Reviewed research article

A calibrated mass balance model for Vatnajökull, Iceland

Martijn de Ruyter de Wildt[1], Johannes Oerlemans[1] and Helgi Björnsson[2]

[1] *Institute for Marine and Atmospheric Research, Universiteit Utrecht, Utrecht, The Netherlands; m.s.deruijterdewildt@phys.uu.nl*

[2] *Science Institute, University of Iceland, Dunhaga 3, 107 Reykjavík, Iceland; hb@raunvis.hi.is*

Abstract – *Vatnajökull (Iceland) is the largest ice cap in the world where the energy and mass balance have been studied with good spatial and temporal resolution. In this paper we use these data to analyze the energy balance and to construct a calibrated and spatially distributed mass balance model. The incoming longwave radiation is best modeled as a function of meteorological variables in the free atmosphere just above the relatively thin katabatic layer, instead of those at the 2 m level. The ratio of changes in the 2 m temperature to changes in the free atmospheric temperature (the climate sensitivity) is smaller than 1. Therefore, when the bulk method is used to compute the turbulent fluxes, the 2 m temperature must be explicitly calculated. Otherwise the sensitivity of Vatnajökull to climatic change would be overestimated. When the model is forced with data from a permanent weather station not on the ice cap, it reproduces the observed mass balance reasonably well. Horizontal precipitation gradients over Vatnajökull are large, which results in a strongly varying sensitivity to external temperature changes over the ice cap. The mass balance and its sensitivity is thus highly dependent on local climatic conditions. For a temperature increase of 1 K and a simultaneous precipitation increase of 5.3%, the mean specific mass balance of Vatnajökull decreases by 0.56 m w.e.*

INTRODUCTION

Many authors have studied the energy and mass balance of glaciers. In earlier studies single points on a glacier were studied (e.g., Ambach, 1963; Munro and Davies, 1978). However, the mass balance depends strongly on altitude which is why others studied the energy balance along glacier transects (e.g., Braithwaite and Olesen, 1990; Munro, 1990; Greuell *et al.*, 1997) and modeled mass balance gradients and sensitivities (e.g., Ambach and Kuhn, 1985; Van de Wal and Oerlemans, 1994; Jóhannesson, 1997). Only a few authors have studied the surface of a glacier or ice cap in a three-dimensional way (e.g., Arnold *et al.*, 1996, Oerlemans *et al.*, 1999). Oerlemans *et al.* (1999) carried out a glacio-meteorological experiment on Vatnajökull (Iceland), which is the first ice cap where the melt process has been observed with good spatial and temporal resolution. In this work we use data from this experiment to analyze the energy balance and present a calibrated mass balance model which is based on a calculation of the surface energy balance. We force this model with data from meteorological stations that are close to but not on the ice cap. For these stations long meteorological records are available with which we can calculate the mean specific mass balance over the last decades and study the sensitivity of Vatnajökull to external climatic change. Precipitation over Vatnajökull is largely unknown, which is why we use this quantity to tune the model to the mass balance data. The mass balance is evaluated at a number of sites distributed across the ice cap. A Digital Elevation Model (DEM) with a horizontal resolution of 500 m is used to obtain the altitude and other geographical features for these points (Figure 1). The mean specific mass balance is then calculated with an interpolation scheme that was especially developed for this purpose.

Vatnajökull (Figure 1) is located in southeastern Iceland and is one of the largest temperate ice caps in the world (8200 km^2). The altitude ranges from

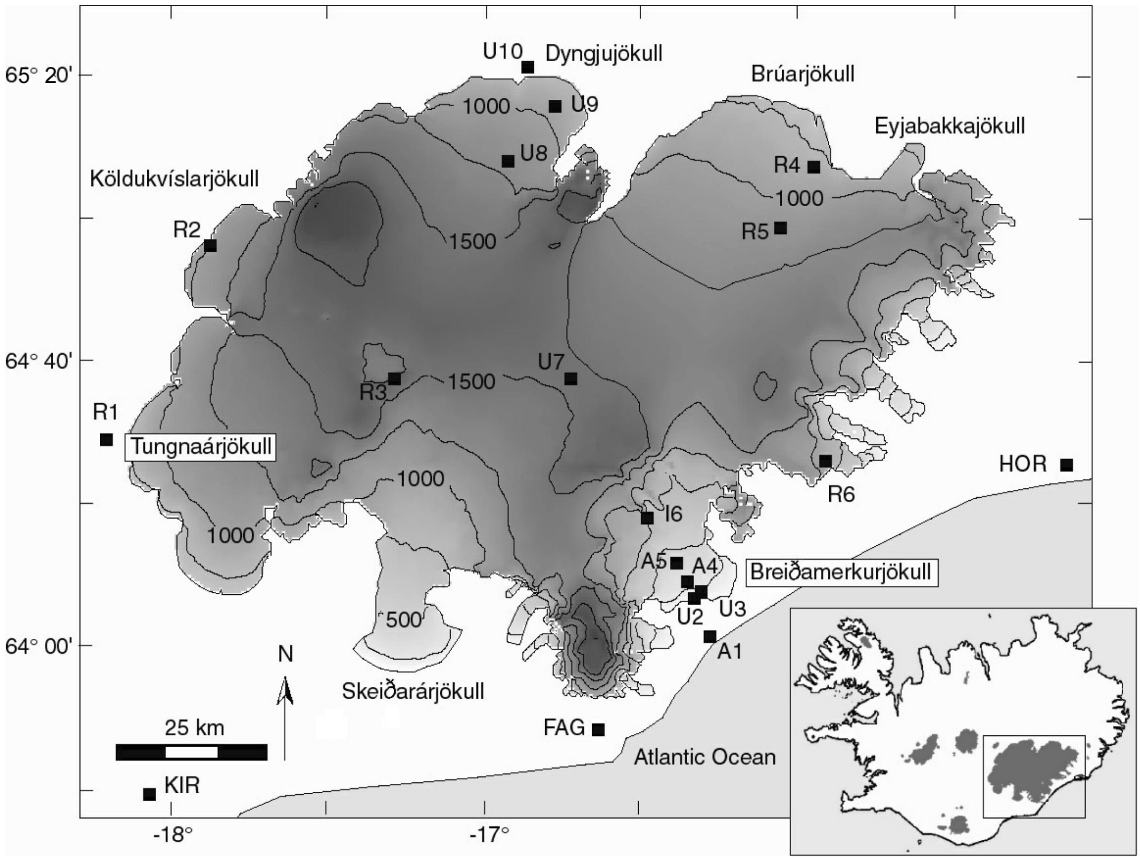

Figure 1. Map of Vatnajökull based on the DEM used by the mass balance model. The horizontal resolution is 500 m and contour lines are shown for each 250 m interval in altitude. Indicated are the sixteen weather stations operated during the 1996 experiment and the names of the lobes. The permanent weather stations of Hornafjörður, Fagurhólsmýri and Kirkjubæjarklaustur are indicated with HOR, FAG and KIR, respectively. – *Kort af Vatnajökli sem sýnir legu sex sjálfvirkra veðurstöðva á jökli og þriggja veðurathuganastöðva utan jökuls.*

sea-level to 2000 m with 88% of the surface between 700 and 1700 m. It comprises several domes overlying volcanic caldera's, some large surging-type lobes and many smaller outlet glaciers. Since the end of the nineteenth century it has been the topic of much research, but only in the latest decades have data been gathered in a systematic way. Overviews of previous research are given by Björnsson *et al.* (1998a) and Williams *et al.* (1997). The experiment reported in this paper was carried out in 1996 by research groups from the universities of Iceland, Utrecht (The Netherlands), Innsbruck (Austria) and from the Vrije Universiteit of Amsterdam (The Netherlands). Twelve Automatic Weather Stations were placed on the ice cap and four close to it (Figure 1). All of these were operated from May 22nd until September 1st and measured, amongst other variables, 2 m temperature, 2 m humidity, 2 m windspeed, pressure and incoming and outgoing shortwave and longwave radiation. Furthermore, close to U2 radiosonde ascents were made twice daily. These produced profiles of temperature, humidity, windspeed and pressure up to an altitude of

10 km. Balloon soundings were made frequently near U3 and near I6 to probe the lowest 500 m of the atmosphere. Cloud observations were made approximately every 3 hours near station U2 and irregularly near station R5. The set-up of the experiment and the data are discussed in more detail in Oerlemans *et al.* (1999).

ENERGY BALANCE PARAMETERIZATIONS

Surface melt occurs when the energy balance at the surface (F) is larger than zero. F is given by

$$F = (1-\alpha)Q + I_{in} + I_{out} + H_s + H_l \quad (1)$$

where α is the surface albedo, Q the global radiation, I_{in} the incoming longwave radiation, I_{out} the outgoing longwave radiation, H_s the turbulent flux of sensible heat and H_l the turbulent flux of latent heat.

Global radiation

Global radiation is computed according to the parameterization given by Greuell *et al.* (1997) which is based on Meyers and Dale (1983). This parameterization is valid for horizontal surfaces. Most of the surface of Vatnajökull slopes very gently, so we apply a first order approximation to take the surface inclination into account. The surface inclination only influences the direct solar radiation, so we distinguish between direct and diffuse shortwave radiation:

$$Q = Q_0 \left[c_1 \cos(\theta_s - \theta_{srf}) + c_2 \cos\theta_s \right] TF \quad (2)$$

where $TF = T_r T_g T_w T_{as} T_{cl} F_{ms} F_{rs} F_{ho}$ and $c_1 + c_2 = 1$. Q_0 denotes the solar irradiance at the top of the atmosphere, θ_s is the solar zenith angle, and θ_{srf} is the inclination of the surface in the direction of the sun. θ_s can be easily calculated from standard astronomical theory (e.g., Walraven, 1978). For clear skies c_1 is 0.9 and c_2 is 0.1 while for entirely overcast skies c_1 is 0 and c_2 is 1. $T_r T_g$, T_w, T_{as} and T_{cl} are transmission coefficients that account for Rayleigh scattering and absorption by other gases than water vapor, absorption by water vapor, aerosol extinction and absorption by clouds, respectively. The first three coefficients can directly be calculated from meteorological variables. T_{as} and T_{cl} are tuned to the data in the same way as described in Greuell *et al.* (1997). The coefficients F_{ms}, F_{rs} and F_{ho} account for the amplification through multiple scattering at the surface, amplification through directly reflected radiation at the surface and attenuation by horizon obstruction. These three coefficients can be obtained from the DEM. Figure 2a displays hourly means of the observed and simulated global radiation for the stations for which cloud observations are available.

Albedo

Snow albedo (α_{sn}) depends on several climatological and surficial quantities and thus changes in time and space. These quantities are grain size, impurity content, cloudiness, solar inclination, liquid water content and surface roughness (e.g., Warren, 1982). The influence of most of these quantities on α_{sn} is not well understood and some simplifications have to be made when α_{sn} is parameterized. We use a parameterization that describes the daily albedo as a function of aging of the snow and snowdepth (Oerlemans and Knap, 1998). We calibrate this parameterization to the data measured on Vatnajökull.

The ice albedo displays large spatial variations over Vatnajökull which cannot be described with a simple parameterization. In the north and northwest (U8, U9 and R2) the mean ice albedo is very low (<0.10). As has been confirmed *in situ*, these low values are caused by black volcanic deposits, tephra. The tephra melts out and accumulates in the ablation zone. To the south and to the east, the ice is cleaner, but the albedo has not been measured on all major outlets and it varies a lot in some areas. For example, the albedo values measured at the stations A4, A5, I6 and R4 are most likely not representative of their surroundings (Reijmer *et al.*, 1999). We therefore use satellite reflectance images to determine the ice albedo of different parts of Vatnajökull. We do not study the snow albedo in this way, because the snow albedo is more homogeneous on a horizontal scale than the ice albedo. Furthermore, there are too few usable satellite images to resolve changes in snow albedo. We use NOAA-AVHRR images which have a resolution of 1.1 km at nadir. The retrieval method is described by De Ruyter de Wildt *et al.* (2002). Figure 3 shows one

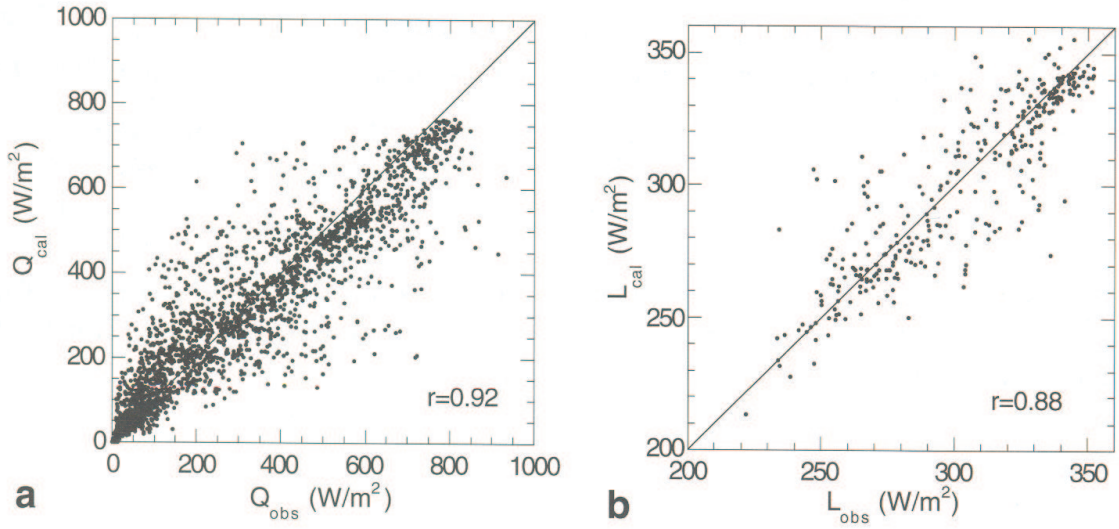

Figure 2. Scatterplots of calculated against observed hourly means of global radiation (a) and of the incoming longwave radiation (b). Only measurements from stations for which cloud observations are available are shown (A1, U2, A4, A5, I6, R4 and R5). For calculation of the incoming longwave radiation data from the radio soundings are used, which are available only twice a day. – *Samband mældrar og reiknaðrar sólgeislunar (a) og jarðgeislunar (b).*

of the images. It displays few clouds and much bare ice and shows that Breiðamerkurjökull and Skeiðarárjökull in the south have broad bands of low albedo and are quite inhomogeneous. Tungnaárjökull and Köldukvíslarjökull in the west are more homogeneous. The same is true for Dyngjujökull in the north but this outlet displays lower albedo values than the other ones (below 0.05 in places). For each model evaluation site we use the mean ice albedo that was determined for that site from the satellite images.

In November 1996 a volcanic eruption took place underneath Vatnajökull, which had a significant impact upon the ice cap (Gudmundsson *et al.*, 1997). The eruption melted the ice, causing a large flood or jökulhlaup south of Skeiðarárjökull. The eruption also covered large parts of Vatnajökull with tephra. From the AVHRR images we find that during the summer of 1997, the average albedo in the accumulation area was about 0.08 lower than in the years prior to and after 1997. In the model we therefore lower the albedo of snow and firn by 0.08 during the melting season of 1997. The eruption had no measurable effect upon the albedo of the ablation areas presumably because there the albedo already was relatively low.

Longwave radiation

I_{in} stems from the lowest part of the atmospheric boundary layer and from the upper-hemisphere slopes that surround a measurement site. For an exact calculation one needs to know the atmospheric profiles of temperature and humidity, but generally the influence of the vertical profiles can be well described with the temperature and water vapor pressure at screen-height, T_a and e_a. We use a slightly altered version of the parameterization given by Greuell *et al.* (1997) which is based on Konzelmann *et al.* (1994) and which calculates I_{in} as a function of T_a, e_a and cloudiness. However, the katabatic layer over Vatnajökull is generally not very thick (Oerlemans *et al.*, 1999) and T_a and e_a are not representative of that part of the atmosphere that generates I_{in}. Near A5, for ex-

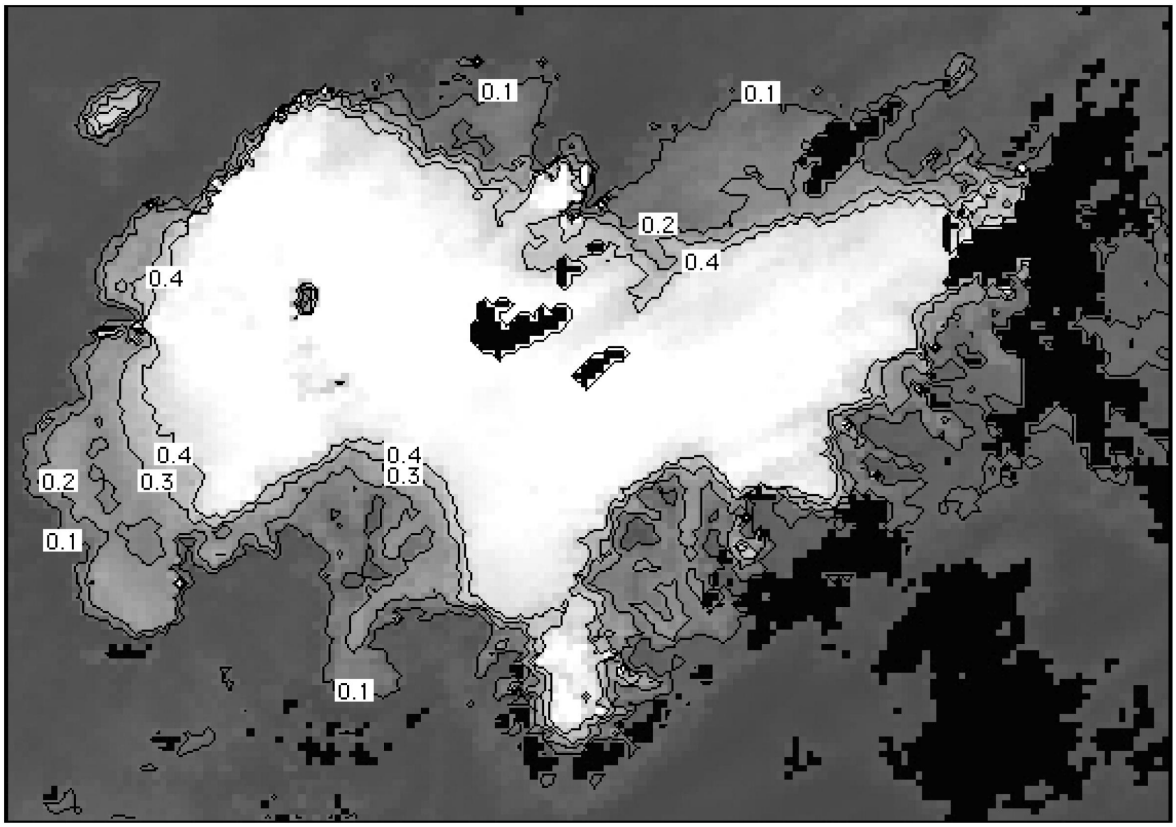

Figure 3. NOAA AVHRR image of Vatnajökull, taken on August 23rd, 1998. It has been processed so that it represents the surface albedo. The black pixels represent clouds. – *Gervitunglamynd sem sýnir endurkaststuðul sólgeislunar á Vatnajökli 23. ágúst 1998. Svörtu reitirnir stafa af skýjum.*

ample, the temperature inversion typically occurs in the lowest 20 to 30 m above the surface. This means that I_{in}, even for this station which is situated in the lower part of the ablation area, is only partly influenced by the katabatic layer and is best described as a function of meteorological variables in the free atmosphere (Meesters and van den Broeke, 1996). When we tune the parameterization to the measurements of I_{in} (Appendix), we indeed find that I_{in} is best described as a function of the temperature and the water vapor pressure in the free atmosphere at the same altitude as the measurement site (T_{atm} and e_{atm}). Apart from this, when the mass balance model is forced with external data, T_a and e_a need to be computed from the external data, which introduces extra uncertainty. When we compute I_{in} from the cloud observations and from T_{atm} and e_{atm} (measured by the radio soundings), we obtain a correlation coefficient between computed and observed I_{in} of 0.86. The residual standard deviation is 17.6 W/m^2 (Figure 2b).

I_{out} depends on the surface temperature and on the surface emissivity, which were not measured. However, the emissivity of snow and glacier ice is very close to 1 (Warren, 1982), while we assume the surface temperature to be equal to T_a when T_a is below the freezing point. Otherwise, the surface temperature is set to 0°C.

Turbulent fluxes

H_s and H_l are calculated with the bulk transfer method (e.g., Munro, 1990). This method requires values of windspeed, temperature and humidity at the surface and at some height above the surface (usually 2 m) as input. The basis for the bulk method, Monin-Obukhov similarity theory, is not strictly valid when a low level wind maximum is present, as is the case over sloping and melting glacier surfaces (Munro and Davies, 1978). In spite of this, recent work (Denby and Greuell, 2000) has shown that the bulk method only slightly overestimates H_s and H_l. The roughness length for momentum (z_0) has been reported to vary considerably in space and time over the ice surface of Breiðamerkurjökull, where values between 3 mm and 6 cm were found (Smeets *et al.*, 1999). The large values were caused by ice hummocks up to almost 2 m in height, which developed during the melting season. The smallest values were measured before these hummocks developed and these values are comparable with those for smooth ice surfaces found in the literature (Morris, 1989). Unfortunately, the aforementioned kind of irregularities can hardly be modeled and, moreover, do not arise in all ablation areas of Vatnajökull. We therefore choose an intermediate value of 5 mm for z_0 over ice. Denby and Greuell (2000) remarked that the error in the calculated turbulent heat fluxes due to an order of magnitude error in z_0 will be roughly 25%. We expect this to be an upper limit for the error present in the calculated turbulent heat fluxes. For snow surfaces, where these problems literally and figuratively do not arise, we use a value of 0.1 mm for dry snow and 2 mm for wet snow. These values are often found for snow surfaces (Morris, 1989). The roughness lengths for heat and moisture are calculated from z_0 with the often-used expressions of Andreas (1987).

Insulation

For most weather stations, the sum of observed net radiation and turbulent fluxes (as computed from observed 2 m variables) matches the energy that is required for the melt observed during the experiment (Figure 4). At sites where tephra covered ice appeared at the surface (I6, U8, U9 and R2) the simulated amount of melt is too high. This is probably caused by insulation of the underlying ice (e.g., Bozhinsky *et al.*, 1986; Kirkbride, 1995), because in the thermal channels of the NOAA satellite these parts of Vatnajökull appear slightly warmer than the rest of the ice cap surface. For the stations I6, U9 and R2 the observed melt corresponds to about 80% of the melt energy that was available when the surface was snow-free. For station U8 there was tephra-covered ice at the surface only during a few days of the 1996 experiment and the effect of insulation is small here. A reduction of melt by 20% is plausible, for this corresponds to a tephra layer of a few cm (e.g., Kirkbride, 1995). We do not attempt to model this effect elaborately, because it concerns only a minor part of the ablation area. Only on Dyngjujökull a significant part of the ablation area is covered by tephra or other debris (Figure 3). Furthermore, information needed to do so (notably thickness and thermal conductivity of the layer) is not available. We simply reduce the melt by 20% when the albedo is 0.15 or lower.

RECONSTRUCTION OF THE MASS BALANCE

We compute the mass balance (B) as the annual sum of ablation and solid precipitation:

$$B = \int_{year} \left[(0; -F/L_m) + P + H_l/L_s \right] dt \qquad (3)$$

where F/L_m is the ablation rate, L_m the latent heat of melting of ice, P the solid precipitation rate and L_s the latent heat of sublimation. The time step that we use is 30 minutes and we let the balance year start on September 21st. We neglect refreezing of meltwater and assume that the meltwater drains away instantaneously. Superimposed ice was not found anywhere on the ice cap during field work in the years 1992–1995 (Björnsson *et al.*, 1998a). Furthermore, the refreezing process is only important when the winter snowpack is cold (e.g., Colbeck, 1975) and the ablation season is short (Greuell and Oerlemans, 1986). Winters in Iceland are relatively mild and the ablation season on Vatnajökull is long. At station U7 (1530 m

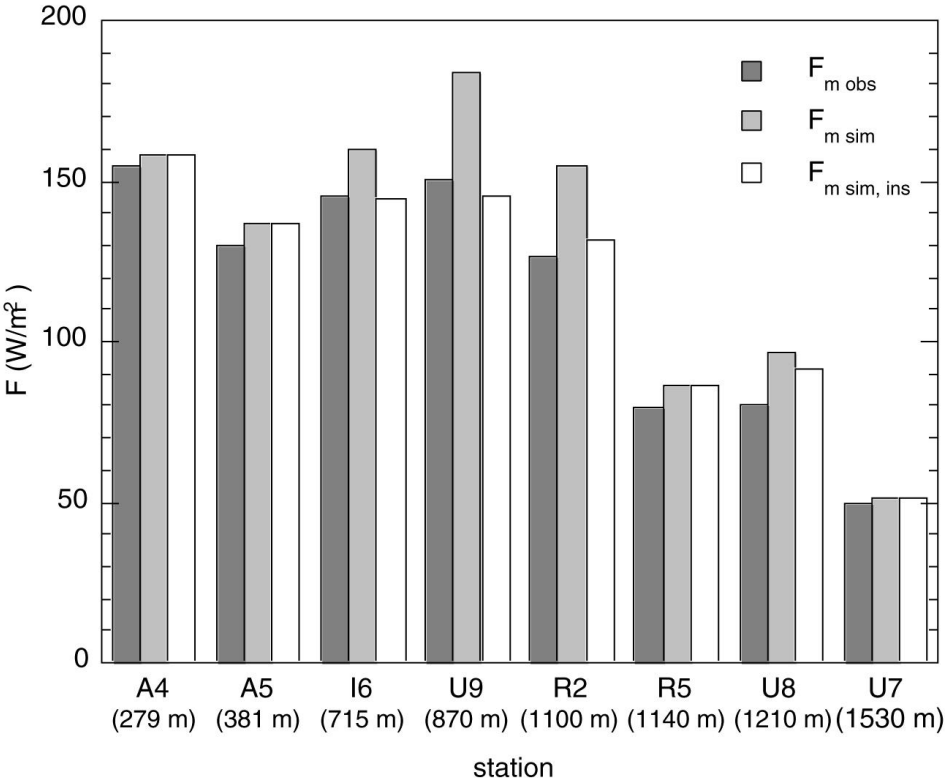

Figure 4. Mean energy used for melting (F_m) during the 1996 experiment. Only those stations are shown where the net radiation was measured during the entire experiment. Shown are the values obtained from the observed melt (obs), the simulated values without the effect of insulation (sim) and the simulated values with this effect (sim, ins). The simulated values are obtained from the observed net radiation and the turbulent fluxes calculated from the observed 2 m values of meteorological variables. – *Orka sem fór til leysingar á ýmsum mælistöðum sumarið 1996 (mælt og reiknað).*

a.s.l.) the ablation season lasts about 120 days and at station U3 (169 m a.s.l.) about 220 days.

Both the ablation rate and the accumulation rate of an ice body are largely determined by the conditions of the atmosphere that surrounds it. We assume all meteorological variables in the free atmosphere, except precipitation, to be horizontally homogenous around Vatnajökull. This seems justified, for the mean temperatures of three stations that lie at very different locations are nearly the same when they are extrapolated to sea-level: 9.3, 9.4 and 9.3°C for A1, U10 and R1, respectively. The tuning of the emissivity of clear skies (Appendix) also indicates that the free atmosphere was horizontally homogeneous, for all stations display the same results. Because of this assumption the model can be forced with data from a single permanent weather station. We use a permanent weather station outside of the ice cap so that we can reconstruct a long mass balance record and study the sensitivity of Vatnajökull to external climatic changes. There are three such stations close to Vatnajökull: one in Hornafjörður, one in Fagurhólsmýri and one in Kirkjubæjarklaustur (Figure 1). Hornafjörður is much more influenced by clouds and damp air from the ocean than most of Vatnajökull. The station in Fagurhólsmýri has no long records of humidity, so the station in Kirkjubæjarklaustur (named KIR hereafter)

Table 1. Observed and simulated sensitivities of temperature and windspeed at screen-height. The sensitivities were obtained by fitting a straight line to daily mean values. – *Mæld (obs.) og reiknuð (cal.) aukning í lofthita og vindhraða í veðurstöðvum við einnar gráðu hlýnun andrúmslofts.*

Station	Altitude (m)	dT_a/dT_{atm}		du_a/dT_{atm} $(ms^{-1}K^{-1})$
		Obs.	Cal.	Obs.
U3	165	0.32±0.05	0.34±0.00	0.30±0.08
A4	279	0.38±0.05	0.38±0.01	0.18±0.10
A5	381	0.41±0.05	0.43±0.01	0.12±0.08
I6	715	0.62±0.05	0.57±0.01	-0.04±0.09
R6	820	0.94±0.06	0.62±0.01	0.17±0.14
R4	830	0.70±0.07	0.62±0.01	0.21±0.11
U9	870	0.74±0.08	0.63±0.01	0.32±0.09
R2	1100	0.62±0.07	0.73±0.02	-0.20±0.13
R5	1140	0.75±0.06	0.75±0.02	0.05±0.12
U8	1210	0.78±0.08	0.78±0.02	0.19±0.12
U7	1530	0.99±0.08	0.91±0.02	-0.12±0.11
R3	1712	0.95±0.09	0.99±0.03	-0.02±0.15

is the most suitable. From this station we use daily means of temperature, vapor pressure, pressure and cloudiness from 1965 to 1999. Upon the temperature we impose a daily cycle with an amplitude of 2.2 K, which is the observed mean daily amplitude at the stations A1, U10 and R1. Precipitation in KIR was found to be not particularly well correlated to the observed winter mass balance of northern and western Vatnajökull. The precipitation measured in Fagurhólsmýri proved to be more useful, which is why we force the model with monthly values from this station.

Variables in the free atmosphere

The temperature in the free atmosphere influences I_{in} (see the Appendix) and the occurrence of solid precipitation. It is calculated from the temperature at KIR (T_0) and the atmospheric lapse rate (γ_{atm}), which is obtained from the radiosonde data (-5.8 K/km). The atmospheric pressure (p) is extrapolated from the pressure measured at KIR (p_0) with the same exponential decrease with altitude as was measured by the radiosondes. Vapor pressure in the free atmosphere (e_{atm}) is calculated by assuming a constant relative humidity with altitude. The radiosonde data showed that this was, on average, the case.

Variables in the katabatic layer

When melting conditions prevail a cool katabatic surface layer develops in which temperature and wind speed deviate from their values in the free atmosphere. All weather stations on the ice cap, except U7, R3 and R6, display a preferred wind direction (Oerlemans *et al.*, 1999). This means that a large part of the ice cap is subject to a more or less persistent katabatic wind regime which shields the surface from fluctuations in the atmosphere. This is clearly reflected in the sensitivities of 2 m variables to external changes (Table 1). The sensitivity of T_a to external temperature changes, dT_a/dT_{atm}, is smallest for the lowest stations and increases with altitude. It approaches 1 for the stations where a katabatic layer was weak or absent (U7 and R3 high in the accumulation area and R6 on an open spot at the very glacier margin). Neglecting the small sensitivities of the 2 m variables leads to an over-estimation of the sensitivity of the turbulent fluxes to external temperature changes (Greuell and Böhm, 1998).

From a physical point of view, parameterizations of the 2 m variables should be based upon the dynamic equations that describe the turbulent exchange of mo-

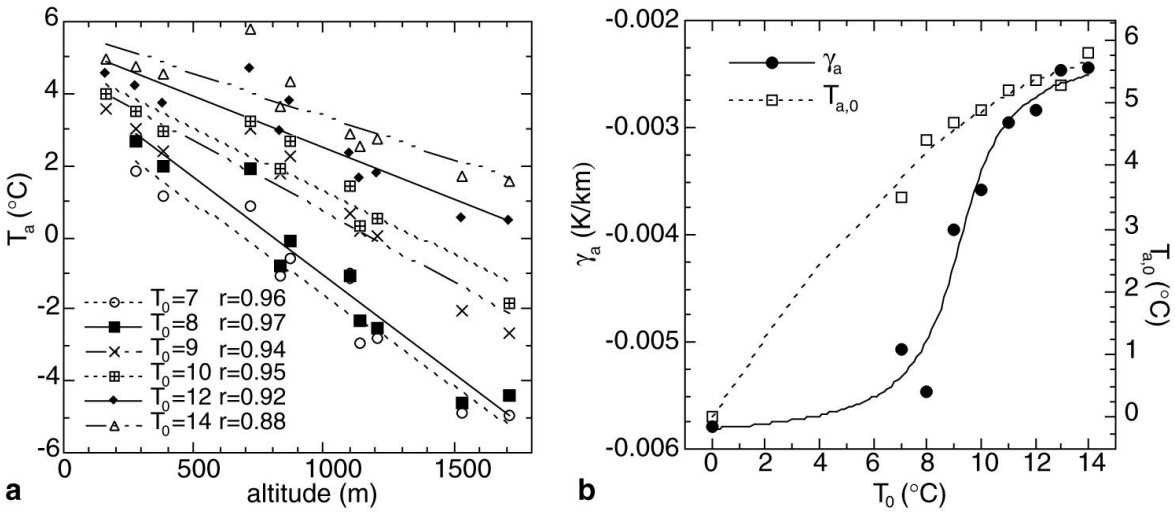

Figure 5. a: Average 2 m temperature (T_a) as a function of surface altitude, for different values of the temperature in Kirkjubæjarklaustur (T_0); b: 2 m lapse rate (γ_a) and 2 m temperature deviation at sea-level ($T_{a,0}$) as a function of T_0, both determined from the linear regressions in plot a. The data points for T_0 = 0°C are boundary conditions. – *a: Tengsl hita á Kirkjubæjarklaustri (T_0) og lofthita í 2 m hæð á Vatnajökli (T_a). b: hitastigull (γ_a) og hitafrávik ($T_{a,0}$) við sjávarmál sem fall af hitastigi á Kirkjubæjarklaustri (T_o).*

mentum, heat and moisture. However, the dynamics of the katabatic layer are very complex and external effects such as advection further complicate the situation. On the other hand, when we analyzed the data we found that T_a can be adequately described by empirical relations that result in correct sensitivities. T_a decreases linearly with altitude (Oerlemans *et al.*, 1999) and we found that this decrease changes with T_0 (Figure 5a). We therefore write

$$T_a = T_{a,0} + \gamma_a z \tag{4}$$

where $T_{a,0}$ is T_a at sea-level, γ_a is the lapse rate of T_a and z is the altitude. The dependence of both coefficients in equation 4 upon T_0 is shown in Figure 5b. For low temperatures hardly any katabatic flow exists and γ_a nearly equals γ_{atm}. For higher temperatures, γ_a grows less negative and $T_{a,0}$ deviates more from T_0. For high values of T_0, γ_a approaches -2.5 K/km. γ_a is best described by an inverse tangent (Figure 5b):

$$\gamma_a = -0.0041 + 0.0012 \arctan\left[0.77(T_0 - 9.2)\right] \tag{5}$$

When we impose the boundary condition that $T_{a,0}$ equals T_0 for T_0 = 0°C, we can describe the dependence of $T_{a,0}$ upon T_0 with (Figure 5b):

$$T_{a,0} = 0.68T_0 - 0.020{T_0}^2 \tag{6}$$

So the higher T_0, the less negative γ_a and the more negative the deviation of T_a from T_{atm}. These results are not surprising because the katabatic layer, which reduces T_a and γ_a, is better developed at higher temperatures. Table 2 shows that mean observed and simulated values of T_a correspond reasonably well. The parameterization of T_a produces mean errors between 0 and 1.2 K, but this error is much smaller than the mean difference between T_{atm} and T_a. The sensitivities that result from the parameterization correspond well with the observed sensitivities (Table 1).

For the 2 m windspeed (u_a) the picture is different. For some stations where the katabatic forcing is strong (U3, A4, U8, U9 and R4), du_a/dT_{atm} is 0.2 to 0.3. However, the uncertainties are large for these stations and even larger for the other stations, which display no significant sensitivities. We therefore do not develop a parameterization for u_a. Instead we use the mean wind speed in the free atmosphere, as measured by the radiosondes. For U3, R2 and R4 this results in a mean wind speed that is much too low, while for I6 it is much too high (Table 2). The former three stations lie at the very margin of the ice cap and experience strong katabatic winds. They represent, however, only a very small part of Vatnajökull. I6 is situated near nunataks that probably provide shelter from the wind.

Table 2. Differences between mean observed and mean simulated daily mean values of 2 m temperature (dT_a) and windspeed (du_a) during the 1996 experiment. – *Meðaltalsmismunur mælds og reiknaðs dagsmeðaltals lofthita og vindhraða í 2 m hæð sumarið 1996.*

Station	Altitude (m)	dT_a (K)	du_a (m/s)
U3	165	0.12	-1.57
A4	279	0.23	-0.44
A5	381	0.38	0.52
I6	715	1.17	1.15
R6	820	1.19	0.04
R4	830	0.27	-1.15
U9	870	1.13	-0.38
R2	1100	0.60	-1.30
R5	1140	0.30	-0.18
U8	1210	0.10	-0.29
U7	1530	0.46	0.54
R3	1715	0.02	0.36

Cloudiness

Cloudiness was observed every three hours close to U2 in the south and less regularly near R5 in the north. Mean values of cloudiness for these two stations and for the stations in Hornafjörður and KIR are shown in Table 3. More information about the mean cloudiness can be gained by assuming the cloud observations of station U2 to be valid for all stations and then comparing mean calculated with mean observed global radiation. The correspondence is good for most stations (Figure 6).

Table 3. Mean observed cloudiness (<n>) for several locations and periods. – *Meðaltal skýjahulu á ýmsum athugunarstöðvum.*

Station	Altitude (m)	<n> summer 1996	<n> summer 1966–1996
U2	50	0.750	
R5	1140	0.694	
Hornafj.	0	0.803	0.808
KIR	35	0.764	0.764

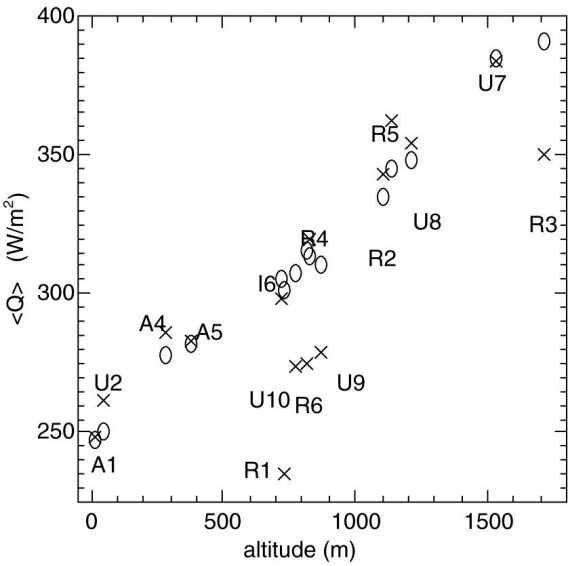

Figure 6. Mean observed (circles) and simulated (crosses) values of the global radiation over Vatnajökull as a function of altitude. In order to minimize errors in the observations, measurements corresponding to high solar zenith angles (>80°) were not used for this comparison. – *Mæld (hringir) og reiknuð (krossar) sólgeislun í mismunandi hæð á Vatnajökli.*

Some stations have large deviations, which are most likely caused by an erroneous cloud factor. Other factors in equation 2 cannot produce such large deviations. For U9, U10, R1, R3 and R6 the cal-

culated values are too high, so for these stations the mean cloudiness and/or the optical thickness must have been higher than at U2. R3 was located near the Grímsvötn depression and was often surrounded by fog. R6 was located near Hornafjörður, where the mean cloudiness was higher than at U2 (Table 3). For U9, U10 and R1 no other indications of a low cloud factor are available. Note, however, that U10 and R1 were not located on but close to the ice cap.

We force the model with daily values of cloudiness, and it seems best to use values measured in KIR. This is expected to introduce errors over the lowermost ablation zones in the north, the southeast (not over Breiðamerkurjökull) and perhaps in the west.

Precipitation and tuning of the model

Because the albedo depends upon the number of days since the last snowfall, it is important to simulate the time between subsequent snowfalls correctly. When we assume that precipitation falls every 5th day, and define days with snowfall as days with precipitation and a daily mean temperature below 3°C (in the free atmosphere above the katabatic layer), the mean albedo in the accumulation area is simulated correctly. The exact amount of annual precipitation is only known at a few locations near the coast and we will have to use it for final tuning. We assume that the precipitation gradients are fixed in time, and for all sites where B has been measured regularly, we calibrate P so that the modeled annual mean B is equal to the observed annual mean B (over the years 1993 to 1999). The resulting precipitation field must be interpolated to the model evaluation sites, but the measurement sites are irregularly distributed over the ice cap and do not resolve many of the altitudinal gradients. Hence, we make a distinction between horizontal and vertical gradients. Precipitation is often found to double in a certain altitude interval so we write:

$$P = P_0 k^{z/1000} \tag{7}$$

where P_0 is the mean annual precipitation at a reference altitude and k is a constant (e.g., for a doubling per 1000 m, k has a value of 2). We fit this equation to the winter mass balance data, which are available for the years 1993 to 1999 (Björnsson *et al.*, 1997, 1998a,b,c, 1999; Gudmundsson, 2000; Sigurdsson, 1997). These data were mainly obtained over the northwestern part of Vatnajökull, including most of the accumulation area. For Öræfajökull (the southernmost dome of Vatnajökull) we also use precipitation data from the permanent weather station in Fagurhólsmýri. When we correct the data for the small amounts of ablation and rainfall that occur during the winter season (both corrections are calculated with the mass balance model), we find a value of 2.3 for k. At each measurement site, we can now reduce P to its value at sea-level. The resulting values vary smoothly over the ice cap and we can interpolate them to the model evaluation sites with Kriging interpolation (e.g., Cressie, 1993). This interpolation technique is suitable for smooth interpolation of unevenly distributed data such as the mass balance observations that we use. At each model evaluation site we then calculate P with equation 7.

RESULTS

Energy balance

Observed and modeled components of the energy balance are displayed in Figure 7. All components change with altitude and this is simulated by the model. Q_{in} and α logically increase with altitude, whereas the net longwave radiation and the turbulent fluxes decrease with altitude. For station U9, and to a lesser extent for A4, A5 and I6, the modeled cloud factor is too low and hence the calculated global radiation is too high. Significant differences in the percentage of reflected shortwave radiation are present for A4, A5, and R2. For A4 and A5 this is caused by the ice albedo which is not representative of the surrounding ice. The model uses a more representative value. In the model results the snow disappears too soon at station R2 so the modeled mean albedo is too low. This is a consequence of tuning the model precipitation with mass balance data from several years. Errors in the turbulent heat fluxes mainly result from errors in the calculated meteorological variables at screen-height. Note that for U9 and R2 part of the available energy is not used for melting due to insulation.

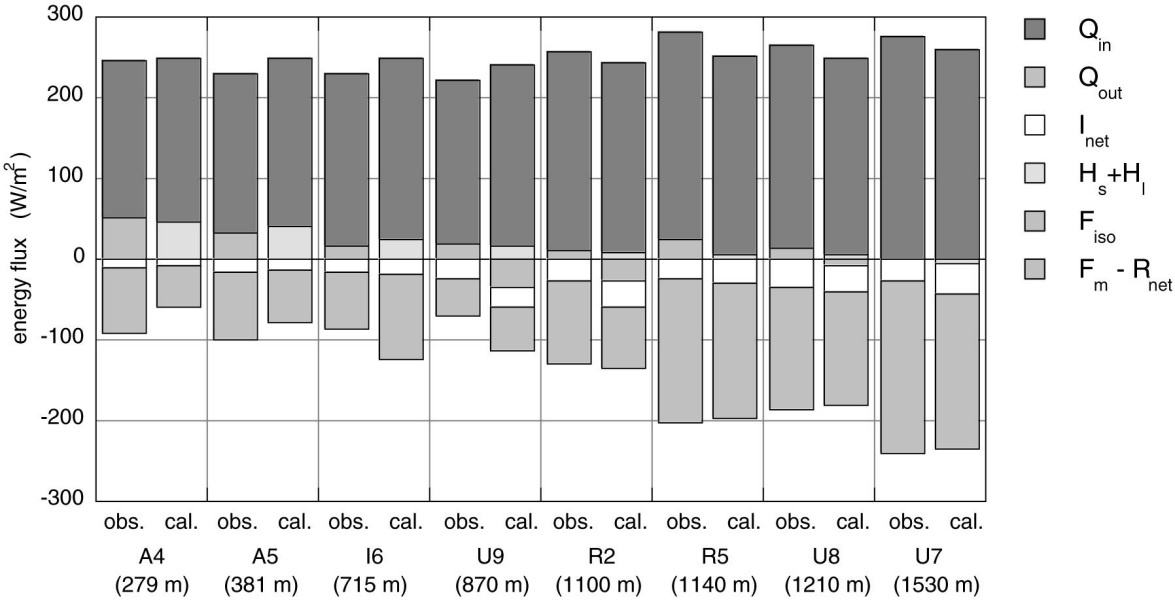

Figure 7. Mean observed (obs.) and calculated (cal.) components of the energy balance during the 1996 experiment. Those stations are shown where all radiation components have been measured. Q_{out} is the reflected shortwave radiation, I_{net} the net longwave radiation, F_{iso} the energy used to heat up the insulating layer, F_m the melt energy and R_{net} the net radiation. Other symbols are explained in the text. H_s, H_l and F_{iso} (if applicable) have not been measured directly, but the sum of these three modeled components should be equal to the observed difference between F_m and R_{net}. *– Mældir (obs.) og reiknaðir (cal.) orkuþættir sem bárust jökli í ýmsum sjálfvirkum veðurstöðvum. Q_in, sólgeislun sem fellur á jökul, Q_{out}, sólgeislun sem endurkastast frá jöklinum, I_{net}, heildarjarðgeislun, H_s, H_l, varmaþættir frá hlýju og röku lofti sem berst inn yfir jökulinn, F_{iso}, orka til upphitunar yfirborðslags, F_m, heildarorka til leysingar, R_{net}, heildargeislun.*

Mean specific mass balance

We compute the mean specific mass balance (B_m) with an interpolation scheme that has been especially designed for this purpose. With this interpolation scheme vertical gradients are taken into account, even when the evaluation sites do not resolve changes in altitude. For each grid point of the DEM we determine the n_s closest evaluation sites that do not differ more than 500 m in altitude from the grid point. Then, because of the limited height differences, a linear relation between mass balance and altitude can be found for the n_s evaluation points. With this relation the mass balance at the altitude of the gridpoint is calculated. In order to avoid discontinuities in the resulting mass balance field, the contribution of each evaluation point is weighted with the inverse of its distance to the grid point. The algorithm works best when for all parts of the ice cap, the entire altitudinal range is represented by the evaluation points. In order to fulfill this condition we use 128 evaluation points, and with n_s = 6 good results are obtained. Figure 8 shows the annual precipitation and B, interpolated over the whole ice cap and averaged over the period 1965–1999. Clearly most precipitation falls in the south and southeast, which is due to predominant southerly to easterly wind direction. Most precipitation falls on top of Öræfajökull, the highest part of Vatnajökull which lies in the south: almost 9 m w.e. annually.

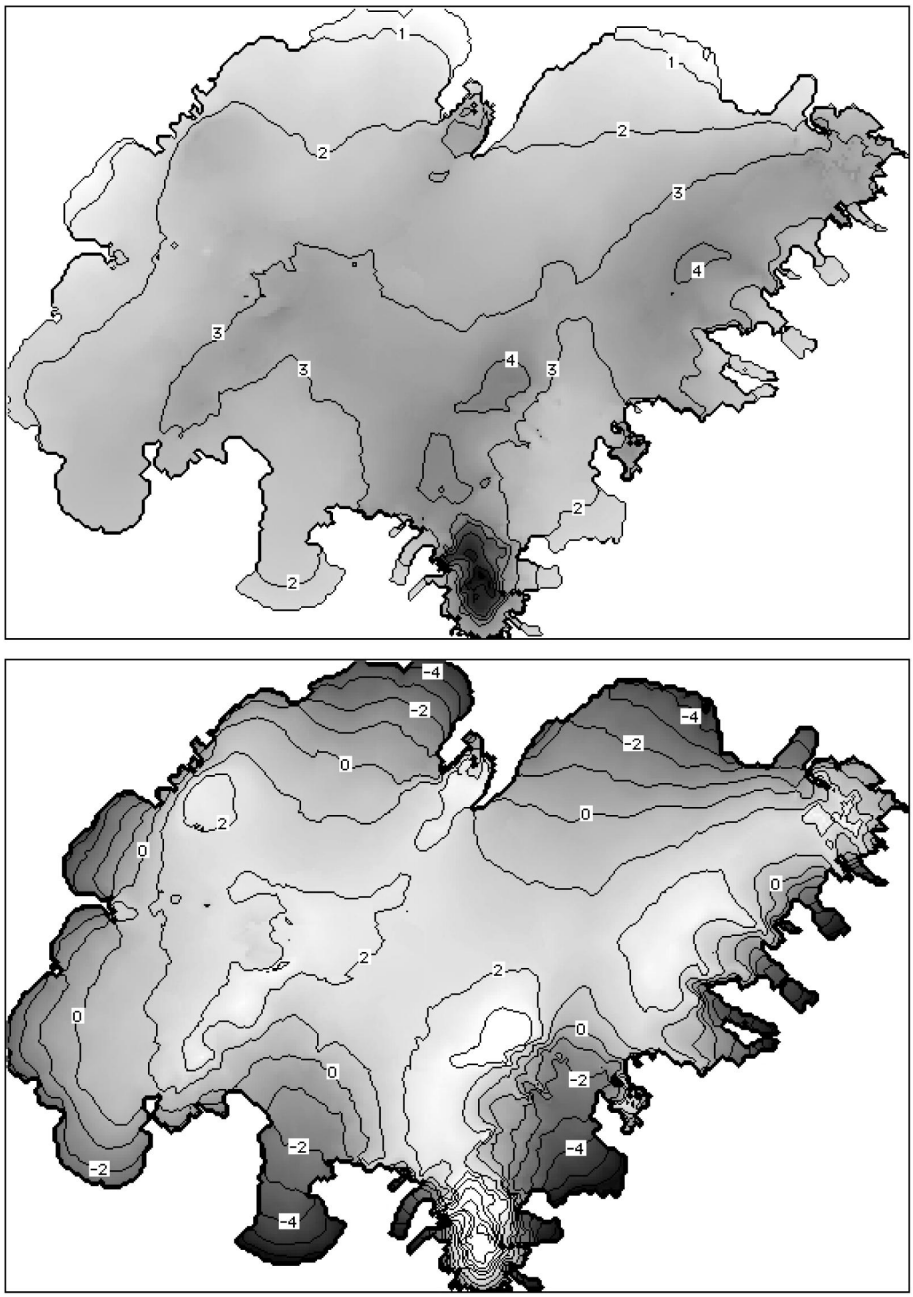

Figure 8. Modeled annual precipitation (top plot) and annual mass balance (bottom plot), averaged over the period 1965–1999. Both quantities are given in m w.e. – *Reiknað meðaltal ársúrkomu (efri mynd) og ársafkomu (neðri mynd) áranna 1965–1999 í m vatnsgildis.*

The outlet glaciers in the south and southeast also receive more precipitation than the ablation areas in the north and northwest, even though the latter are situated on higher grounds. Consequently, the Equilibrium Line Altitude (ELA) in the south and southeast is lower than in the northwest (Figure 8). The ELA ranges from about 1000 m at some locations in the south and southeast to over 1400 m on Dyngjujökull and Köldukvíslarjökull in the northwest. The accumulation area ratio of Vatnajökull is found to be 64%.

Figure 9 shows the simulated mean specific mass balance of the northwestern drainage basins of Vatnajökull as a function of time. The winter mass balance has slightly increased between 1965 and 2000, whereas the summer mass balance has decreased. As a result the annual B_m decreased from slightly positive in the 1970s to near-neutral in the 1990s (changes in glacier surface not taken into account). In general, fluctuations in winter and summer balance are more or less equally large. Interestingly, the most extreme positive and negative annual values both occur in the early 1990s and are both due to summer conditions. In 1991 the summer was exceptionally sunny and warm, whereas in 1992 the summer was cool with considerable snowfall in June and August (Björnsson *et al.*, 1998a). The model reproduces B_m over the years 1993 to 1999 fairly well (we estimate the uncertainty in the measured B_m to be 0.25 m w.e.) The effect of the albedo lowering due to the volcanic eruption of November 1996 is adequately simulated: without the additional lowering of the snow albedo during the summer of 1997, the simulated mass balance would have been considerably too high. Table 4 displays correlation coefficients between observed and modeled B_m for all drainage basins where B has been measured. In general the correlations are good, although for the winter mass balance the correlations are low for some drainage basins. This is obviously caused by the distribution of precipitation over Vatnajökull, which varies from year to year, whereas we use a fixed distribution of precipitation. Vatnajökull is sufficiently high and large to act as a topographic barrier, so the distribution of precipitation depends upon the large-scale atmospheric circulation.

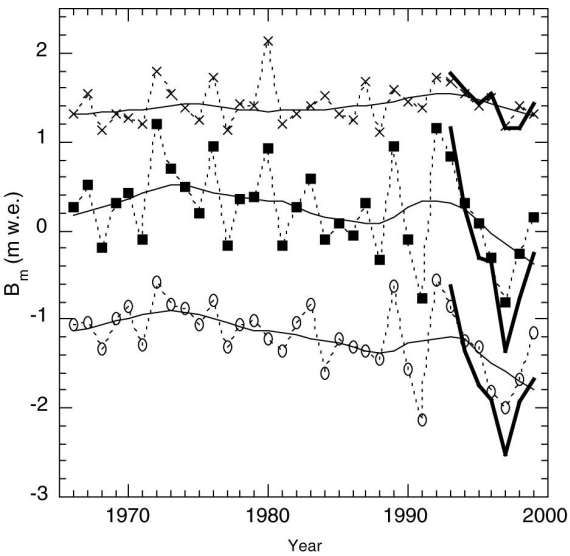

Figure 9. Observed (thick solid lines) and reconstructed (dashed lines) mean specific mass balance for the northwestern part of Vatnajökull (formed by the drainage basins Tungnaárjökull, Köldukvíslarjökull, Dyngjujökull and Brúarjökull). Shown are the winter balance (crosses), the summer balance (open circles) and the annual balance (solid squares). The thin solid lines are smoothed curves. – *Mæld (þykk lína) og reiknuð (brotin lína) afkoma á norðvesturhluta Vatnajökuls; vetrarafkoma (krossar), sumarafkoma (opnir hringir), ársafkoma (svartir ferningar); þunnar heildregnar línur sýna meðaltöl.*

THE SENSITIVITY OF VATNAJÖKULL TO CLIMATIC CHANGE

We compute the sensitivity of B_m to a change in a variable x as (Oerlemans, 1996):

$$C_x = \delta B_m / \delta x = \left[B_m(x+dx) - B_m(x-dx) \right] / 2dx \tag{8}$$

where dx is the change in variable x. We compute the sensitivity to changes in the two most important atmospheric variables, temperature (dT = 1K) and precipitation (dP = 10%). We perturb these variables over the period 1965–1999 and then compute the average change in B_m over this period. Table 5 shows C_T and C_P for different parts of Vatnajökull. Most obviously,

Table 4. Correlation coefficients between observed and modeled mean specific mass balance for the drainage basins where the mass balance has been measured. – *Fylgni mældrar og reiknaðrar afkomu á nokkrum vatnasvæðum.*

Region	Area (km^2)	Winter	Summer	Annual
Tungnaárjökull	491	0.86	0.81	0.93
Köldukvíslarjökull	330	0.65	0.89	0.98
Dyngjujökull	1239	0.77	0.92	0.96
western Brúarjökull	956	0.68	0.82	0.95
eastern Brúarjökull	780	0.65	0.90	0.97
Eyjabakkajökull	121	0.42	0.76	0.86

C_T varies strongly over Vatnajökull (third column). In regions with high precipitation in the south and east C_T is, depending on hypsometry, up to 70% higher than in the drier regions in the northwest. The same relation has been observed for a set of climatologically very different glaciers (Oerlemans and Fortuin, 1992) and our results fit well in this picture: C_T is almost -0.8 m w.e./K for glaciers with high precipitation and about -0.50 m w.e./K for drier glaciers. Thus, the precipitation gradients over Vatnajökull cause the sensitivity to temperature changes to vary considerably over the ice cap. Differences in precipitation also result in differences in C_P, but this sensitivity varies less over Vatnajökull.

Under the influence of increasing concentrations of greenhouse gases in the atmosphere, an annual temperature increase of about 0.3 K per decade is expected for Iceland in the next decades (Houghton *et al.*, 1996). For Vatnajökull an external temperature increase of 1 K implies a decrease of B_m with 0.71 m w.e. When we add an increase in precipitation of 5.3% for a warming of 1 K (Huybrechts *et al.*, 1991), the decrease is smaller but still 0.56 m w.e.

DISCUSSION AND CONCLUSIONS

We have presented a mass balance model for Vatnajökull which has been calibrated with an extensive data set. This made it possible to study differences between the free atmosphere and the katabatic layer. The calibration of the parameterization for I_{in} demonstrated that over Vatnajökull, I_{in} is better described as a function of T_{atm} and e_{atm} than as a function of T_a and e_a. This is a consequence of the moderate size of Vatnajökull, which allows advection of relatively warm air over the ice cap. This air strongly determines I_{in}, as the katabatic layer is not very thick. Nevertheless, the katabatic layer is well enough developed for T_a to deviate significantly from T_{atm}. The model presented in this paper uses the bulk method to compute the turbulent fluxes, which requires temperature, humidity and windspeed at the 2 m level. This means that it is important to make a distinction between these variables in the free atmosphere and in the katabatic surface layer. Most importantly, T_a differs from T_{atm}, such that the sensitivity dT_a/dT_{atm} is smaller than 1 (Table 1). If T_a were not explicitly calculated but simply set equal to T_{atm} (dT_a/dT_{atm}=1), using the bulk method would result in a value of C_T that is too high. The over-estimation of C_T in such a case is large: about 70%, which is much larger than Greuell and Böhm (1998) found for the Pasterze in the Austrian Alps (22%). Other mass balance models that are based upon a calculation of the energy balance often use a simplification of the bulk method (e.g. Oerlemans, 1992, Van de Wal and Oerlemans, 1994):

$$H_s = C(T - T_s) \tag{9}$$

where C is the transfer coefficient, T is the air temperature and T_s is the surface temperature. These models use the temperature in the free atmosphere (T_{atm}) for T in equation 9 and do not require T_a. For $C = H_s/(T_{atm} - T_s)$ we find values of 3.5–5

Table 5. Static sensitivities of the mean specific mass balance of various parts of Vatnajökull to changes in free atmospheric temperature and precipitation. All quantities are given as the mean during the period 1965 to 1999. – *Metin breyting í afkomu á ýmsum hlutum Vatnajökuls við einnar gráðu hlýnun (C_T) og 10% aukningu í úrkomu (C_P).*

Region	Area (km^2)	P_m (m)	C_T m $(w.e./K)$	C_P m (w.e./10%)
Síðujökull	514	2.7	-0.76	0.31
Tungnaárjökull	491	2.0	-0.77	0.30
Köldukvíslarjökull	330	1.7	-0.50	0.28
Dyngjujökull	1239	2.0	-0.49	0.28
W-Brúarjökull	956	2.3	-0.62	0.29
E-Brúarjökull	780	2.5	-0.73	0.30
Eyjabakkajökull	121	2.4	-0.74	0.27
Fláajökull	173	3.3	-0.73	0.32
E-Breiðamerkurjökull	631	3.0	-0.69	0.32
W-Breiðamerkurjökull	384	3.1	-0.80	0.35
E-Skeiðarárjökull	1002	3.1	-0.61	0.34
W-Skeiðarárjökull	434	2.8	-0.75	0.31
Vatnajökull	8198	2.6	-0.65	0.31

$Wm^{-2}K^{-1}$, which is somewhat lower than the values used by Oerlemans (1992) (7 $Wm^{-2}K^{-1}$). Other authors (e.g., Kuhn, 1989; Braithwaite, 1995) calculate C from 2 m variables and obtain values between 6 and 31 $Wm^{-2}K^{-1}$. These values compare well with the values that we find for $C = H_s/(T_a - T_s)$ and that are two to three times as high as $H_s/(T_{atm} - T_s)$ (i.e. 9–13 $Wm^{-2}K^{-1}$). So when T_a is explicitly calculated in the model, the bulk method can be used, but when only T_{atm} is available and if equation 9 is used, a correct transfer coefficient equal to $H_s/(T_{atm} - T_s)$ should be applied.

Another difference with some other models (e.g. Oerlemans, 1992) is the dependence of snow albedo upon the age of surface snow. For Vatnajökull the occurrence of summer snow fall in the accumulation area is strongly related to temperature. Modeling the albedo of snow as a function of snow age will increase C_T with respect to a model with constant snow albedo. When we use a constant mean value of 0.72 for the albedo of snow, we find that C_T is roughly 25% lower.

We studied the sensitivity of B_m of Vatnajökull to external climatic changes with the model. Meteorological stations in the vicinity of Vatnajökull are all located south of the ice cap, so we could only force the model with data from these weather stations. This introduces errors in cloudiness and precipitation, which are not well known over Vatnajökull. Especially the distribution of precipitation may vary from year to year, dependent on the predominant wind direction. In spite of this, the mean specific mass balance over seven years could be simulated reasonably well. The fact that Vatnajökull shapes its own precipitation gradients results in strongly varying sensitivities to temperature changes over the ice cap. In the south and southeast, where the climate is more maritime, C_T is higher than in the drier northwest. The variation of C_T over Vatnajökull stresses the importance of an optimization procedure as that described in this work. Even for much smaller ice caps or glaciers, the precipitation can vary considerably irrespective of altitude, and this strongly affects their sensitivity to climate change.

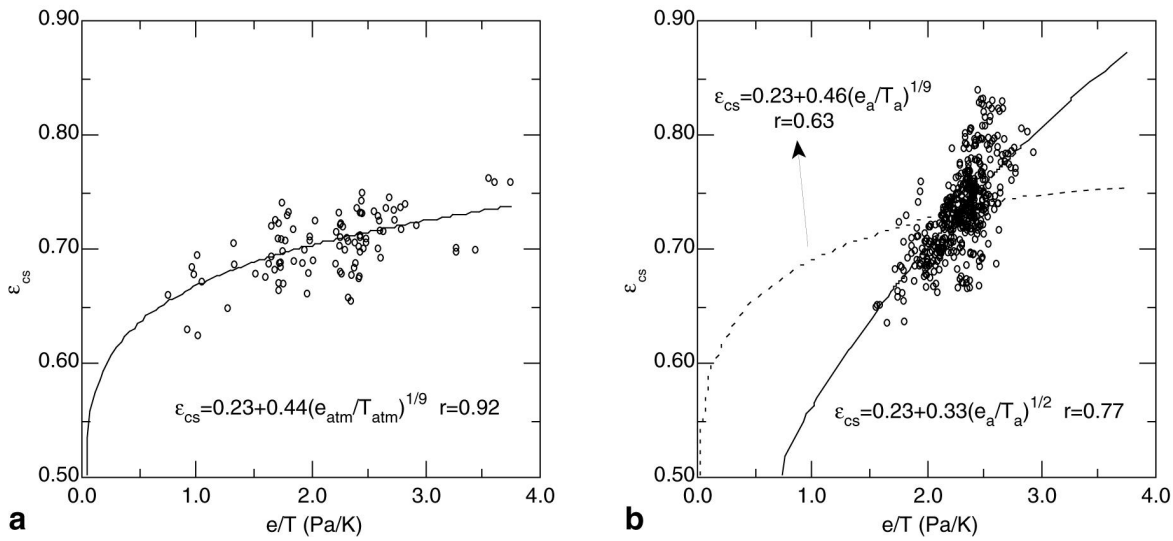

Figure 10. Clear-sky emissivity as a function of e/T. In plot a, e and T of the free atmosphere are used, which are available twice for each day. In plot b, hourly values of e and T at screen-height (2 m) are used. In both plots fits of equation 11 to the data are shown (solid lines), and in plot b a fit of equation 11 with a theoretical acceptable value of 9 for m is also shown (dotted line). – *Geislunarstuðull frá andrúmslofti við heiðan himinn sem fall af hlutfalli gufuþrýstings og lofthita (e/T).*

APPENDIX: Parameterization for the incoming longwave radiation

The incoming longwave radiation is given by:

$$I_{in} = \left[\varepsilon_{cs}(1-n^p) + \varepsilon_{oc}n^p\right] f_s \sigma T_{atm}^4 + L_{sl} \quad (10)$$

where ε_{cs} is the emissivity of a clear sky, n the cloudiness, p a constant (integer), ε_{oc} the emissivity of clouds, σ the Stefan-Boltzmann constant, L_{sl} the longwave radiation received from surrounding upper-hemisphere slopes and f_s a constant that is estimated from the DEM for each measurement site. The emissivity of a clear sky is written as

$$\varepsilon_{cs} = 0.23 + b(e_{atm}/T_{atm})^{1/m} \quad (11)$$

where b and m are constants. The integer m is theoretically expected to be not smaller than 7 (Konzelmann *et al.*, 1994). For all stations T_{atm} and e_{atm} are obtained from the radio soundings that were made twice a day near station U2. By doing so we assume that the atmospheric profiles above U2 are also valid for the rest of Vatnajökull. The parameterization contains four constants (b, m, ε_{cs} and p) that need to be tuned to the data. First, b and m are determined by fitting the parameterization to the clear-sky (n=0) measurements of I_{in}. We use the hourly mean values of I_{in} that are closest in time to the radio soundings. For all stations we find comparable values of b and m, with mean values of 0.438 and 9, respectively (Figure 10a). This shows that the assumption of a horizontally homogeneous atmosphere is justified. The residual standard deviation for ε_{cs} is 0.022. The value of b compares well with those found by Konzelmann *et al.* (1994), 0.443, and by Greuell *et al.* (1997), 0.475 and 0.407.

The value of m is, as theoretically expected, larger than 7. Konzelmann *et al.* (1994), who used T_a and e_a and not T_{atm} and e_{atm} in their calculations, found a value of 8 for a location near the equilibrium line of the Greenland ice sheet. There, T_a and e_a obvi-

ously describe the state of that part of the boundary-layer that generates I_{in}. The Greenland ice sheet is so large that advection of relatively warm air from ice free areas does not affect I_{in}. Greuell *et al.* (1997) also used T_a and e_a and found that a value of 8 for m is suitable for a site high in the accumulation area of the Pasterze (Austria). On the tongue of that glacier, however, it was not and they noted that this was presumably caused by the shallow katabatic layer over the tongue. We find the same for Vatnajökull when T_a and e_a are used in the equations 10 and 11 in stead of T_{atm} and e_{atm} (Figure 10b). We then find a value of 2 for m (with a residual standard deviation for ε_{cs} of 0.028), which is not acceptable from a theoretical point of view (Konzelmann *et al.*, 1994). However, in this case a more realistic value of 9 correspond less well to the data (residual standard deviation for ε_{cs} is 0.035). This means that the katabatic layer over Vatnajökull is generally not thick enough to influence I_{in} and that relatively warm air above the katabatic layer plays a larger role.

When an expression for ε_{cs} is available, ε_{oc} can be determined by fitting the parameterization to data that were measured when the sky was entirely overcast (n=1). The values that we find for the different stations lie between 0.937 and 0.958 and do not display any variation with altitude. We therefore use the mean value of 0.952 henceforth. This value is in line with those found by Konzelmann *et al.* (1994), 0.952, and by Greuell *et al.* (1997), 0.976. Next, p is determined from the data for which 0<n<1 (for n=0 and n=1, p has no influence). For all stations the same value is found, namely 3, except for R5 where it has a value of 4. For this station, however, only few data are available. We therefore use a value of 3 throughout.

Acknowledgements

We would like to thank all the people involved in the 1996 Vatnajökull experiment, and especially Friedl Obleitner, Finnur Pálsson, Paul Smeets and Jan de Wolde for providing information about the data. We also thank Trausti Jónsson from the Icelandic Meteorological Office for providing climatological data. Several anonymous reviewers provided useful comments. This work was funded by SRON (Space Research Organization of the Netherlands) under project number eo-030. The Vatnajökull experiment was supported by the European Union (Framework IV, contract EV5V-CT95-0105), The National Power Company of Iceland and NWO (Netherlands Foundation for Scientific Research) through GOA (Netherlands Geosciences Foundation).

ÁGRIP

Reiknilíkan af afkomu Vatnajökuls

Mæligögn um orkubúskap í sex sjálfvirkum veðurstöðvum og afkomu Vatnajökuls frá 1995–1996 eru notuð til þess að stilla af líkön sem lýsa afkomu hans með veðurþáttum frá veðurstöðvum utan jökulsins. Líkanið er síðan notað til þess að líkja eftir afkomu áranna 1965 til 1999 eftir mældu hitastigi og úrkomu á Kirkjubæjarklaustri. Líkanið bendir til þess að við hlýnun um 1°C, sem fylgdi 5,3% aukning í úrkomu, myndi afkoma Vatnajökuls rýrna um 0,56 m að vatnsgildi, jafndreift yfir jökulinn.

REFERENCES

Ambach, W. 1963. Untersuchungen zum Energieumsatz in der Ablationszone des grönlandischen Inlandeises (Camp IV-EGIG, 69°40'05" N, 49°37'58" W). *Meddelelser om Grønland* 174(4).

Ambach, W. and M. Kuhn 1985. Accumulation gradients in Greenland and mass balance response to climatic changes. *Z. Gletscherk. Glazialgeol.* 21, 311–317.

Andreas, E. 1987. A theory for the scalar roughness and the scalar transfer coefficients over snow and sea ice. *Boundary-Layer Meteorol.* 38, 159-184.

Arnold, N. S. I. C. Willis, M. J. Sharp, K. S. Richards and W. J. Lawson 1996. A distributed surface energy-balance model for a small valley glacier. I. Development and testing for Haut Glacier d'Arolla, Valais, Switzerland. *J. Glaciol.* 42(140), 77–89.

Björnsson, H., F. Pálsson, M. T. Guðmundsson and H. H. Haraldsson 1997. *Afkoma, Hreyfing og afrennsli á vestan- og norðanverðum Vatnajökli jökulárið 1995–1996.* Rep. RH-24-97, Science Institute, University of Iceland, 57 pp.

Björnsson, H., F. Pálsson, M. T. Guðmundsson and H. H. Haraldsson 1998a. Mass balance of western and northern Vatnajökull, Iceland, 1991–1995. *Jökull* 45, 35–58.

Björnsson, H., F. Pálsson, M. T. Guðmundsson and H. H. Haraldsson 1998b. *Mass balance and meteorological obserations on Vatnajökull 1997.* Rep. RH-03-98, Science Institute, University of Iceland.

Björnsson, H., F. Pálsson, M. T. Guðmundsson and H. H. Haraldsson 1998c. *Mass balance and meteorological observations on Vatnajökull 1998.* Rep. RH-14-98, Science Institute, University of Iceland.

Björnsson, H., F. Pálsson, M. T. Guðmundsson and H. H. Haraldsson 1999. *Mass balance and meteorological observations on Vatnajökull 1999.* Rep. RH-24-99, Science Institute, University of Iceland.

Bozhinskiy, A. N., M. S. Krass and V. V. Popovnin 1986. Role of debris cover in the thermal physics of glaciers. *J. Glaciol.* 32(111), 255–266.

Braithwaite, R. J. and O. B. Olesen 1990. A simple energy-balance model to calculate ice ablation at the margin of the Greenland ice sheet. *J. Glaciol.* 36(123), 222–288.

Braithwaite, R. J. 1995. Aerodynamic stability and turbulent sensible-heat flux over a melting ice surface, the Greenland ice sheet. *J. Glaciol.* 41(137), 153–160.

Colbeck, S. C. 1975. A theory for water flow through a layered snow pack. *Water Resour. Res.* 11(2), 261–266.

Cressie, N. A. 1993. *Statistics for Spatial Data.* Wiley, New York.

Denby, B. and W. Greuell 2000. The use of bulk and profile methods for determining the surface heat fluxes in the presence of glacier winds. *J. Glaciol.* 46(154), 445–452.

De Ruyter de Wildt, M. S., J. Oerlemans and H. Björnsson 2002. A method for monitoring glacier mass balance using satellite albedo measurements: application to Vatnajökull (Iceland). *J. Glaciol.* 48(161), 267–278.

Greuell, W. and R. Böhm 1998. 2 m temperatures along melting mid-latitude glaciers, and implications for the sensitivity of the mass balance to variations in temperature. *J. Glaciol.* 44(146), 9–20.

Greuell, W., W. H. Knap and P. C. Smeets 1997. Elevational changes in meteorological variables along a midlatitude glacier during summer. *J. Geophys. Res.* 102, 25,941–25,954.

Greuell, W. and J. Oerlemans 1986. Sensitivity studies with a mass balance model including temperature profile calculations inside the glacier. *Z. Gletscherk. Glazialgeol.* 22, 101–124.

Gudmundsson, M. T., F. Sigmundsson and H. Björnsson 1997. Ice-volcano interaction of the 1996 Gjálp subglacial eruption, Vatnajökull, Iceland. *Nature* 389 (30), 954–957.

Gudmundsson, M. T. 2000. Mass balance and precipitation on the summit plateau of Öræfajökull, SE-Iceland. *Jökull* 48, 49–54.

Houghton, J. T., L. G. Meiro Filho, B. A. Callander, N. Harris, A. Kattenberg and K. Maskell, *eds.* 1996. *Climate change 1995: the science of climate change.* Cambridge University Press, Cambridge.

Huybrechts, P., A. Letréguilly and N. Reeh 1991. The Greenland ice sheet and greenhouse warming. *Global Planet. Change* 3(4), 399–412.

Kirkbride, M. P. 1995. Processes of transportation. *In* J. Menzies, *ed. Modern glacial environments: processes, dynamics and sediments. Vol. 1. Glacial environments.* Butterworth-Heinemann, Oxford, 261–292.

Konzelmann, T., R. S. W. van de Wal, W. Greuell, R. Bintanja, E. A. C. Henneken and A. Abe-Ouchi 1994. Parameterisation of global and longwave incoming radiation for the Greenland Ice Sheet. *Global and Planetary Change* 9, 143–164.

Kuhn, M. 1989. On the computation of heat transfer coefficient from energy-balance gradients on a glacier. *J. Glaciol.* 22(87), 263–272.

Jóhannesson, T. 1997. The response of two Icelandic glaciers to climate warming computed with a degree-day glacier mass balance model coupled to a dynamic glacier model. *J. Glaciol.* 43(144), 321–327.

Meesters, A. and M. R. van den Broeke 1997. Response of the longwave radiation over melting snow and ice to atmospheric warming. *J. Glaciol.* 43(143), 66–70.

Meyers, T. P. and R. F. Dale 1983. Predicting daily insolation with hourly cloud height and coverage. *J. Clim. Appl. Meteorol.* 2, 343–348.

Morris, E. M. 1989. Turbulent transfer over snow and ice. *J. Hydrol.* 105, 205–223.

Munro, D. S. 1990. Comparison of melt energy computations and ablatometer measurements on melting ice and snow. *Arctic and Alpine Res.* 22(2), 153–162.

Munro, D. S. and J. A. Davies 1978. On fitting the log-linear model to wind speed and temperature profiles over a melting glacier. *Boundary-Layer Meteorol.* 15, 423–437.

Oerlemans, J. 1992. Climate sensitivity of glaciers in southern Norway: application of an energy-balance model to Nigardsbreen, Hellstugubreeen and Alfotbreen. *J. Glaciol.* 38(129), 223–232.

Oerlemans, J. 1996. Modelling the Response of Valley Glaciers to Climatic Change. *In* C. Boutron, *ed. ERCA – Physics and chemistry of the atmospheres of the Earth and other objects of the solar system.* Les Editions de Physique, Les Ulis, 91–123.

Oerlemans, J., H. Björnsson, M. Kuhn, F. Obleitner, F. Palsson, C. J. P. P. Smeets, H. F. Vugts and J. de Wolde 1999. Glacio-meteorological investigations on Vatnajökull, Iceland, summer 1996: an overview, *Boundary-Layer Meteorol.* 92(1), 3–26.

Oerlemans, J. and J. P. F. Fortuin 1992. Sensitivity of glaciers and small ice caps to greenhouse warming. *Science* 258, 115–117.

Oerlemans, J. and W. H. Knap 1998. A one-year record of short-wave radiation and albedo in the ablation zone of the Morteratschgletscher, Switzerland. *J. Glaciol.* 44, 147, 231–238.

Reijmer, C., W. H. Knap and J. Oerlemans 1999. The surface albedo of the Vatnajökull ice cap, Iceland: A comparison between satellite-derived and ground-based measurements. *Boundary-Layer Meteorol.* 92(1), 125–144.

Sigurdsson, O. 1997. Jöklabreytingar 1930–1960, 1960–1990 og 1995–1996. *Jökull* 47, 101–104.

Smeets, C. J. P. P., P. G. Duynkerke and H. F. Vugts 1999. Observed wind profiles and turbulence fluxes over an ice surface with changing surface roughness. *Boundary-Layer Meteorol.* 92(1), 101–123.

Van de Wal, R. S. W. and J. Oerlemans 1994. An energy balance model for the Greenland ice sheet. *Global Planet. Change* 9, 115–131.

Walraven, R. 1978. Calculating the position of the sun. *Solar Energy* 20, 393–397.

Warren, S. G. 1982. Optical properties of snow. *Rev. Geophys. Space Phys.* 20(2), 67–89.

Williams Jr., R. S., D. K. Hall, O. Sigurdsson and J. Y. L. Chien 1997. Comparison of satellite-derived with ground-based measurements of the fluctuations of the margins of Vatnajökull, Iceland, 1973–92. *Ann. Glaciol.* 24, 72–80.

One of the Vatnajökull weather stations. – *Hannes H. Haraldsson við sjálfvirka veðurstöð á Vatnajökli.* Photo./*Ljósm.* Finnur Pálsson.