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Modelling the southern extent of the last Icelandic ice-sheet

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ABSTRACT: A three-dimensional, time-dependent numerical ice-sheet model is used to investigate the southern extent of the last Icelandic ice-sheet since the Last Glacial Maximum (LGM, ca. 20 000 ^{14}C yr BP). Ice development over southern Iceland is forced using a linear relationship between mass balance and altitude based on observations over Sólheimajökull, a southern outlet glacier of the Mýrdalsjökull ice-cap. A continentality factor is introduced that raises inland equilibrium-line altitudes (ELAs) and also slackens the mass-balance/altitude gradient driving the model. With this calibration the present-day ice distribution can be reconstructed. Growth from ice-free to total ice cover is assessed and the sensitivity of ice extent to ELA change is shown to be non-linear. The model indicates that an ELA lowering of 500 m, consistent with a ca. 5 °C temperature depression for southern Iceland at the LGM, would enable glacier ice to cover the whole land surface within the modelled area and inundate any putative ecological refugia. Modelling with a 300 m ELA depression, consistent with previous reconstructions of Younger Dryas (YD, 10 000–11 000 ^{14}C yr BP) ice extent, indicates that the principal outcrops of the Sólheimar ignimbrite were glaciated at this time, suggesting that an origin during this stadial (and a correlation with North Atlantic Ash Zone I / Vedde Ash) is problematic. Copyright © 2003 John Wiley & Sons, Ltd.

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KEYWORDS: glacier modelling; Iceland; deglaciation; refugia; Sólheimar ignimbrite.

Introduction

It is important to study glacier fluctuations in Iceland for a number of reasons. Glaciers in Iceland advance and retreat in response to widespread changes in North Atlantic circulation patterns, which are increasingly being identified as key components of the global climate system (Bond *et al.*, 1997; Dowdeswell *et al.*, 1997; Ingólfsson *et al.*, 1997). If we can better understand the complex linkages between climatic changes and glacier fluctuations, glacial–geological evidence of past glacier extent can be used in conjunction with glacio-climatological modelling to reconstruct past climates over the island. In addition to this, improved constraints on the extent of Icelandic ice at the Last Glacial Maximum (LGM, also known as Weichselian Maximum, ca. 20 000 yr BP) and during subsequent decay stages are crucial to the understanding of biogeographical change in the region. Glacier extent provides two critical lines of evidence for palaeoecology. Firstly, ice-free, or nunatak areas are necessary to create the physical space for possible refugia. Secondly, if ice-free areas are identified, the inferred envelopes of climatic conditions necessary to reconstruct the surrounding glaciers also constrain ecological reconstructions by defining possible growing seasons, moisture regimes and temperature ranges.

To date, most attempts to reconstruct the glacial history of Iceland, particularly back to the LGM, have been hampered by sparse glacial–geological evidence, incomplete stratigraphy, and inadequate dating control, leading to contrasting interpretations. During global glacial maxima through the late Pliocene and into the Pleistocene, ice sheets covered much, if not virtually all, of the present land surface of Iceland (Ingólfsson, 1991; Geirsdóttir and Eiríksson, 1994) and the glacial–geological record forms the principal palaeoenvironmental archive of these periods. Little evidence survives of non-glacial terrestrial environments older than 10 000 ^{14}C yr BP (Norðdahl, 1990; Hjartarson and Ingólfsson, 1988; Ingólfsson, 1991; Rundgren and Ingólfsson, 1999). As a result, the direct record of glaciation preserved in landforms and sediments is critical to our understanding both of the LGM and the last glacial–interglacial transition on the island. Evidence of the extent, thickness and flow of former glaciers has been preserved both offshore and onshore (Fig. 1). Offshore, moraine banks have been identified, the extent of till

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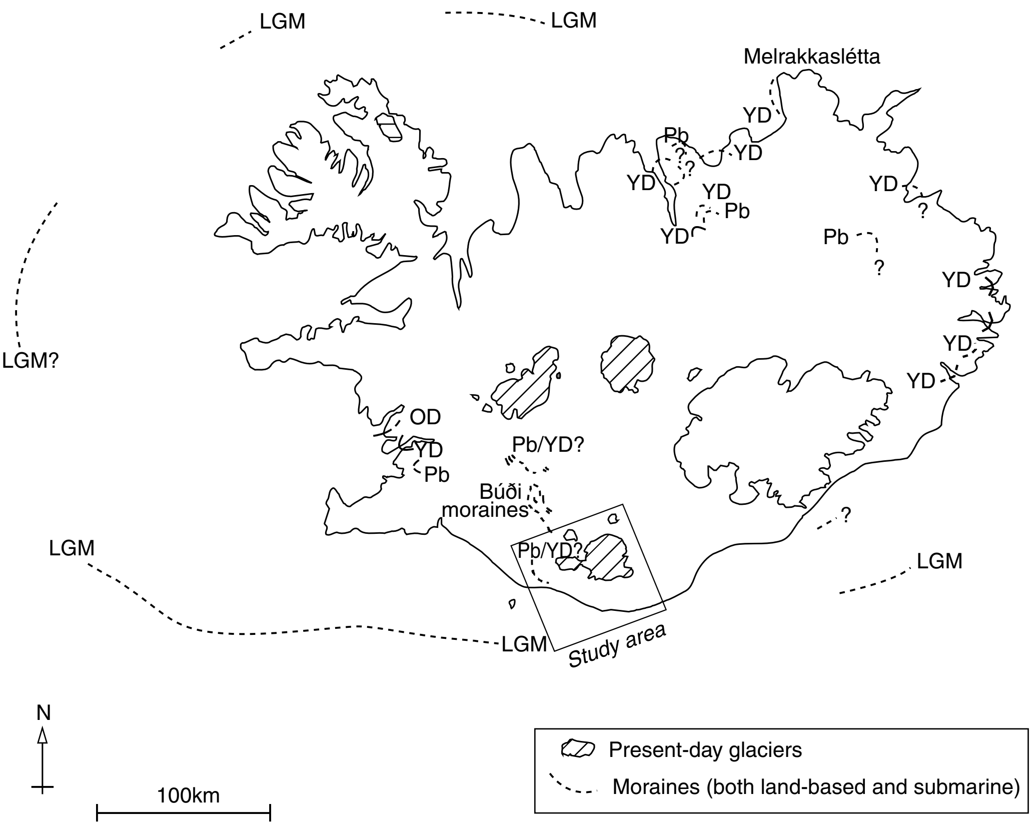

Figure 1 Iceland showing possible limits of major ice sheets, and the decay stages of the last Glacial–Interglacial transition. Notation: LGM, Last Glacial Maximum (inferred limits only; no dates available); OD, Older Dryas; YD, Younger Dryas; Pb, Preboreal; ?, undetermined. Inland moraine localities are from Norðdahl (1990), Geirsdóttir *et al.* (2000) and Norðdahl and Einarsson (2001); submarine moraines are from Andrews *et al.* (2000). Also shown is the study area examined in this paper (rectangle)

sheets has been delimited and minimum ages have been established for deglaciation (Ólafsdóttir, 1975; Egloff and Johnson, 1978; Boulton *et al.*, 1988; Andrews *et al.*, 2000). Despite this progress, the timing and maximum extent of the last glaciation are poorly constrained by offshore data. The last ice sheet may have extended to the current 200 m bathymetric contour, or glaciers in key sectors may have terminated close to the current coast (Einarsson and Albertsson, 1988; Norðdahl 1990; Ingólfsson, 1991; Andrews *et al.*, 2000) (Fig. 1). Onshore, ice thickness has been constrained by the heights of volcanoes that have formed table mountains, as these structures were formed partially below and partially above past glacier surfaces (Walker, 1965; Gudmundsson, 1997). Altitudes of table mountain plateau surfaces have been compared with striae, trimlines, blockfield limits and the extent of alpine glaciation in coastal areas to produce reconstructions of overall ice-sheet thickness at the LGM (Einarsson, 1959; Einarsson, 1963; Hoppe, 1982; Hjort *et al.*, 1985; Ingólfsson, 1988; Norðdahl, 1991; Norðdahl and Hjort, 1995). Crucially, however, much of this evidence is undated—or undatable—and the ice-surface gradients established cannot be used to test alternative offshore reconstructions of ice margins.

Whereas the maximum extent of ice thus remains unclear, empirical constraints on the late Pleistocene to early Holocene decay stages and readvances of the last ice sheets are more certain, as well-dated ice-marginal positions have been identified in the southwest and northeast of the island (Hjartarson and Ingólfsson, 1988; Norðdahl, 1990; Norðdahl, 1991; Ingólfsson and Norðdahl, 1994, Geirsdóttir *et al.*, 1997; Ingólfsson *et al.*, 1997; Geirsdóttir *et al.*, 2000; Hardardóttir *et al.*, 2001; Norðdahl and Einarsson, 2001). Preboreal (Pb, ca. 9 700–9 900 ^{14}C yr BP) and YD limits have been identified across Melrakkaslétta (northeast Iceland) and the Pb ice limit has also been identified at the prominent Búði morainal-complex in southwest Iceland (Hjartarson and Ingólfsson, 1988; Norðdahl, 1990; Eiríksson *et al.*, 1997; Geirsdóttir *et al.*, 1997) (Fig. 1). Earlier research in southwestern Iceland pointed towards poorly defined YD ice limits on the southwestern ice shelf (Hjartarson and Ingólfsson, 1988), but it is now thought that the Búði moraines also mark the southwestern YD ice limit, at which ice may have calved directly into a marine environment (Eiríksson *et al.*, 1997; Geirsdóttir *et al.*, 1997; Geirsdóttir *et al.*, 2000). Although these data therefore constrain Holocene ice-margins within most major fjord systems and on Melrakkaslétta, the northwestern and southeastern margins of the ice sheet yet remain unconstrained and give limited overall control on the pattern of the glaciation.

Supplementary approaches to the use of field data are therefore required to reconstruct the last Icelandic ice-sheet and its decay stages and to augment the available glacial–geological evidence. In this paper, we use a time-dependent numerical ice-sheet model to investigate the response of ice cover in southern Iceland to specified inputs of precipitation and temperature. To demonstrate the wider value of this approach to Quaternary science, we use the model to tackle the issue of whether significant ice-free areas could have existed in southern Iceland during the LGM and through the YD. The existence, or otherwise, of ice-free areas during these glacial maxima is of crucial importance both to theories of ecological refugia (cf. Rundgren and Ingólfsson, 1999) and correlations of onshore ignimbrite deposits with offshore ash zones (cf. Lacasse *et al.*, 1995).

The study area

We focus on the southern coastal district of Iceland containing four current glaciers: the Mýrdalsjökull ice-cap and the smaller glaciers and icefields of Eyjafjallajökull, Tindfjallajökull and Torfajökull (Fig. 2). This contemporary pattern of glaciation is important for model calibration. The maximum extent of the last ice sheet in this sector is unknown, but it is the site of putative refugia for plants (Steindórsson, 1963) and invertebrates (Lindroth, 1957; Lindroth, 1968), and the presence of alpine landscape elements has been used by Sigbjarnarson (1983) to argue for limited glaciation. Buckland and Dugmore (1991) have disputed the existence of refugia, stating that contemporary biogeographical patterns could instead be the product of ecological change driven by human impacts during the last millennium. Furthermore, the geomorphological record could reflect selective glacier erosion with landform preservation beneath thin, cold-based glaciers (Buckland and Dugmore, 1991). Terminal moraines below the current Markarfljót sandur surface have been identified in seismic data by Haraldsson (1981) and have been linked to a reconstruction of a former Markarfljót glacier based on trimlines, striae and the extent of subaerial lava flows (Jóhannesson, 1985). This glacial stage is undated, but it is probably either from the LGM or a stage during the overall sequence of ice-sheet decay (Jóhannesson, 1985), and it may mark the southern extent of the Búði morainal complex (Geirsdóttir *et al.*, 1997). Arguments for a restricted glaciation of this area during the YD have also been developed from the interpretation of ignimbrite deposits south of Mýrdalsjökull (Lacasse *et al.*, 1995).

This southern region of Iceland therefore exemplifies many of the problems currently faced in the reconstruction of the LGM: the maximum extent of ice coverage is unknown, and empirical evidence is limited and can be collated and interpreted in varied ways. The area thus presents an ideal environment for testing the utility of ice-sheet modelling to unresolved issues of Quaternary science in Iceland.

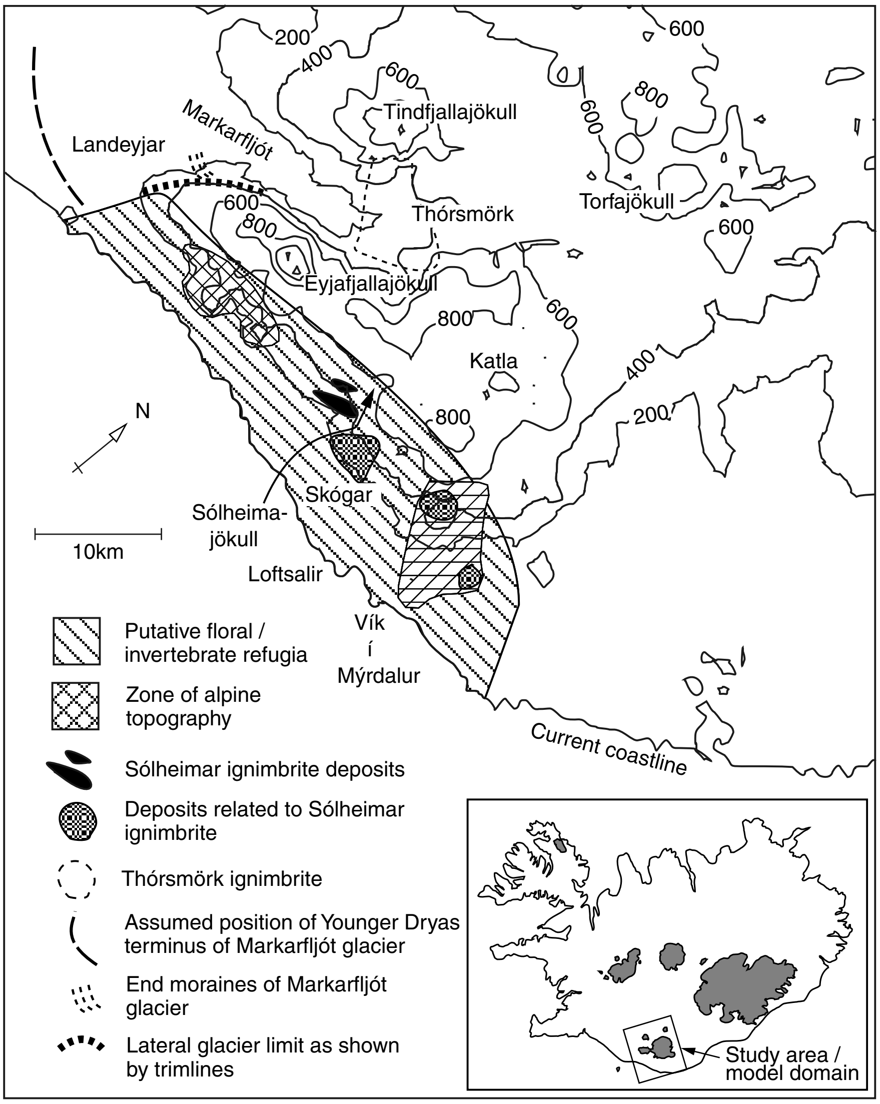

Figure 2 The study area, showing the bedrock topography in 200-m contours. Putative refugia (after Steindórsson, 1963; Lindroth, 1957), areas of alpine topography (after Sigbjarnarson, 1983), and outcrops of the Sólheimar ignimbrite (after Carswell, 1983; Jónsson, 1988) and related deposits (after Lacasse *et al.*, 1995) are all thought to define various limits to glaciation. Trimlines on Eyjafjöll (after Jóhannesson, 1985) and moraines identified by Haraldsson (1981) and Jóhannesson (1985) offer additional fragmentary empirical constraints to decay stages of the last glaciation of the region. The Thórsmörk ignimbrite has been extensively eroded since its deposition ca. 57 000 ^{14}C yr BP (distribution after Jørgensen, 1981). Today, the Mýrdalsjökull ice-cap overlies Katla, and small glaciers and icefields overlie Eyjafjalljökull, Tindfjallajökull and Torfajökull. Inset: the location of the study area within Iceland. This rectangle also denotes the domain of the model and shows the area covered by Figs 4, 5 and 7

Approach and methods

The model generates ice extent, volume and thickness in response to prescribed inputs of precipitation and air temperature. Glacial–geological evidence is compared with, but does not drive, the modelling. Our overall approach is heuristic, using the model to postulate glaciologically and climatologically consistent scenarios rather than to produce an 'exact' simulation. Field evidence can be used to constrain possible scenarios. We believe it is more important to focus on the sensitivity of the ice system to possible climatic fluctuations in an attempt to separate the improbable from the probable, rather than to attempt to produce rigid, deterministic simulations of past ice extent. In this way, we hope to demonstrate that numerical modelling and glacial geology can contribute mutually towards improved reconstructions of palaeo-ice extent.

The model

The model splits conceptually into two parts: an ice-sheet model and a climate model. The ice-sheet model (described in Mineter and Hulton, 2001) takes what is now a relatively standard approach to modelling large ice masses (cf. Huybrechts *et al.*, 1996). The climate model is based around the derivation of a simple integrated surface mass balance value based on altitude and continentality. We see the model as an heuristic device that can elucidate the major relationships between ice limits as interpreted from geomorphology and the climates and ice dynamics that might be responsible for such ice-sheet behaviour. Despite the relatively complex physics and non-linearity of ice-sheet behaviour, the ice-sheet model can be seen as a relatively 'dumb' responder to the imposed climate signals.

The ice-sheet model is three-dimensional, calculating time-dependent ice-mass continuity in response to a driving surface mass balance

$$\frac{\partial H}{\partial t} + \nabla \cdot q = \frac{\partial H}{\partial t} + \nabla \cdot (\bar{\mathbf{v}} H) = M \qquad (1)$$

where H is ice thickness (m), q is ice mass flux (m^3 yr^{-1}), ∇ is a two-dimensional divergence operator, t is time (yr), $\bar{\mathbf{v}}$ is depth-averaged horizontal velocity (m yr^{-1}) and M is the rate of surface mass balance (m yr^{-1}).

Equation (1) can be rewritten as a non-linear diffusion problem

$$\frac{\partial H}{\partial t} + \nabla \cdot D \nabla H = M \qquad (2)$$

where D is a 'diffusivity' equal to the scalar part of equation (1). The model uses the well-known, so-called 'shallow ice approximation' to derive vertically averaged ice flux for columns of ice based on surface slope and ice thickness (e.g. Huybrechts, 1986). Thus

$$D = \frac{2A(\rho g)^n}{n+2} H^{n+2} [\nabla(H) \cdot \nabla(H)]^{\frac{n-1}{2}} \qquad (3)$$

where A is the linear flow rate (equal to 10^{-16} Pa^{-3} yr^{-1}), g is acceleration as a result of gravity, ρ is ice density (910 kg m^{-3}) and n is the flow law exponent, normally $n = 3$.

The model then solves the combined surface mass balance and horizontal mass flux divergence over a finite-difference grid to derive a new set of ice thicknesses over a given time-step. The modelled ice sheet thus evolves through time.

As in previous modelling work (cf. Hulton *et al.*, 1994), we do not in this instance attempt to solve the thermal field and assume the ice to be isothermal, although the model can operate in a fully thermodynamic mode (cf. Mineter and Hulton, 2001). In effect, we assume that the ice sheet is uniformly relatively warm and is entirely warm-based. Despite this implicit assumption, for simplicity here we do not prescribe a sliding mechanism. Variations in basal melt rates resulting from variations in geothermal heat flux are not modelled, and are assumed to have a minimal effect on the modelled flow regime. These simplifications increase calculation speeds tremendously, and in experimental modelling runs the thermal dependence of ice stiffness on equilibrium forms, and the inclusion of basal sliding, were found to be negligible. However, these assumptions are potentially unreasonable, and their possible influences on the results are considered in the discussion.

For purposes of simplicity alone, isostatic readjustment is not included in these modelling experiments. The marine margin is also treated simply. The modelled ice sheet is not permitted to extend seawards as an ice shelf, and complete ice loss is forced to occur at 200 m water depth.

Derivation of mass balance

The ice-sheet mass-balance used to force the model (equation 1) is specified as an integrated function (separate accumulation and ablation components are not calculated) that is a linear function of altitude up to a specified threshold altitude

$$M = aZ \quad (0 < Z < x)$$
$$M_x = b \quad (Z \geq x) \qquad (4)$$

where M is mass balance (m yr^{-1}), Z is altitude relative to equilibrium-line altitude (ELA) (m), a and b are coefficients (in yr^{-1} and m yr^{-1} respectively), and x is the altitude (m) at which maximum mass balance, M_x, is reached. This provides a simple ice-thickness/mass-balance positive feedback mechanism to ice-sheet growth. The threshold values ($Z = x$, M_x) occur because of lower moisture supply and minimal ablation rates at higher altitudes, and so are used to define a

$$a = \frac{M_x}{x} \qquad (5)$$

The linear form of this relationship differs from previous ice-sheet models in which the mass-balance/altitude curve has been characterised as a truncated hyperbola (Boulton *et al.*, 1984, 1995; Hulton *et al.*, 1994). The rationale for using a linear relationship in this instance is based on mean values of annual mass-balance/altitude curves derived over the 31 yr period 1966–1996. Mackintosh (2000) collated these annual curves from meteorological data collected at weather stations at nearby Vík í Mýrdalur, Loftsalir/Vatnskarðshólar and Skógar (Fig. 2).

In general, the model is forced by controlling the mass-balance/altitude gradient and the level of the ELA at each model location.

The spatial variation of mass balance with continentality

The spatial variation in climate across southern Iceland is modelled by allowing the shape and overall level of

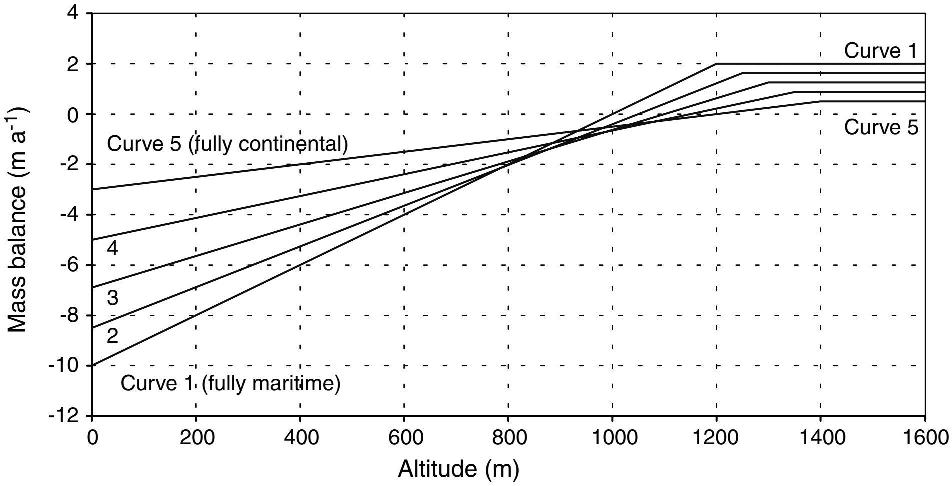

Figure 3 Form of the mass-balance/altitude relationship used to force the model. Curves 1 and 5 on this figure define end-members of an envelope of mass-balance/altitude experienced throughout the model domain. In this scenario, mass balance in areas with $C = 0$ (i.e. maritime areas) would vary with altitude as shown by curve 1. Conversely, areas with $C = 1$ (i.e. inland regions) would display a mass-balance/altitude relation as in curve 5. In areas where $0 < C < 1$, an intermediate curve is used, and the exact form of the curve is a linear function of continentality. Hence, for each model run, values of maximum mass balance M_x, maximum mass balance (or threshold) altitude x, and equilibrium-line altitude ELA, must be defined for the two end-member curves, for which C is either 0 or 1

the mass-balance/altitude relationship to vary as a function of continentality. The region containing Mýrdalsjökull and Eyjafjallajökull experiences strong climatic contrasts from south to north. With distance inland (or northwards) there is a strong reduction in precipitation, largely a result of the rain-shadow effect of the ice sheets themselves. The continentality, C, at any location is defined by the relative proportion of land and sea that lies within a hypothetical circle of radius R drawn around that location. Thus

$$C = \frac{Y_L}{Y} \tag{6}$$

where Y_L is the total land area contained within a circle of radius R, and Y is the total area of the same circle. On the basis of sensitivity tests, the radius of the circle, R, was selected to be 30 km, approximately the distance from the coastline to the current ice divides on Mýrdalsjökull and Eyjafjallajökull. The value of C thus varies between 0 (most maritime) and 1 (most continental). The overall mass balance field is defined by prescribing, for each continentality end-member ($C = 0$, $C = 1$), the ELA, the threshold altitude (x) and the maximum mass balance (M_x) at this altitude. Each of these values is defined to vary as a linear function of C between the two end members. More maritime areas, with $C \rightarrow 0$, take lower values of ELA and x, but higher values of M_x, than do continental curves. These differences in conditions and the resultant differences in mass-balance/altitude gradient all serve as proxies for the effect that variation in precipitation has on mass balance. In more maritime areas, greater precipitation equates to greater snowfall above the freezing level, tending also to lower the ELA overall. Conversely, such moisture also provides a sensible heat source and melt rates at lower elevations are consequently enhanced. A hypothetical range of mass-balance/altitude relationships is displayed in Fig. 3.

Topographic data

An ice-free topographic data set covering southern Iceland was required in order to provide inputs for basal altitude throughout the model domain. The model uses a 1-km topographic grid. A bathymetric data set for the south of Iceland was obtained as a 10-km digital elevation model (DEM) from the ETOPO-5 data set, published on the Web as part of the Global Resource Information Database (GRID) of the United Nations Environment Programme (UNEP). This 10-km DEM was reinterpolated to a 1-km DEM and merged with a 1-km DEM of southern Iceland obtained on the Web from the US National Imagery and Mapping Agency (NIMA). The areas covered by the Mýrdalsjökull and Eyjafjallajökull ice-caps were cut out and replaced with a subglacial topography. Subglacial topography for the Mýrdalsjökull ice-cap is based on material summarised by Dugmore and Sugden (1991) and data for Eyjafjallajökull was derived from Strachan (2001). The resulting data sets (subglacial topography and the combined DEM of southern Iceland) were then merged into the Lambert Azimuthal Projection, providing an ice-free 1-km DEM over the entire model domain (Fig. 4a).

Recent work by Björnsson *et al.* (2000) has demonstrated some discrepancies with our subglacial topography for Mýrdalsjökull. The main difference is that Björnsson *et al.*'s (2000) data show that subglacial topography on the southern crater rims of Katla is somewhat higher than in the same location in our data set, but this difference becomes immaterial once glaciers begin to grow. All of our reconstructions of the LGM begin with the current pattern of glaciation, so the issue of these basal ice topography discrepancies is of minor importance.

Model parameterisation and sensitivity tests

An initial step was to discover which values of the climatic parameters considered in the model would grow ice to an equilibrium distribution that best matches the present-day distribution of ice in southern Iceland. Therefore, multiple forward models were run with contrasting climatic inputs, and the geographical distribution of ice produced by each run was compared with the present-day ice distribution.

As a first experiment, the model was run using a maritime ($C = 0$) ELA value (=1000 m) based on the measured position of the ELA at Sólheimajökull, a southern outlet glacier of

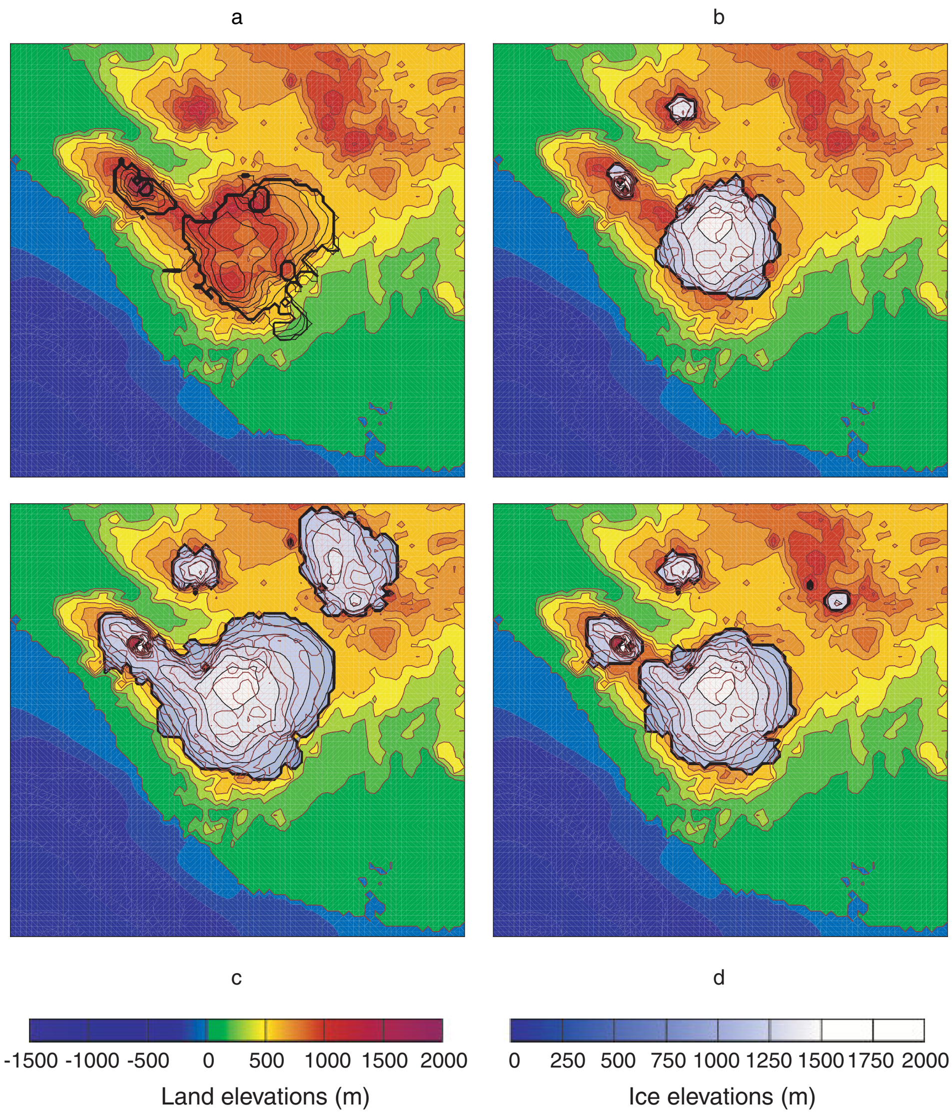

Figure 4 (a) A topographic model for the study area showing the limits of present glaciation. (b) Results of a modelling experiment where maritime ($C = 0$) ELA = 1000 m, $M_x = 4$ m yr^{-1} and $x = 100$ m; and continental ($C = 1$) ELA = 1200 m, $M_x = 1$ m yr^{-1} and $x = 1400$ m. The run time was 10 000 model years. This underestimates current ice distribution. (c) Revised modelling experiment where maritime ($C = 0$) ELA = 850 m, $M_x = 4$ m yr^{-1} and $x = 950$ m; and continental ($C = 1$) ELA = 1050 m, $M_x = 1$ m yr^{-1} and $x = 1250$ m. The run time was 10 000 model years. This overestimates current ice distribution. (d) The best-fit model reconstruction of current ice distribution is achieved with a parameterisation of continental ELA = 1150 m, $M_x = 1.5$ m yr^{-1}, $x = 1350$ m (200 m above continental ELA), and maritime ELA = 850 m, $M_x = 3$ m yr^{-1} and $x = 950$ m (100 m above maritime ELA). The result after a run time of 10 000 model years provides the best fit to current ice distribution

Mýrdalsjökull, whereas corresponding initial values of M_x (=4 m yr^{-1}) and x (=100 m) were estimated on the basis that above the ELA, in this highly maritime regime, mass balance tends quickly towards a maximum value. Continental ($C = 1$) values for these parameters were based on the premise that interior regions of Iceland are drier, therefore experience higher ELAs and maximum mass balance altitudes, but lower maximum mass balance. The values for ELA, M_x and x therefore were set initially to 1200 m, 1 m yr^{-1} and 1400 m respectively for the continental end-member.

The model was run for 10 000 model years and the resulting distribution of ice is shown in Fig. 4b. This modelled ice distribution provides a poor match with the current glacier configuration in southern Iceland. Ice extent is underestimated over Mýrdalsjökull and Eyjafjallajökull, and ice is absent over Torfajökull. This mismatch between model and reality is informative as it tells us that, although these parameters provide a reasonable first estimate, a lowering of ELA is required to match current glacier distribution. Thus further runs were carried out with progressive lowering of the ELA surface, until ice extent after 10 000 model years more closely matched that of the present day.

A better-fit model (Figure 4c) is obtained with maritime ($C = 0$) values of ELA = 850 m, $M_x = 4$ m yr^{-1} and $x =$ 950 m; and continental ($C = 1$) ELA = 1050 m, $M_x = 1$ m a^{-1} and $x = 1250$ m. However, modelled ice still overextends into areas that are in reality ice-free. Ice grows between Eyjafjallajökull and Mýrdalsjökull such that the two ice sheets merge into one, and large centres of ice develop over the Tindfjallajökull and Torfajökull massifs. With distance inland, the mismatch between model and reality becomes greater, suggesting that although the modelled ELA is reasonably estimated in maritime areas, the ELA is set too low for regions further inland. Consequently, a new series of model runs was initiated, with continental ELA increased by 50 m with each model run, a reduced maritime maximum mass balance and increased continental maximum mass balance. The best reconstruction of current ice distribution is achieved with a parameterisation of continental ELA = 1150 m, $M_x =$ 1.5 m yr^{-1}, $x = 1350$ m (200 m above continental ELA), and maritime ELA = 850 m, $M_x = 3$ m yr^{-1} and $x = 950$ m (100 m above maritime ELA). The result after a run time of 10 000 model years is shown in Fig. 4d. It is called the 'best-fit model' (BFM) as it produces a good reconstruction of the current distribution of ice in southern Iceland, assuming that these glaciers are currently in equilibrium with climate as expressed through mass balance.

Model results

A series of model runs was carried out in order to determine what magnitude of ELA lowering is required to develop glaciation to the current coastline. In this set of experiments, the model's initial conditions were set to the end result of the BFM. In other words, modelling was initiated from a point at which the distribution of ice across the model domain matches the present glaciation of the region. Figure 5 shows stages of ice development from present glaciation under two different conditions: Fig. 5a and Fig. 5b show stages in the development to Fig. 5c, which is driven by a 300 m lowering of the ELA in the BFM, run for 5000 y. Crucially, with the ELA 300 m lower than present, the model does not grow ice to the current coastline. In Fig. 5d, ELA is modelled at 500 m below the present level (hence maritime ELA is set at 350 m, and continental ELA is at 650 m). Under these circumstances ice reaches the current coast throughout the model domain.

Discussion

Relating ELA to climate

Changes in ELA used to force various model experiments can be used to define an envelope of possible changes in temperature and/or precipitation. The derivation of this relationship is based on the method used by Hulton *et al.* (1994). At one extreme, a disruption in ELA may be caused entirely by a change in air temperature. At the opposite extreme, ELA might fluctuate solely in response to changes in precipitation. If ELA alters only in response to fluctuations in temperature, and a lapse rate of 0.9 °C per 100 m altitude applies to southern Iceland, then an ELA lowering of 100 m could result directly from a cooling of 0.9 °C. Conversely, if ELA changes as a result only of fluctuations in precipitation, the change necessary can be estimated from conditions

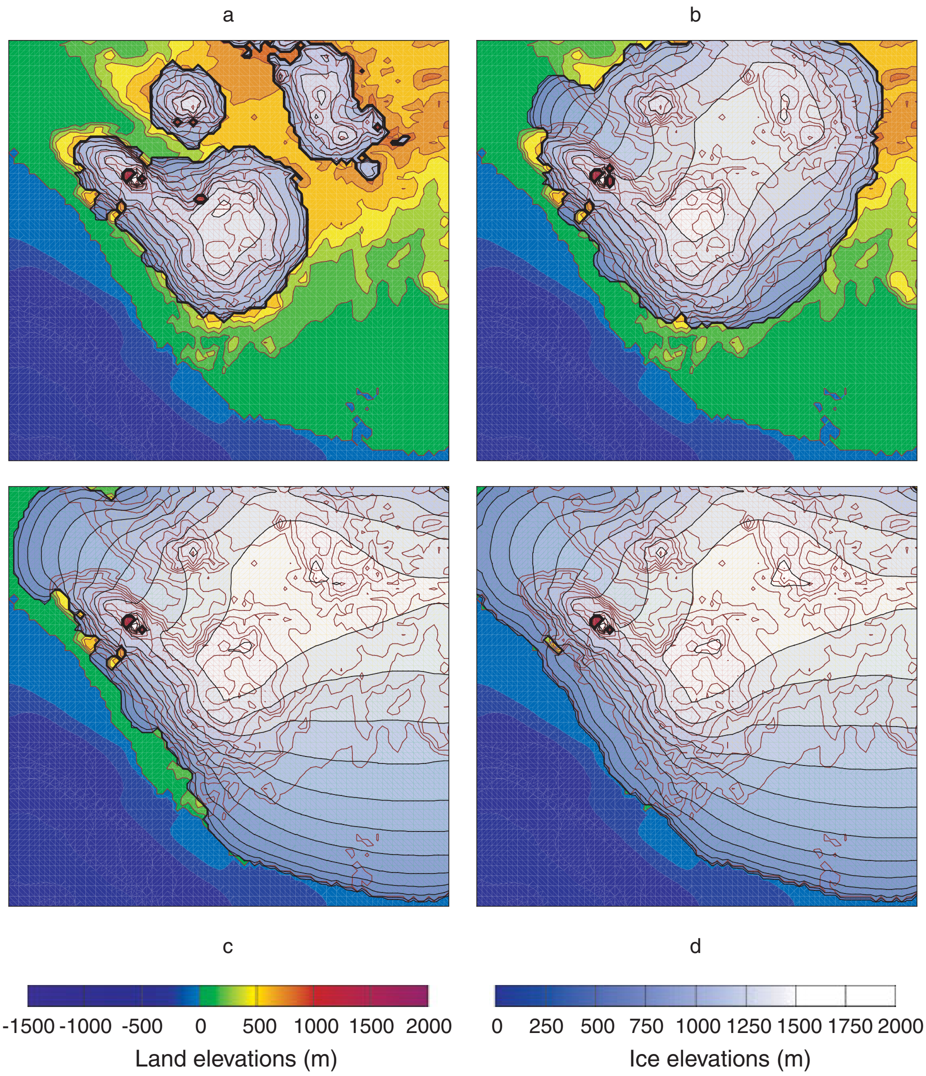

Figure 5 Stages in the development of a glaciation driven by a 300 m lowering of the ELA in the best-fit model are shown in (a) and (b). Equilibrium is reached within 5000 yr and is shown in (c). (d) The end result (equilibrium after 5000 model years) of a lowering of ELA to 500 m below the present level (maritime ELA = 350 m, continental ELA = 650 m). Ice reaches the current coastline throughout the modelled domain

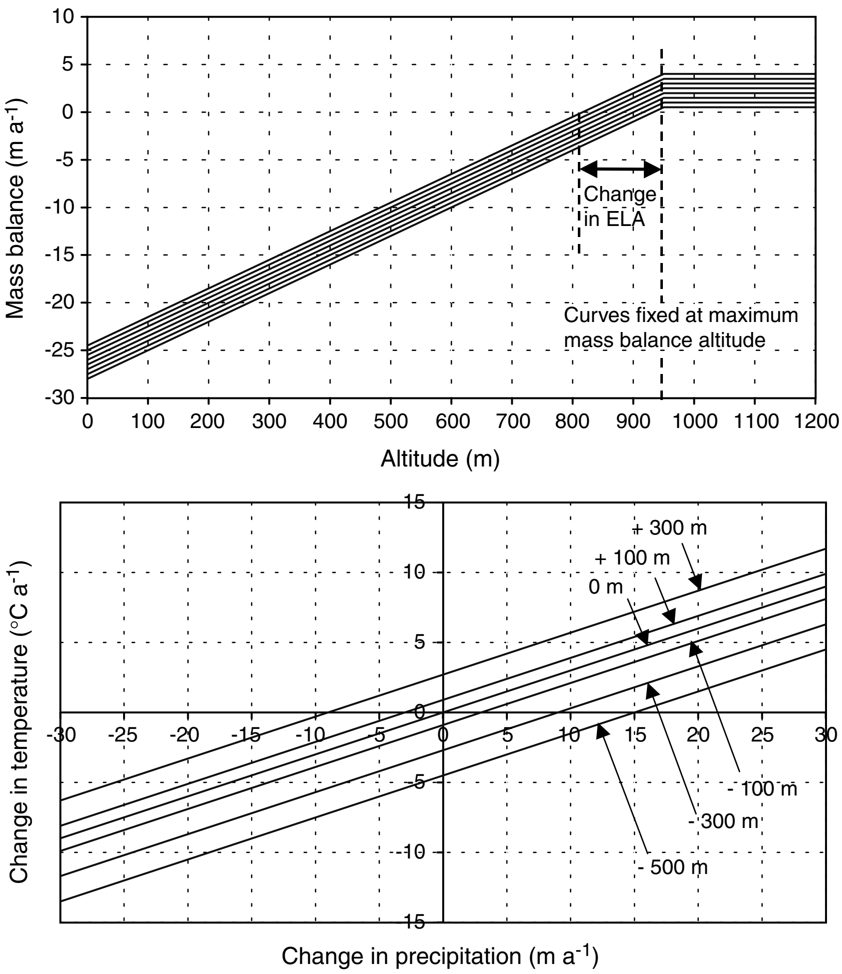

Figure 6 (a) The relationship of ELA to changes in mass balance. The value for M_x is fixed at 950 m, and it is assumed that no melting of ice takes place there. This assumption implies that, given constant temperatures, changes in mass balance are equivalent to changes in precipitation. Because the mass-balance/altitude relationship is linear, ELA changes are inversely proportional to precipitation changes. (b) The relationship between the ELA and the climatic envelope of temperature and precipitation in southern Iceland. See text for further explanation

at the maximum mass balance altitude x (Fig. 6a). If it is assumed that negligible melting occurs at this altitude, then any changes in mass balance can be fully attributed to changes in precipitation. Assuming a linear relationship between mass balance and altitude, ELA is inversely proportional to the amount of precipitation. This suggests that ELA lowering of 100 m could result from an increase in precipitation of 1.5 m yr^{-1} (Fig. 6a).

As Iceland is located in the zone swept by the changing location of atmospheric and oceanic polar fronts (Lamb, 1979; Broecker and Denton, 1990), neither temperature nor precipitation have remained constant for any length of time. Therefore, it is likely that ELA fluctuations have resulted from a combination of change both in temperature and precipitation. Based on the methodology of Hulton *et al.* (1994), it is possible to plot an envelope of temperature and precipitation fluctuations that relate to ELA changes (Fig. 6b). As a result, an ELA lowering of 100 m may be interpreted as resulting from any one of the following climatic changes:

1. cooling of 0.9 °C, no change in precipitation;
2. precipitation increase of 3 yr^{-1}, no change in air temperature;
3. cooling of 0.5 °C, precipitation increase of 1.7 m yr^{-1};
4. any alternative increase in precipitation combined with reduction in air temperature as defined by the −100 m line (Fig. 6b).

The model requires an ELA lowering of 500 m to grow ice to the current coast. This could be explained by a maximum temperature fall of 4.5 °C, or a maximum precipitation increase of 15 m yr^{-1}, or by intermediate values of these two parameters as defined by the line labelled −500 m (Fig. 6b). As maximum mass balance values over the world's ice sheets rarely exceed 6 m yr^{-1}, explanations of ELA lowering that invoke a precipitation flux in excess of this value must be treated with caution. Palaeoclimatic evidence from the Greenland ice-sheet and the North Atlantic suggest that LGM temperatures around southern Iceland may have been approximately 5–10 °C cooler than at present (Lamb, 1979; Broecker and Denton, 1990; Johnsen *et al.*, 1995; Rundgren and Ingólfsson, 1999). Assuming that precipitation levels did not change by more than a few metres per year, a ≥5 °C cooling probably caused the ELA surface across southern Iceland to fall sufficiently to glaciate the entire region.

Wider implications

Ice cover during the LGM

As our modelling experiments show that glaciers could extend to the present-day coastline with a 500 m ELA depression, the existence of coastal refugia in southern Iceland throughout the

LGM is not supported by the model. Although the model does not directly consider the effects of glacial isostasy on increasing relative sea-levels or areal shifts in coastline both as a result of changing relative sea-levels (cf. Hardardóttir *et al.*, 2001) and Holocene coastal aggradation (cf. Haraldsson, 1981), these processes would only have reduced further the amount of land available to support LGM coastal refugia. The weight of advancing ice would have acted to depress coastal land, leading to a northward encroachment of the coastline towards the ice limits, and the absence of Holocene aggradation on the coastal plain has exaggerated this change. Haraldsson's (1981) seismic data suggests that >50 m of Holocene aggradation has occurred in the Landeyjar district and similar, if not greater, Holocene aggradation has probably occurred in front of Sólheimajökull (Maizels and Dugmore, 1985; Maizels, 1991, 1994) and Mýrdalssandur (Maizels, 1994). It therefore is probable that during the LGM, even though absolute sea-level may have been lower, ice in southern Iceland calved directly into the sea, as it did during the YD (Geirsdóttir *et al.*, 2000).

The polycentric distributions of elements of the modern biota and separate concentrations of flightless forms of some species of invertebrates previously have been used to argue for the existence of refugia on the southern coast of Iceland during the LGM (Lindroth, 1957, Steindórsson, 1963; Ingólfsson, 1991; Rundgren and Ingólfsson, 1999). According to this line of reasoning, polycentric species distributions reflect Holocene expansions from geographically isolated ice-free areas. Alternatively, it has been argued that these patterns reflect the ecological changes of deforestation and soil erosion wrought by people since the ninth century AD Norse colonisation, which acted to isolate biota into separate enclaves (Buckland and Dugmore, 1991). The glacier modelling reported here supports the idea that the biogeographical patterns primarily reflect human impacts rather than serve as legacies of LGM refugia, as it effectively rules out the possibility that any parts of the area were ice-free during the LGM. The existence of areas of 'alpine landforms' (Sigbjarnarson, 1983) also has been used to argue for limited glaciation during the LGM. However, given the probable inundation of the whole study area by the last ice sheet, as suggested by the model, we suggest that these areas of alpine topography most probably reflect variations in effectiveness

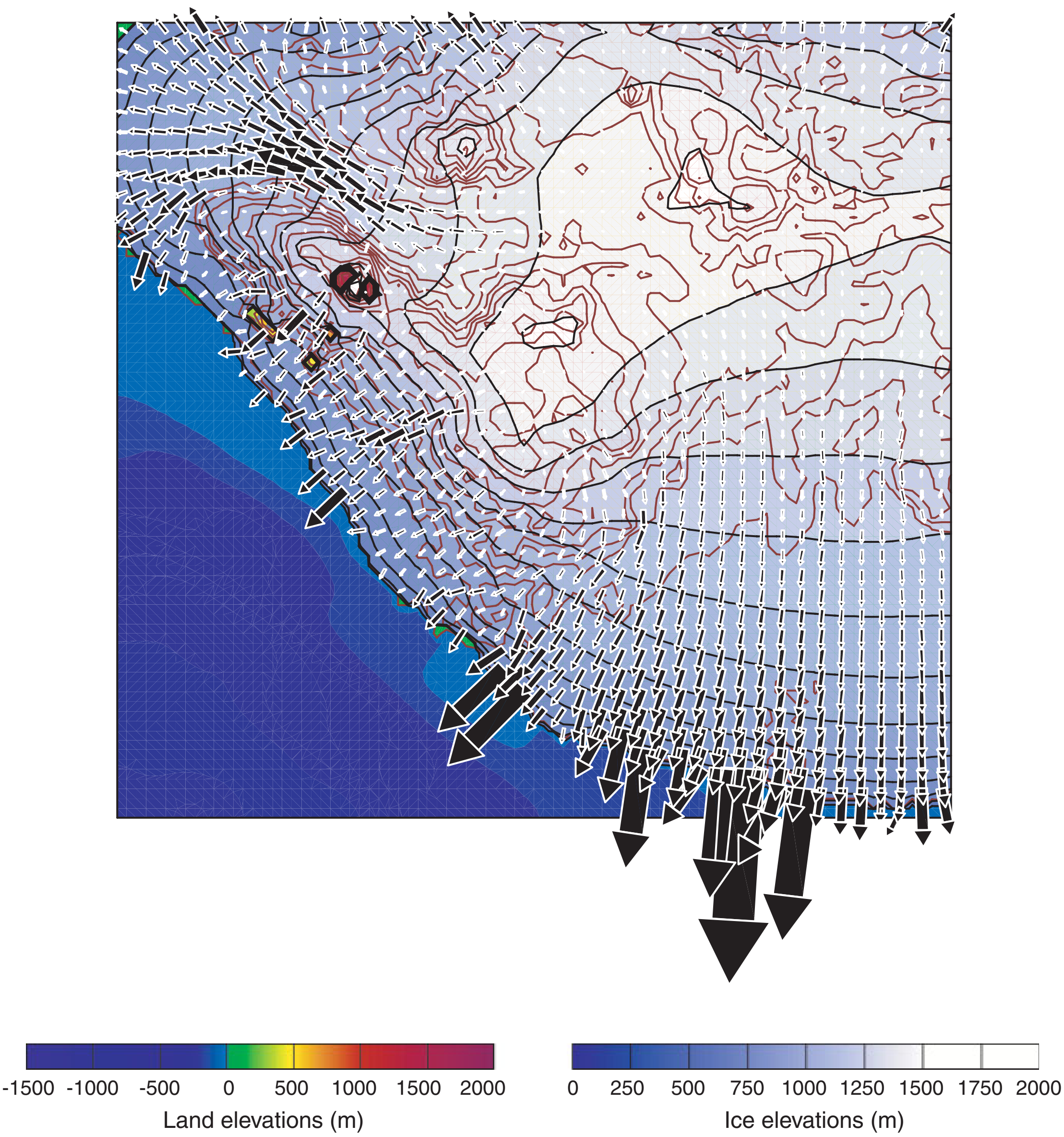

Figure 7 A reconstruction of the glaciation of southern Iceland with ELA 500 m lower than at present. Ice flux through each grid cell is illustrated with arrows where the arrow scale is proportional to the flux of ice. Notable ice streams develop in the Markarfljót valley in the northwest of the study area, and across Mýrdalssandur in the south of the study area. Bedrock contours (brown) are shown beneath the ice elevations (blue)

of ice-sheet erosion at ice maximum conditions rather than subaerial exposure at that time. 'Alpine' areas may have been preserved under thin, cold-based ice, whereas surrounding areas were preferentially eroded by thicker, warm-based ice (cf. Sugden and John, 1976). It is notable that the modelling experiments predict limited flow intensity where such areas of alpine topography previously have been identified (compare Fig. 7 with Fig. 2).

The model of total glaciation (Fig. 7) shows ice thicknesses in the order of 1 km and strong ice flow developing in the Sólheimajökull and Markarfljót valleys, indicating that these should be areas of significant glacial erosion. This idea is supported by radio-echo sounding of Sólheimajökull (Mackintosh *et al.*, 2000), which indicates the presence of a major subglacial trough extending to below present sea-level. Likewise, seismic studies by Haraldsson (1981) indicate the presence of a major bedrock trough lying beneath the current Markarfljót sandur. It is notable that these ice streams develop in the model despite the omission of these bedrock troughs from the topographic data set used by the model.

The model of total glaciation (Fig. 7) also shows limited ice flow over the region containing the Thórsmörk ignimbrite deposits (Fig. 2). The Thórsmörk ignimbrite was probably formed ca. 57 000 ^{14}C yr BP, and is thought to be the onshore correlative of North Atlantic Ash Zone II (Lacasse *et al.*, 1995). Despite being formed within the last glacial cycle and therefore probably subjected to <30 000 yr of glaciation, and underlying an area where modelled ice flow is limited, this ignimbrite therefore has been eroded through its complete thickness, which in places exceeds 100 m (Jørgensen, 1981). The ignimbrite has therefore been eroded at twice the rate of the bedrock walls of the main glacial troughs of northern Iceland (Bentley and Dugmore, 1998). This rapid erosion is not inconsistent with the limited overlying ice flow simulated by the model, because the ignimbrite deposits are poorly consolidated and susceptible to rapid erosion. Therefore strong ice flow is not required to erode large thicknesses of the Thórsmörk ignimbrite, thus the model results are consistent with its rapid post-depositional erosion.

Ice cover during the Younger Dryas

The extent of the last glaciation of the Markarfljót valley has been reconstructed by Jóhannesson (1985) using a combination of trimlines, striae and the extent of subaerial lava flows. These data are consistent with a putative terminal moraine buried beneath the sandur and located by Haraldsson (1981) using seismic data (Fig. 2). Haraldsson (1981) interpreted this moraine to represent the terminal position of the Markarfljót valley glacier during the YD. Our modelling of the glaciation resulting from a 300 m lowering of ELA leads to a glacier terminus in this vicinity (Fig. 5c). This 300 m lowering is consistent with the probable climatic conditions of this region during the Younger Dryas (Rundgren and Ingólfsson, 1999), so our modelling supports the conclusions of this earlier geomorphological research.

Our results also show that a narrow ice-free strip of land may have existed between the ice-sheet margin and the current coastline in southern Iceland during the YD (Fig. 5c). However, as noted above for the LGM chronozone, it is most likely that this area would in fact have been submerged during the YD, as relative sea-level during the YD may have been ca. 70 m higher than today (Geirsdóttir *et al.*, 2000; Hardardóttir *et al.*, 2001), and coastal sandar may have been less extensive (Haraldsson, 1981). Both of these factors would have led to a northward shift in the coastline towards the ice margin.

At the same glacial stage in the model, and therefore also linked to the YD chronozone, ice covers all of the hillsides rising above the sandur both to the east and west of Sólheimajökull (Fig. 5c). Outcrops of the Sólheimar ignimbrite have been mapped on these hillsides up to an altitude of 450 m (Carswell, 1983; Jónsson, 1988; Fig. 2), and previously have been correlated with the formation of North Atlantic Ash Zone I (also known as Vedde Ash) ca. 10 600 ^{14}C yr BP (Lacasse *et al.*, 1995) to suggest that the area below 450 m was ice-free, thus could have acted as a refugium, during the YD. That YD ice may have entirely covered this area casts doubt on this theory. Two possibilities are suggested here. If the area were covered in thick ice (as in Fig. 5c), a subaerial ignimbrite deposit could not have been formed at the same time as the ash zone was deposited across the neighbouring ocean floor. In this case, the Sólheimar ignimbrite may have been formed before the YD, and may be significantly older than North Atlantic Ash Zone I. This is supported by evidence that the surface of the ignimbrite has been glaciated since its deposition (Jónsson, 1988). Morphologically contrasting 'Sólheimar ignimbrite' deposits found at Víkurhóll ('pumice hill'), west of Sólheimajökull (cf. Newton, 1999), which lie outwith the modelled ice margin of this time, suggest that geochemically similar silicic Katla pumices may have been formed on at least two occasions in the pre-Holocene period. Alternatively, the ignimbrite may have been partially deposited over thin ice as ice cover retreated from the YD maximum, and could have melted its way down through or into the relatively thin ice margin. This would also explain the contrasting morphology between the deposits at Sólheimajökull and Víkurhóll (Newton, 1999). Either way, the Sólheimar ignimbrite cannot be used to support the hypothesis that ice-free conditions existed in southern Iceland at the YD Maximum, thus it is unlikely that refugia existed in this part of the island during glacial maxima.

Limitations of the model and avenues for future research

Any exercise in glacier modelling is by definition an exercise in simplifying reality, and the work presented here constitutes no exception. The interpretation of glacier fluctuations in Iceland is complex. Glaciers may be partially decoupled from climate as a result of topographic feedback caused by glacier growth or decay, changing glacier processes such as meltwater routing and subglacial geothermal activity, and simple topographic thresholds such as pinning points and calving lines (Mercer, 1961; Kuhn, 1985; Payne and Sugden, 1990; Kirkbride and Spedding 1996). In addition, Icelandic glaciers exhibit spatially varied degrees of sensitivity to climate change, and this sensitivity can change over short distances (<1 km) and limited periods of time (10–100 yr) (Sigurðsson and Jónsson, 1995). Our model takes into account the effect of continentality on landward values of precipitation and temperature, and prescribes maximum mass balance at given altitudes in separate runs, thus accounting for some spatial variance in sensitivity and topographic feedback caused by ice-cap growth. However, the model does not explicitly consider the effects of volcanic eruptions, geothermal activity or jökulhlaups on ice-sheet growth or decay.

Given that Mýrdalsjökull and Eyjafjallajökull are both currently underlain by active volcanoes (Katla and Eyjafjöll respectively), and that volcanic eruptions of Katla have historically led to jökulhlaups emanating both from the southeast and southwest of Mýrdalsjökull (Björnsson *et al.*, 2000), it is reasonable to suppose that significant basal melting may occur in response to heightened geothermal activity underneath the ice caps. Subglacial geothermal melting may generate large quantities of meltwater at the base, the routing of which to the ice margins may enhance basal sliding. Periodic volcanically instigated jökulhlaups therefore may lead to short-term (of the order of a few years) variations in flow dynamics, although their overall effect on millennial-scale ice-growth is probably negligible. Hence the absence of a sliding mechanism in the model is probably unrealistic, although some sensitivity experiments indicate that including this process has a minimal effect on the extent of the glacial configuration in response to climatic forcing. In effect, a similar extent is supported by an ice sheet with a slightly lower profile and greater variations in velocity across the model. It does not affect our substantive conclusions.

A further limitation to this research, to which we alluded above, is that we have not explicitly modelled here the effects of changes in relative sea-level off southern Iceland, despite recent glacial–geological evidence indicating that significant relative sea-level oscillations took place in the region throughout the period of investigation (Geirsdóttir *et al.*, 2000; Hardardóttir *et al.*, 2001). It has been suggested that during the YD, higher relative sea-level led to an inland incursion of the marine margin in southwestern Iceland, at least to the Búði morainal complex, and that YD ice may have calved actively into the sea at this point (Geirsdóttir *et al.*, 2000). Our model results show that with a 300 m ELA depression, which may represent the climatic situation in the YD, ice does not extend to the current coastline in southern Iceland. However, if, as proposed, the coastline were located further to the northeast during the YD, considerable volumes of ice may have calved actively into the sea within the model domain. This could have induced increased mass loss and enhanced ice flow towards the palaeocoastline, as occurs at the margins of present-day calving glaciers (Van der Veen, 1996). Changing sea-levels therefore could clearly have an impact on patterns of ice flow and ice flux in southern Iceland. However, this does not alter our conclusion that ecological refugia could not have existed in the area during the glacial advances. The model shows that with a 300 m ELA depression, ice completely overruns the areas of proposed refugia. Even taking into consideration the hypothesis that ice calved actively into a marine margin during the YD and/or the LGM, the proposed refugia would still have been overrun by ice. Nevertheless, future modelling efforts in the region would benefit from active consideration of changing sea-levels, their effects on lateral marine-front incursion, and improved definition of ice–sea interaction.

One further criticism that may be levelled at the modelling reported here is that it does not incorporate the possible effects of incursions of ice from outside the model domain. We need to evaluate the model's treatment of such effects. The model itself assumes a zero flux gradient across the boundary. This means it acts 'passively', neither enhancing nor restricting changes induced within the model domain. It cannot by definition be influenced by changes beyond the domain boundary. How reasonable is this assumption for the situations considered? Given that during the YD, and particularly the LGM, an ice divide probably existed to the northeast of the modelled domain (Rundgren and Ingólfsson, 1999), it is possible that large fluxes of ice may have entered the model domain from the north and northeast, casting doubt on our reconstructions of ice flow through southern Iceland. Large influxes of ice may have entered the northern sector of the modelled region, and it is probable that during glacial maxima, ice in the sector around Tindfjallajökull and Torfajökull flowed in a south to southwesterly direction, and was mostly sourced by flow from an ice divide to the north of the region, rather than flowing in a north to northwesterly direction, as shown in our reconstruction (Fig. 7). The ice stream flowing to the north of Tindfjallajökull in Fig. 7 is therefore likely to be an artefact of the modelling approach. However, the Katla–Eyjafjöll massif probably acted to deflect southwesterly ice flow during glacial maxima, such that the major ice streams along the Markarfljót and from Sólheimajökull, simulated in our results (Fig. 7), were probably real. A further complication is that major ice streams may have entered the model domain to the west (sourced from the southern lowlands west of Eyjafjallajökull) and from the east (originating from the site of the southern margin of the current Vatnajökull). Such enormous ice streams may have truncated the relatively smaller ice streams flowing southwest from Mýrdalsjökull—shown on our reconstruction of possible LGM conditions-but would surely themselves have eradicated any potential ecological refugia. To investigate these hypotheses in detail requires numerical modelling of glacier–climate interaction over the whole of Iceland, combined with an ability to down-scale the results back to southern Iceland.

It is thus clear that the model has a number of limitations. However, as an heuristic device that can be used to investigate the links between climatic changes and glacier response, and the results of which can be tested against and used to strengthen glacial-geological evidence from southern Iceland, the model performs well. If utilised with adequate consideration of their limitations, such models can be used to test and shed new light on controversial hypotheses such as the existence of ecological refugia. As such, this work is intended to act as a preliminary step in encouraging dialogue between numerical modellers, glacial geologists and Quaternary scientists.

Conclusions

1. It is possible to recreate the current glaciation of southern Iceland with an ELA of 850 m a.s.l. at the coast, and mass balance of 3 m yr^{-1} at 950 m a.s.l. Further inland, the modelled ELA surface rises to approximately 1150 m and maximum mass balance of 1.5 m yr^{-1} is attained at 1350 m altitude.
2. The reproduction of contemporary glaciation is an effective exercise in parameterisation as the glacier distribution includes four separate ice caps of contrasting physiography.
3. Modelling experiments beginning with the contemporary distribution of glaciers require an ELA depression of 300 m to bring the ice close to the present coast, and 500 m to cover all the land within the modelled domain.
4. A 500 m depression of ELA is consistent with a maximum temperature fall of 4.5 °C, or a maximum precipitation increase of 15 m yr^{-1}, or a combination of lesser changes in these two parameters. These changes could have been produced by the LGM in this region.
5. Ice-free ecological refugia are unlikely to have existed on the southern coast of Iceland at the LGM if regional ELA was depressed by 500 m and glaciers reached climatic equilibrium.
6. The existence of areas of 'alpine landforms' in southern Iceland most likely reflects variations in glacial erosion at ice maximum conditions rather than restricted glaciation and/or subaerial exposure at that time.

7. A putative terminal moraine in the Markarfljót valley, identified by Haraldsson (1981), could reflect glaciation with an ELA 300 m lower than today. This is consistent with a proposed formation during the Younger Dryas stadial. At this glacial stage, most of the current outcrops of the Sólheimar ignimbrite would be glaciated, indicating that an origin during the stadial (and a correlation with North Atlantic Ash Zone One/Vedde Ash) is problematic. The morphologically contrasting deposits of Vikurhóll ('pumice hill') lie outwith the modelled ice margin of this time, suggesting that geochemically similar pumices may have been formed on at least two occasions.
8. Ice-sheet modelling can act as an effective heuristic tool for examining a variety of climatic and geomorphological problems.

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References

Andrews JT, Hardardóttir J, Helgadóttir G, Jennings AE, Geirsdóttir Á, Sveinbjörnsdóttir ÁE, Schoolfield S, Kristjánsdóttir GB, Smith LM, Thors K, Syvitski JPM. 2000. The N and W Iceland Shelf: insights into Last Glacial Maximum ice extent and deglaciation based on acoustic stratigraphy and basal radiocarbon AMS dates. *Quaternary Science Reviews* **19**: 619–631.

Björnsson H, Pálsson F, Guðmundsson MT. 2000. Surface and bedrock topography of the Mýrdalsjökull ice cap, Iceland: the Katla caldera, eruption site and routes of jökulhlaups. *Jökull* **49**: 29–47.

Bentley MJ, Dugmore AJ. 1998. Landslides and the rate of glacial trough formation in Iceland. *Quaternary Proceedings* **6**: 11–16.

Bond G, Showers W, Cheseby M, Lotti R, Almasi P, deMenocal P, Priore P, Cullen H, Hajdas I, Bonani G. 1997. A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science* **278**: 1257–1266.

Boulton GS, Smith GD, Morland LW. 1984. The reconstruction of former ice sheets and their mass balance characteristics using a non-linearly viscous-flow model. *Journal of Glaciology* **30**: 140–152.

Boulton GS, Thors K, Jarvis J. 1988. Dispersal of glacially derived sediment over part of the continental-shelf of south Iceland and the geometry of the resultant sediment bodies. *Marine Geology* **83**: 193–223.

Boulton GS, Hulton NRJ, Vautravers M. 1995. Ice-sheet models as tools for palaeoclimatic analysis: the example of the European ice sheet through the last glacial cycle. *Annals of Glaciology* **21**: 103–110.

Broecker WS, Denton GH. 1990. The role of ocean-atmosphere reorganizations in glacial cycles. *Quaternary Science Reviews* **9**: 305–341.

Buckland PC, Dugmore AJ. 1991. 'If this is a refugium, why are my feet so bloody cold?' The origins of the Icelandic biota in the light of recent research. In *Environmental Change in Iceland: Past and Present*, Maizels JK, Caseldine C (eds). Kluwer: Dordrecht; 107–125.

Carswell DA. 1983. The volcanic rocks of the Sólheimajökull area, southern Iceland. *Jökull* **33**: 61–71.

Dowdeswell JA, Hagen JO, Björnsson H, Glazovsky AF, Harrison WD, Holmlund P, Jania J, Koerner RM, Lefauconnier B, Ommanney CSL, Thomas RH. 1997. The mass balance of circum-Arctic glaciers and recent climate change. *Quaternary Research* **48**: 1–14.

Dugmore AJ, Sugden DE. 1991. Do the anomalous fluctuations of Sólheimajökull reflect ice-divide migration? *Boreas* **20**: 105–113.

Egloff J, Johnson GL. 1978. Erosional and depositional structures of the Southwest Iceland Insular Margin: thirteen geophysical profiles. In *Geological and Geophysical Investigations of Continental Margins*, Watkins JS, Montadert L, Dickerson PW (eds). American Association of Petroleum Geologists: OK; 43–63.

Einarsson T. 1959. Studies of the Pleistocene in Eyjafjörður. *Societas Secretarium Islandica* **33**.

Einarsson Th. 1963. Pollen-analytical studies on vegetation and climate history of Iceland in late and post-glacial times. In *North Atlantic Biota and their History*, Löve A, Löve D (eds). Pergamon: Oxford; 255–365.

Einarsson T, Albertsson KJ. 1988. The glacial history of Iceland during the past 3 million years. *Philosophical Transactions of the Royal Society of London, Series B* **318**: 637–644.

Eiríksson J, Símonarson LA, Knudsen KL, Kristensen P. 1997. Fluctuations of the Weichselian ice sheet in SW Iceland: a glaciomarine sequence from Sudurnes, Seltjarnarnes. *Quaternary Science Reviews* **16**: 221–240.

Geirsdóttir Á, Eiríksson J. 1994. Growth of an intermittent ice sheet during the late Pliocene and early Pleistocene. *Quaternary Research* **42**: 115–130.

Geirsdóttir Á, Hardardóttir J, Eiríksson J. 1997. The depositional history of the Younger Dryas–Preboreal Búði moraines in south-central Iceland. *Arctic and Alpine Research* **29**: 13–23.

Geirsdóttir Á, Hardardóttir J, Sveinbjörnsdóttir ÁE. 2000. Glacial extent and catastrophic meltwater events during the deglaciation of Southern Iceland. *Quaternary Science Reviews* **19**: 1749–1761.

Gudmundsson HJ. 1997. A review of the Holocene environmental history of Iceland. *Quaternary Science Reviews* **16**: 81–92.

Haraldsson H. 1981. The Markarfljót sandur area, southern Iceland. Sedimentological, petrological and stratigraphic studies. *Striae* **15**: 1–60.

Hardardóttir J, Geirsdóttir Á, Sveinbjörnsdóttir ÁE. 2001. Seismostratigraphy and sediment studies of Lake Hestvatn, southern Iceland: implications for the deglacial history of the region. *Journal of Quaternary Science* **16**: 167–179.

Hjartarson Á, Ingólfsson Ó. 1988. Preboreal glaciation of southeastern Iceland. *Jökull* **38**: 1–16.

Hjort C, Ingólfsson Ó, Norðdahl H. 1985. Late Quaternary geology and glacial history of Hornstrandir, northwest Iceland. A reconnaissance study. *Jökull* **35**: 9–29.

Hoppe G. 1982. The extent of the last inland ice sheet of Iceland. *Jökull* **32**: 3–11.

Hulton N, Sugden D, Payne A, Clapperton C. 1994. Glacier modeling and the climate of Patagonia during the Last Glacial Maximum. *Quaternary Research* **42**: 1–19.

Huybrechts P. 1986. *A Three-dimensional Time-dependent Numerical Model for Polar Ice Sheets: some Basic Testing with a Stable and Efficient Finite Difference Scheme*. Technical Report 86/1, Vrije Universiteit Brussel Geografisch Instituut: Brussels.

Huybrechts P, Payne T. The EISMINT Intercomparison Group. 1996. The EISMINT benchmarks for testing ice-sheet models. *Annals of Glaciology* **23**: 1–14.

Ingólfsson Ó. 1988. Glacial history of the lower Borgarfjörður area, western Iceland. *Geologiska Föreningens I Stockholm Förhandlingar* **110**: 293–309.

Ingólfsson Ó. 1991. A review of the late Weichselian and early Holocene glacial and environmental history of Iceland. In *Environmental Change in Iceland: Past and Present*, Maizels JK, Caseldine C (eds). Kluwer: Dordrecht; 13–29.

Ingólfsson Ó, Norðdahl H. 1994. A review of the environmental history of Iceland, 13 000–9000 yr BP. *Journal of Quaternary Science* **9**: 147–150.

Ingólfsson Ó, Björck S, Haflidason H, 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.

Jóhannesson H. 1985. Um endaslepptu hraunin undir Eyjafjöllum og jökla siðasta jökulskeiðs. *Jökull* **35**: 83–95.

Johnsen SJ, Dahl-Jensen D, Dansgaard W, Gundestrup N. 1995. Greenland paleotemperatures derived from GRIP borehole temperature and ice core isotope profiles. *Tellus* **47B**: 624–629.

Jónsson J. 1988. *Geological Map Eyjafjöll*. Neðri-Ás Research Institute: Hveragerði, Iceland.

Jørgensen KA. 1981. The Thórsmörk ignimbrite: a review. In *Tephra Studies*, Self S, Sparks RSJ (eds). Reidel: Dordrecht.

Kirkbride M, Spedding N. 1996. The influence of englacial drainage on sediment-transport pathways and till texture of temperate valley glaciers. *Annals of Glaciology* **22**: 160–166.

Kuhn M. 1985. Mass budget imbalances as criterion for a climatic classification of glaciers. *Geografiska Annaler* **66A**: 229–238.

Lacasse C, Sigurðsson H, Jóhannesson H, Paterne M, Carey SN. 1995. The source of Ash Zone I in the North Atlantic. *Bulletin of Volcanology* **57**: 18–32.

Lamb HH. 1979. Climatic variation and changes in the wind and ocean circulation: the Little Ice Age in the northeast Atlantic. *Quaternary Research* **11**: 1–20.

Lindroth CH. 1957. *The Faunal Connections Between Europe and North America*. Wiley: Chichester.

Lindroth CH. 1968. The Icelandic form of *Carabus problematicus* Hbst. (Col. Carabidae). A statistical treatment of subspecies. *Opuscala entomologica* **33**: 157–182.

Mackintosh A. 2000. *Glacier fluctuations and climate change in Iceland*. Unpublished PhD thesis, University of Edinburgh.

Mackintosh AN, Dugmore AJ, Jacobsen FM. 2000. Ice-thickness measurements on Sólheimajökull, southern Iceland and their relevance to its recent behaviour. *Jökull* **48**: 9–15.

Maizels JK, Dugmore AJ. 1985. Lichenometric dating and tephrochronology of sandur deposits, Sólheimajökull, southern Iceland. *Jökull* **35**: 69–78.

Maizels JK. 1991. The origin and evolution of Holocene sandur deposits in areas of jökulhlaup drainage, Iceland. In *Environmental Change in Iceland: Past and Present*, Maizels JK, Caseldine C (eds). Kluwer: Dordrecht; 267–302.

Maizels JK. 1994. The geomorphic significance of jökulhlaup drainage on Icelandic sandar: morphological and stratigraphic basis for modelling long-term (Holocene) sandur evolution. *Münchener Geographische Abhandlungen* **B12**: 177–204.

Mercer JH. 1961. The response of fjord glaciers to changes in the firn limit. *Journal of Glaciology* **10**: 850–858.

Mineter MJ, Hulton NRJ. 2001. Parallel processing for finite difference modelling of ice sheets. *Computers and Geosciences* **27**: 829–838.

Newton AJ. 1999. *Ocean-transported pumice in the North Atlantic*. Unpublished PhD thesis, University of Edinburgh.

Norðdahl H. 1990. Late Weichselian and early Holocene deglaciation history of Iceland. *Jökull* **40**: 27–49.

Norðdahl H. 1991. A review of the glaciation maximum concept and the deglaciation of Eyjafjörður, north Iceland. In *Environmental Change in Iceland: Past and Present*, Maizels JK, Caseldine C (eds). Kluwer: Dordrecht; 31–47.

Norðdahl H, Einarsson T. 2001. Concurrent changes of relative sea-level and glacier extent at the Weichselian-Holocene boundary in Berufjordur, Eastern Iceland. *Quaternary Science Reviews* **20**: 1607–1622.

Norðdahl H, Hjort C. 1995. Lateglacial raised beaches and glacier recession in the Þistilfjörður-Bakkaflói area, northeast Iceland. *Jökull* **43**: 33–44.

Ólafsdóttir Þ. 1975. Jökulgarður á sjávarbotni út af Breiðafirði (English summary: A moraine ridge on the Iceland shelf, west of Breiðafjörður). *Náttúrufræðingurinn* **45**: 31–36.

Payne A, Sugden D. 1990. Topography and ice-sheet growth. *Earth Surface Processes and Landforms* **15**: 625–639.

Rundgren M, Ingólfsson Ó. 1999. Plant survival in Iceland during periods of glaciation? *Journal of Biogeography* **26**: 387–396.

Sigbjarnarson G. 1983. The Quaternary alpine glaciation and marine erosion in Iceland. *Jökull* **33**: 87–98.

Sigurðsson O, Jónsson T. 1995. Relation of glacier variations to climate changes in Iceland. *Annals of Glaciology* **21**: 263–270.

Steindórsson S. 1963. Ice age refugia in Iceland as indicated by the present distribution of plant species. In *North Atlantic Biota and their History*, Löve A, Löve D (eds). Pergamon: Oxford; 303–320.

Strachan SM. 2001. A geophysical investigation of the Eyjafjallajökull glaciovolcanic system, South Iceland, using radio echo sounding. Unpublished PhD thesis, University of Edinburgh; 209 pp.

Sugden DE, John BS. 1976. *Glaciers and Landscape*. Arnold: London.

Van der Veen CJ. 1996. Tidewater calving. *Journal of Glaciology* **42**: 375–385.

Walker GPL. 1965. Some aspects of Quaternary volcanism in Iceland. *Transactions of the Leicester Literary and Philosophical Society* **59**: 25–40.

