25.0	5.0	5.5		
V10-210799	25.8	5.0	5.0	6.5	
V11-220799	6.0	7.0	7.5		
V12-220799	2.8	6.0	5.7	5.0	
V13-220799	12.5	3.5	4.0	4.0	
V14-220799	18.7	4.0	5.5	6.0	

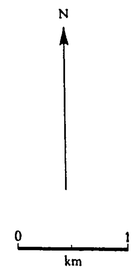
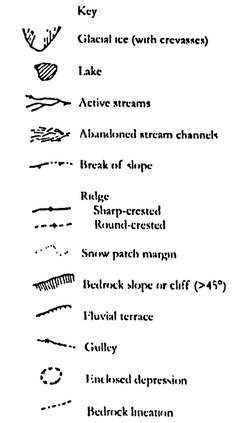
Table AIII-2 (cont.)

Thallus	Maximum diameter (mm)	Distance from fixed mark (mm)			
		a	b	c	d
SK01-300799	14.0	4.0	3.6		
SK02-300799	11.0	3.0	3.0	4.5	
SK03-300799	16.5	5.5	6.0		
SK04-300799	20.4	5.0	4.5	6.0	3.5
SK05-300799	16.0	5.0	5.5	4.5	
SK06-300799	15.0	5.5	4.5		
SK07-300799	17.0	6.0	6.0	4.0	3.0
SK08-300799	16.0	4.5	5.5	3.0	
SK09-310799	15.0	6.0	3.5	3.5	
SK10-310799	19.5	3.5	4.0	5.0	
SK11-310799	17.5	6.0	3.5	4.3	
SK12-310799	13.5	4.5	3.3		
SK13-310799	9.5	5.0	3.5		
SK14-310799	8.0	4.5	4.4	2.5	
SK15-310799	5.5	6.0	6.0		
SK16-310799	4.0	5.0	5.5		
H01-010899	95.0	5.0	5.0	4.5	3.8
H02-010899	17.0	4.0	5.0		
H03-010899	36.0	4.7	4.5		
H04-010899	77.3	4.5	4.0	5.0	
H05-010899	47.0	4.8	4.0	4.0	4.0
H06-020899	25.0	5.0	5.5		
H07-210899	25.8	5.0	5.0	6.5	
H08-220899	6.0	7.0	7.5		
H09-220899	2.8	6.0	5.7	5.0	
H10-220899	12.5	3.5	4.0	4.0	
H11-220899	18.7	4.0	5.5	6.0	

Key to localities:

V – Virkisjökull
SK – Skyndidalur
H – Hoffellsdalur

Figure 6.2

Geomorphological map of Skyndidalur, Hoffellsdalur and the area adjacent to Lambatungnajökull.

Compiled from stereographic aerial photographs taken in 1989 by *Landmaelingar Islands.* Cartography by Tom Bradwell, March 1998.