MODELLING LONG-TERM SUMMER AND WINTER BALANCES AND THE CLIMATE SENSITIVITY OF STORBREEN, NORWAY

BY

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ABSTRACT. Measurements of winter balance (b_w) and summer balance (b_s) have been carried out at Storbreen since 1949. Here we apply a simple mass balance model to study the climate sensitivity and to reconstruct the mass balance series prior to 1949. The model is calibrated and validated with data from an automatic weather station (AWS) operating in the ablation zone of Storbreen since 2001. Regression analysis revealed that b_w was best modelled using precipitation data southwest of the glacier. Results from the model compared well with reported mass balance values for the period 1949–2006, obtained correlations (r) for b_w and b_s varied between 0.83 and 0.87 depending on model set up. Reconstruction of the mass balance series for the period 1924/1925–1948/1949 suggested a cumulative mass deficit of *c.* 30 m w.e. mainly due to highly negative summer balances, but also lower b_w than the average for 1949–2006. Calculated change in specific mass balance for a ±1°C change in air temperature was ±0.55 m w.e., whereas a ±10 % increase in precipitation represented a change of ± 0.20 m w.e. Model results further indicated that for a 2°C warming, the ablation season will be extended by *c.* 30 days and that the period of ice melt at the AWS location will increase from *c.* 40 to *c.* 80 days.

Key words: glacier, mass balance, albedo, temperature, precipitation, precipitation, climate

Introduction

To predict how a glacier will react to climate change or to reconstruct a glacier's mass balance back in time, a mass balance model is needed to relate the surface mass balance to prevailing climate conditions. A whole hierarchy of models exists, from simple regression models to complex surface energy balance models (Hock 2005). While precipitation is prescribed from precipitation measurements in most models, the ablation routines differ greatly. Various versions of temperature–index models, or degree–day models, are widely used as they only require either synthetic or measured temperature as input for modelling the melt (e.g. Laumann and Reeh 1993; Braithwaite 1995). Surface energy balance models compute each of the relevant energy fluxes between the atmosphere and the surface and require additional meteorological measurements and observations, for example cloudiness and humidity. For studies of climate sensitivity to free-atmosphere changes, data from stations outside the glacier boundary layer should be used (Greuell and Böhm 1998). The input data required for an energy balance model is comprehensive, and obtaining long and homogeneous time-series of the required meteorological data is difficult. Attempts have been made to use distributed energy-balance models over glaciers with input data outside the glacier (e.g. Klok and Oerlemans 2002). A limiting factor is nevertheless the lack of temporal and spatial input data to run energy balance models on time scales from years to decades and many of the parameters needed for the calculations must be estimated. If no meteorological stations are operated in the vicinity of the glacier, the observations of cloudiness, radiation and wind speed needed to drive an energy balance model may not reflect the conditions at the glacier. Hence, to model a long time-series several simplifications to the fully distributed energy-balance approach are necessary.

To judge the performance of a mass balance model, its results should be compared with observed data. A series of winter- and summer balances is preferable as it provides more information to test and validate a model. Mass balance data are available from many glaciers around the world, but few observations go back more than 50 years (Dyurgerov 2002; Haeberli *et al*. 2007). The longest time series of a glacial mass balance started in 1946 at Storglaciären in northern Sweden, followed by Storbreen in southern Norway which began in 1949. In this paper we use the long and continuous series of winter and summer balances from Storbreen to test a simplified mass balance model. The mass balance record of Storbreen correlates well with the mass balances of other glaciers in southern Norway and can be used to reconstruct other mass balance series back in time (Rasmussen *et al*. 2007). A previous study using the first 17 years of mass balance data showed that the glacier ablation was well correlated to summer temperature and positive degree days and the mass balance was reconstructed back to 1816 using temperature and precipitation records from Bergen (Liestøl 1967). In another model study, circulation indices and spring–summer temperatures were used to model the mass balance of Storbreen back to 1781 (Nordli *et al*. 2005). Meteorological observations carried out on Storbreen in the summer of 1955 (Liestøl 1967) revealed that net radiation is the most important contributor to the ablation at Storbreen. An **automatic weather station (AWS)** has been operated in the ablation zone of Storbreen since September 2001 providing a near-continuous series of meteorology and surface energy balance data. Analysis of the first five years of data revealed that variations in temperature and reflected shortwave radiation (albedo) explained most of the inter-annual variation in melt, whereas the seasonal mean incoming shortwave radiation was remarkably constant between the years (Andreassen *et al*. 2008). Here we have applied and tested a simplified mass balance model for Storbreen based on energy balance principles. Data from the AWS have been used to calibrate and validate the model. The model includes parameterisation of snow albedo and is driven by temperature and precipitation data from weather stations outside the glacier. We compare the modelled and measured values for the period 1949–2006 and reconstruct the mass balance back in time. The main objectives are to:

- establish a simple mass balance model that captures the variations in the seasonal components of the mass balance;
- use the model to reconstruct the mass balance prior to 1949;
- study the sensitivity of the model to different model set ups and choices of parameters;
- and study the climate sensitivity of the glacier and implications of climate change on the mass balance.

Setting

Storbreen (61°36' N, 8°8' E) is located in the Jotunheimen mountain massif in southern Norway (Fig. 1). The glacier has a total area of 5.4 km^2 and ranges in altitude from 1390 to 2090 m a.s.l. The glacier has been mapped repeatedly, the most recent map is from 1997. Areas calculated from the 1951 and 1997 maps reveal an area reduction of about 0.4 km^2 in this period (Andreassen 1999). Length change observations reveal a net retreat of about 60 m from 1997 to 2006 (data: NVE; e.g. Kjøllmoen *et al*. 2007).

Storbreen is located just east of the main water divide between east and west in southern Norway and receives precipitation from both directions (Liestøl 1967). The glacier is part of an east–west mass balance transect in southern Norway where mass turnover is largest near the western coast and decreases towards the drier interior (Andreassen *et al*. 2005). Storbreen is in this respect considered as a continental glacier, with a smaller mass balance turnover and less winter precipitation than glaciers situated farther west.

Data

Mass balance measurements

The mass balance measurements on Storbreen were started in spring 1949 by the Norwegian Polar Institute (Liestøl 1967). NVE (Norwegian Water Resources and Energy Directorate) took over the measurements in 1994. Although the principal methods have not changed much over the years, the amount of field work has varied. In the first 15 years the monitoring programme at Storbreen was comprehensive, often three or more snow density pits were dug, snow depth was measured at about 600 points and ablation was measured on 30 stakes evenly distributed on the glacier (Liestøl 1967). Based on experience of the snow pattern, the observations were gradually reduced in the 1960s (Østrem and Liestøl 1964). For a period in the 1970s and 1980s measurements were sparse and reduced to a minimum, for some years snow accumulation was only measured at 10–20 points. From the mid-1990s the mass balance programme was

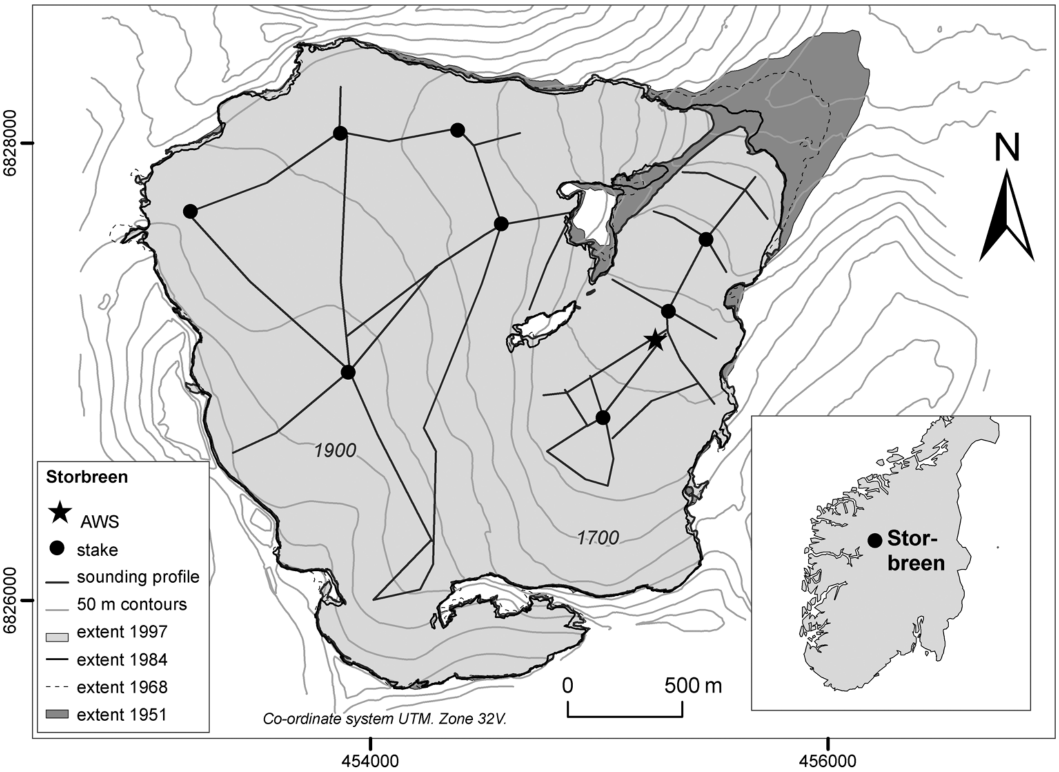

Fig. 1. Location map of Storbreen, southern Norway, showing stake positions and sounding lines for mass balance measurements (2007), position of the Automatic Weather Station (AWS) and glacier extent in 1951, 1968, 1984 and 1997. Glacier contours (50 m) are from 1997

extended. Since then mass balance has been measured at 6–10 stakes, snow depth at 100–150 points along sounding profiles, and snow density at one or two locations (Fig. 1). The mass balance at Storbreen is calculated using the stratigraphic method, i.e. between two successive 'summer surfaces' (surface minima). Consequently, the measurements describe the state of the glacier after the end of melting and before fresh snow starts to fall. The winter (b_w) and summer balance (b_s) are calculated separately and the resulting net balance (b_n) is calculated as the sum of the two components,

$$b_n = b_w + b_s \tag{1}$$

where b_s is negative. The mass balance profiles are made by plotting point measurements of winter and summer balance versus altitude, and their mean values for each 50 m elevation interval are determined. The curves are extrapolated to the lower and upper parts. Another approach, however, was used until the 1980s when hand-contoured maps of accumulation and ablation were made from the observations. The areas within each height interval (50 m) were planimetered and the total amount of accumulation and ablation was calculated for each height interval, and profiles $b_w(z)$, $b_s(z)$ and $b_n(z)$ were created. The mass balance results were previously reported in the series Norsk Polarinstitutt Skrifter and the level of details has varied. Only $b_n(z)$ profiles and specific values were reported in many of the years before 1987. For six years (1955, 1957–60 and 1967) only specific values (no profiles) were reported. In general, reports from Storbreen for the period 1949–1984 do not include the date of the measurements. From 1985 these dates are available. Typically, accumulation measurements are carried out in early May and ablation measurements in the middle of September.

The mass balance of Storbreen has been negative in most years and the glacier had a total mass loss of −16.9 m w.e. for the period 1949–2006 or −0.26 m

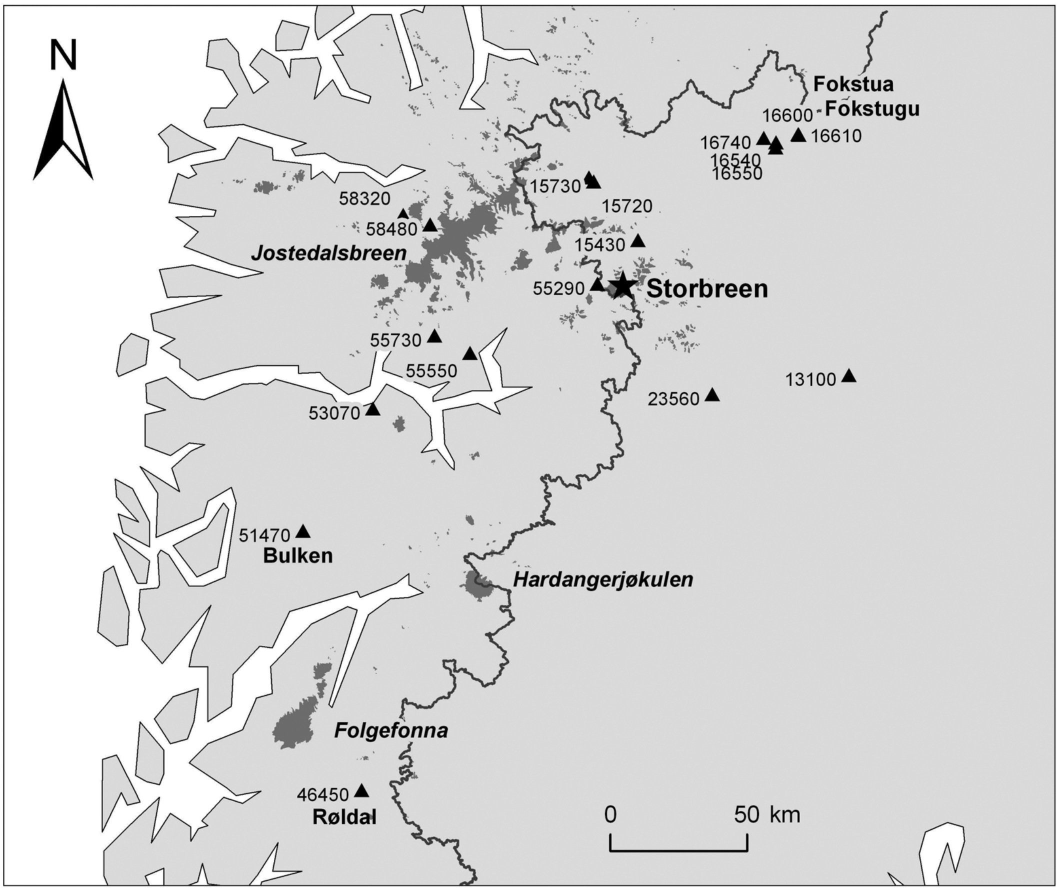

Fig. 2. Map showing location of meteorological stations (▲) in southern Norway used in this study. Glaciers are shaded in dark grey. The main water divide between east and west is shown as a thick grey line. Information about the stations is given in Tables 1 and 2

w.e. per year (data: NVE; Kjøllmoen *et al*. 2007). The mean **winter balance (b_w)** was 1.43 m w.e, the mean **summer balance (b_s)** was −1.72 m w.e. Storbreen had a transient mass surplus in the period 1989–1995 due to high b_w and decreased b_s. The increase in b_w is related to the highly positive **North Atlantic Oscillation (NAO)** index in this period (Pohjola and Rogers 1997; Hanssen-Bauer and Førland 1998; Rasmussen and Conway 2005) although correlation between the NAO-index and b_w is weaker at Storbreen (r=0.61) than at more maritime glaciers such as Nigardsbreen (r=0.71) and Ålfotbreen (r=0.78) located further west (Nesje *et al*. 2000).

Meteorological observations at Storbreen

An automatic weather station (AWS) has been operating in the ablation zone of Storbreen at *c*. 1570 m a.s.l. since September 2001 (Fig. 1). The station is part of the Institute of Marine and Atmospheric research Utrecht (IMAU) network of AWS on glaciers. The AWS measures air temperature, relative humidity, wind speed, wind direction, surface height, air pressure, and the incoming and outgoing shortwave and longwave radiation. The data from the station have been used to calculate the energy balance budget of the glacier surface for the nearly continuous period of data from 6 September 2001–10 September 2006 (Andreassen *et al*. 2008). The **surface energy balance (SEB)** was calculated using an energy balance model previously applied for Antarctica (Van den Broeke *et al*. 2005). The surface energy balance model (hereafter called the SEB-model) calculates the turbulent fluxes, outgoing longwave radiation, subsurface heat flux and melt energy by solving the SEB for the surface tem-

Table 1. Precipitation stations used to calculate correlation with b_w at Storbreen. The station information is taken from www.eklima.no. Mean annual (A) and winter (W) precipitation (P) are calculated for the period 1949–2006. Winter is here defined as the calendar months Jan–Apr and Oct–Dec. Correlation (r) and coefficient of determination (r^2) between winter precipitation sum (P_w) for the period 15 September to 14 May and b_w are also given. Figure 2 shows the location of the stations.

Station information		Digital data	Altitude	North	East		P (mm)			
No.	Name	From	(m a.s.l.)	(°)	(°)	Region	A	W	r^2	r
23560	Beito	09.03.1905	754	61.24	8.86	East	742	317	0.03	0.17
15730	Bråtå-Slettom	01.11.1998	664	61.90	7.90	East	–	–	–	–
15720	Bråtå[1,2]	01.09.1950[2]	712	61.91	7.86	East	537	328	0.60	0.77
15430	Bøverdalen	25.03.1905	701	61.72	8.24	East	440	224	0.49	0.70
16600	Fokstua[3]	01.03.1923[4]	952	62.12	9.28	East	–	–	–	–
16610	Fokstugu	01.06.1968	972	62.11	9.29	East	448	181	0.41	0.59
13100	V Gausdal	01.01.1896	580	61.34	9.77	East	589	229	0.00	–0.02
58480	Briksdal	01.07.1895	40	61.69	6.81	West	1351	849	0.57	0.75
51470	Bulken	09.03.1905	323	60.65	6.22	West	1808	1124	0.69	0.83
55550	Haflso	09.03.1905	246	61.29	7.19	West	1032	644	0.63	0.80
58320	Myklebust i B	01.07.1895	315	61.71	6.62	West	1612	952	0.65	0.81
46450	Røldal	25.02.1905	393	59.83	6.83	West	1606	1002	0.70	0.84
55730	Sogndal	01.07.1895	421	61.33	6.93	West	1525	941	0.68	0.83
53070	Vik i Sogn	01.01.1895	65	61.07	6.58	West	1086	708	0.68	0.82

Notes: [1]Period 1951–2006; [2]Not in operation from 01.06.1998 and replaced by 15730; [3]Period 1957–2006; [4]Not operational from 01.05.1968 and replaced by 16610 from 01.06.1968.

perature, the other fluxes are taken directly from the AWS measurements. A detailed description of the station, the data treatment and the surface energy balance calculations is found in Andreassen *et al.* (2008). In this paper we use the results from the Storbreen AWS and the SEB-model over the AWS-period 2001–2006 for calibration and validation purposes.

To study the spatial variability of the albedo of Storbreen a portable albedometer was used to map the surface albedo in 2006 and 2007. The portable albedometer was designed by IMAU, including a Kipp and Zonen CM3 Pyranometer measuring the incoming and outgoing radiation and a Garmin GPS 16x determining the position of each point measurement. Data were stored on a Campbell logger. At each location, the surface was photographed and surface type, surface condition and weather was noted. In total more than 100 point measurements were made during three days (2 and 3 August 2006 and 7 August 2007). Measurements were made on different surface types: ice, snow and firn, as well as on different surface conditions (dry, wet, clean, dirty). Some of the measurements will be referred to in this paper.

Meteorological observations outside Storbreen

The station network of the Norwegian Meteorological Institute (hereafter called met.no) comprises manual and automatic weather stations and precipitation stations (where only precipitation and snow depth are measured). Although the network is extensive, few stations are situated in high elevated areas. Moreover, stations are terminated or relocated from time to time and records contain data gaps (Engen-Skaugen *et al.* 2007). As an example, the closest weather station to Storbreen, Sognefjellshytta (1413 m a.s.l., located 8 km west of the glacier), has been operated since 1979 (as a manual station from 1979–1989, as an automatic station from 1997). Data are lacking for the intermediate period 1989–1997 and the records contain many gaps, making it difficult to use these data in long-term modelling. Moreover, the stations Liestøl (1967) used to model Storbreen are not longer operational (Elveseter, Fannaråken, Sikkilsdal and Luster) or inhomogeneous due to relocations (Bergen). To model the whole mass balance series for the period 1949–2006 and also reconstruct the mass balance back in time, we needed to select other stations with long-term, digitally available observations (Fig. 2, Table 1, and Table 2). Daily data of precipitation and temperature were downloaded from the eKlima portal (www.eklima.no). In general, meteorological observations series in Norway are digitally available from 1957 (as from this year on all observations were recorded digitally). Supplementary long-term data series of temperature from before 1957 were provided by Øyvind Nordli

Table 2. Correlation between daily temperatures observed at AWS at Storbreen for 2001–2006 and 4 weather stations (See Fig. 2 for location). The stations 15730 Bråtå, 16740 Kjøremsgrendi and 16610 Kjøremsgrendi are replacement stations after relocated stations. Station information on these stations is also included in the table.

Station information No.	Name	Digital data From	Altitude (m a.s.l.)	To	North (°)	East (°)	Region	r	r^2	No of values
15720	Bråtå-Slettom[1]	01.09.1950	01.06.1998	712	61.91	7.86	East	–	–	–
15730	Bråtå[2]	01.11.1998	–	664	61.90	7.90	East	0.90	0.80	1803
16540	Dombås- Kirkenær	01.07.1972	01.05.1976	645	62.08	9.13	East	–	–	–
16550	Dombås / Dombås II	01.01.1864	01.06.1972	643	62.07	9.13	East	–	–	–
16740	Kjøremsgrende[3]	01.07.1976	–	626	62.09	9.05	East	0.90	0.81	1818
16600	Fokstua[4]	01.03.1923	01.05.1968	952	62.12	9.28	East	–	–	–
16610	Fokstugu	01.06.1968	–	972	62.11	9.29	East	0.95	0.91	1818
55290	Sognefjellshytta[5]	01.12.1978	–	1413	61.57	8.00	East	0.97	0.95	1579

Notes:[1]Temperature data available from 1.1.1957
[2]Replacement for 15720
[3]No temperature data until 1.10.1977
[4]Not operational from 01.05.1968 and replaced by 16610 from 01.06.1968
[5]No data between 01.06.1989 and 01.07.1997. Several data gaps in the series.

(met.no). In addition, a few precipitation series were downloaded from NVE's database Hydra II (as many precipitation series had been made digitally available through data entering by NVE personnel).

Correlation between precipitation (P) and winter balance (b_w)

Long-term precipitation records are generally more accessible than the other meteorological elements. As Storbreen is located close to the main water divide between east and west in southern Norway, we analysed precipitation data for a selection of long-term precipitation series (Table 1, Fig. 2). Resulting mean precipitation sums for selected stations, calculated for the period 1949–2006, illustrate the strong gradient in precipitation between stations west and east of the watershed in southern Norway (Table 1). The station with highest annual precipitation is Bulken (1808 mm), the lowest is Bøverdalen (440 mm), which is the precipitation station closest to Storbreen (Table 1, Fig. 2). To compare measured winter balance and accumulated winter precipitation, we calculated **precipitation sums, P_w**, from 15 September to 14 May in the following year. The **correlation (r)** and coefficient of **determination (r^2)** between winter precipitation and b_w were obtained for each station for the period of mass balance measurements 1949–2006.

Data from the southwestern precipitation station Røldal showed the best correlation between P_w and measured b_w (r=0.84) (Table 1). Similar correlations were found for Bulken (r=0.83) and Sogndal (r=0.83). It is striking that the P_w at the most eastern stations, Beito and Vestre Gausdal, correlate so poorly with b_w. The correlation with Storbreen b_w was remarkably similar within the western region. All western stations, except Brigsdal, were better correlated with the b_w at Storbreen than the closer-by located stations Bråtå and Bøverdalen north of Storbreen, but east of the water divide. We also tested combinations of two or more stations, the best combination was of Røldal and Bulken (r=0.85, r^2=0.72), which we therefore chose to use as input to the mass balance model.

Correlation between air temperature at weather station and Storbreen

We also calculated correlations between daily mean air temperature at Storbreen AWS and a selection of weather stations for the period 7 September 2001–10 September 2006 (Table 2). The best correlation was found for the nearest station Sognefjellshytta (r=0.97), but Fokstugu, 60 km northeast of the glacier, had nearly the same correlation (r=0.95). For stations Bråtå and Kjøremsgrende, correlations were slightly lower (r=0.90). None of the stations could provide continuous series for the entire period of mass balance measurements (1949–2006), but a combination of Fokstua and Fokstugu or the Kjøremsgrendi-Dombås series

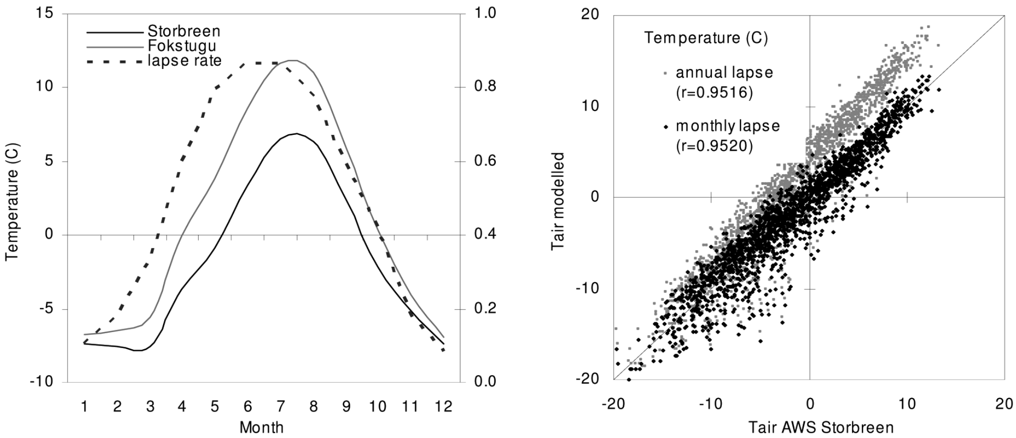

Fig. 3. Left: Monthly temperatures and lapse rates calculated from Fokstugu (972 m a.s.l.) and AWS at Storbreen (1570 m a.s.l.). Right: Modelled daily mean air temperature (T_{air}) of the AWS at Storbreen using monthly and annual lapse rates from Fokstugu. Daily mean temperature at Storbreen AWS is observed at *c*. 5.6 m above the ice surface, whereas the Fokstugu temperature is measured at the 2 m level. The period for both figures is 7 September 2001–10 September 2006

could be linked as a series (Øyvind Nordli, met.no, personal communication). The Fokstua station 16600 was in operation from 1923 to 1968 and then relocated and replaced by 16610 Fokstugu from 1968 (Table 2). Temperature records from the two stations are considered to be homogeneous for spring, summer and autumn temperatures (March–November), but inhomogeneous for winter temperatures (Øyvind Nordli, met.no, personal communication). Winter temperatures are not considered to be so important though, as little melting occurs in winter (Andreassen *et al*. 2008). At Kjøremsgrendi temperature data are lacking for one year. We therefore decided to use the combined series Fokstua/Fokstugu as temperature input to our model. A closer look at the temperature records of Storbreen and Fokstugu revealed that monthly data are highly correlated over the period 2001–2006 (r=0.97). Furthermore, the temperature difference between Storbreen AWS and Fokstugu is larger in summer than in winter, probably due to cooling of the air by the cold glacier surface. Calculated monthly lapse rates show a clear seasonal cycle, with a maximum in June and July and a minimum in January and December (Fig. 3a). Air temperature at the Storbreen AWS altitude, modelled from temperature data from Fokstugu, significantly improves when monthly lapse rates are applied instead of a constant lapse (Fig. 3b). Note that although the r values are similar, the annual lapse rate produces far too high temperatures in summer. Furthermore, temperatures at Fokstugu are sometimes significantly lower in winter compared to the AWS, probably due to less mixing of the air at Fokstugu than on the glacier.

We also calculated **mean summer temperature for June, July, August, T_{JJA}**, from the Fokstua/Fokstugu dataset and compared this to the calculated b_s at Storbreen. We used the specific b_s ignoring effects on the b_s of changing glacier geometry. For the period 1949–2006 the correlation (r) between T_{JJA} and b_s was 0.856. For the period 1949–1967 when station 16600 Fokstua was used, r was 0.866, and for the period 1968–2006 when station 16610 Fokstugu was used, r was 0.882. The difference in r for the two periods is only 0.015 and does not indicate a systematic shift in summer temperature due to the change in station location.

Mass balance model

The mass balance model calculates the mass balance, b(z), as the sum of ablation and solid precipitation:

$$b(z)=\sum\left\{\min\left(0;\frac{-Q}{L_m}\right)+P_{solid}\right\} \qquad (2)$$

where Q is energy available for melt, L_m is latent heat of melting and P_{solid} is precipitation as snow. Melting (and runoff) occurs when the surface energy flux is positive at a rate Q/L_m (after Oerlemans 2001, p 44). The contributions from ablation and

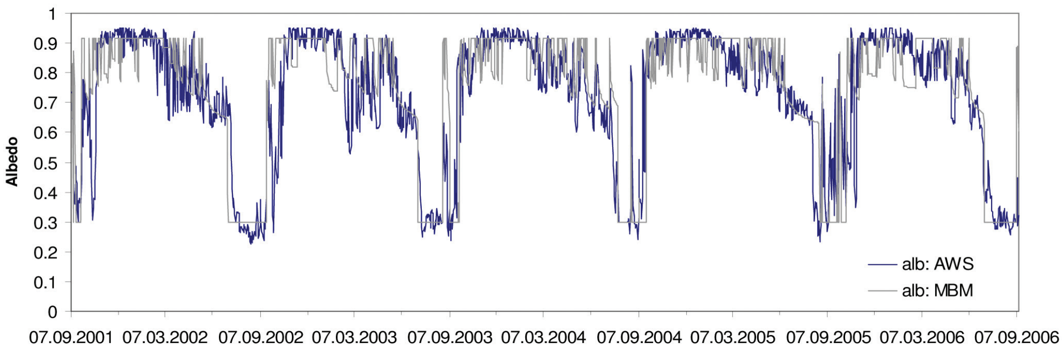

Fig. 4. Albedo simulated with the mass balance model (alb: MBM) for the period 2001–2006 at the AWS altitude compared with albedo measured at the AWS (alb: AWS)

solid precipitation are added together to yield the specific balance. If the melting energy is negative ($Q < 0$) there is no ablation and the contribution from ablation is set to zero. The mass balance model is a profile model which calculates mass balance in a total of 15 points from 1,375 to 2,075 m a.s.l. A finite time interval of one day is used as the time step.

Ablation

Melt at the glacier surface is controlled by the surface energy balance. The melt energy can be calculated as the sum of incoming and outgoing shortwave (solar) and longwave radiative fluxes, turbulent heat fluxes (sensible heat and latent heat flux) and subsurface heat flux. In our approach, we use a simplified model proposed by Oerlemans (2001) where melt energy, Q, is parameterised as:

$$Q = \tau(1 - \alpha)Q_{TOA} + c_0 + c_1 T_{air} \tag{3}$$

where τ is total transmissivity (the ratio between mean shortwave incoming radiation and top of the atmosphere radiation), is surface albedo, Q_{TOA} is top of atmosphere radiation, T_{air} is air temperature, and the term $c_0 + c_1T_{air}$ represents the sum of the longwave radiation and turbulent heat fluxes. The contributions of energy by rain and refreezing are neglected.

The ablation routine was calibrated with data and **surface energy balance (SEB)** calculations derived from the AWS for the period 6 September 2001–10 September 2006 (Andreassen *et al.* 2008). In the following paragraphs we describe input data, selected values and model parameterisations in more detail.

The air temperature, T_{air}, at the AWS-altitude (1,575 m a.s.l.) was calculated from daily air temperature data from Fokstua/Fokstugu using monthly lapse rates (as shown in Fig. 3a–b). A linear temperature lapse rate of 0.65°C/100 m was applied at all glacier altitudes. A constant value of transmissivity, τ, of 0.43 was chosen based on measurements of the AWS in the main ablation season. Daily values of Q_{TOA} were calculated using standard formula. The constants c_0 and c_1 were derived from eq. 3 using the mean seasonal surface energy balance fluxes calculated at the AWS (Andreassen *et al.* 2008; Table 2). Several sets of c_0 and c_1 were tested by comparing simulations with the SEB-model and MBM-model and we chose c_0=–20 W/m^2 and c_1=8.5 W/(m^{2}K) as references for the model.

In modelling studies ice albedo is typically held constant, whereas snow albedo is often parameterised as function of snow depth and age (Oerlemans and Knap 1998) or accumulated temperature (Brock *et al.* 2000; Pellicciotti *et al.* 2005). We used a parameterisation by Brock *et al.* (2000) who parameterised albedo as a function of snow depth and accumulated temperature using separate parameterisations for snow albedo of deep snow (α_{ds}) and shallow snow (α_{ss}). We used their parameterisation, but tuned the constants so it fitted reasonably with observed albedo measurements at the AWS. Whereas Brock *et al.* (2000) and Pellicciotti *et al.* (2005) used daily maximum positive (relative to 0°C) temperatures since snowfall for their routines, Klok and Oerlemans (2004) calculated accumulat-

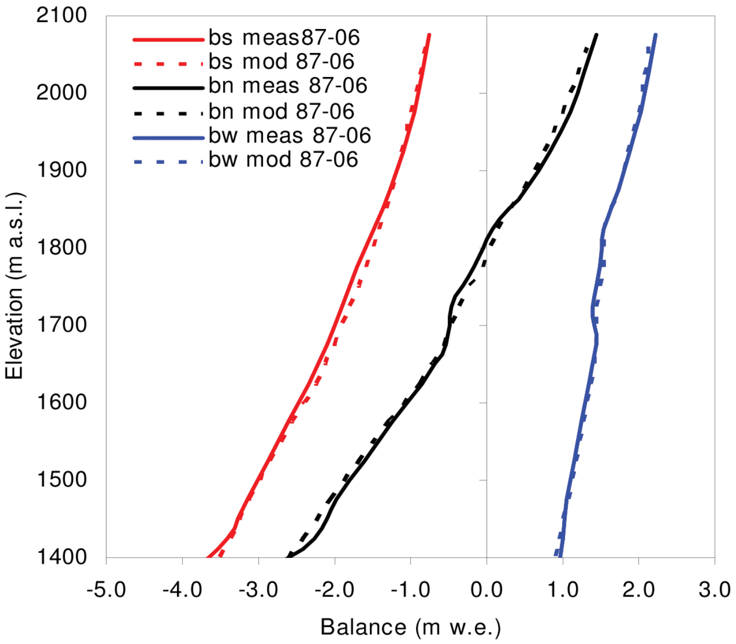

Fig. 5. Mean measured and modelled winter (b_w), summer (b_s) and net (b_n) balance profiles for Storbreen for the period 1987–2006

ed temperatures relative to –20°C to account for decay of snow albedo at temperatures below the melting point, following a study by Winther (1993). We tested both approaches, and chose to use –5°C as the minimum for the accumulated temperature for Storbreen. The equations used were:

$$\alpha_{ds}=0.86–0.08\log 10T_{sum}(i) \qquad (4)$$

$$\alpha_{ss}=\alpha_{ice}+0.35\exp(–0.06T_{sum}(i)) \qquad (5)$$

$$\alpha=\left[1-\exp\left(-\frac{d}{0.01}\right)\right]\alpha_{ds}+\exp\left(-\frac{d}{0.01}\right)\alpha_{ss} \qquad (6)$$

The variables d and T_{sum} are dimensionless and defined as d=[d_{sd}] where d_{sd} is snow depth in m w.e., T_{sum}=[T_{acc}] where T_{acc} is accumulated temperature in °C relative to –5°C since last snowfall, and i is the altitudinal interval. When d_{sd} <=0 albedo is set to albedo of ice (below 1750 m a.s.l.) or firn (above 1750 m a.s.l.). We defined snowfall as observed P at station Bulken ≥ 0.003 m w.e. The constants and tuning parameters in eq. 4–6 were derived by adjusting the original values of Brock *et al.* (2000) until modelled albedo captured the overall evolution in albedo measured at the AWS. The MBM-model produced overall satisfactory results; the model captured the main decay of snow albedo and many of the individual snow falls in late summer (Fig. 4). However, much of the variability in the winter season was not reproduced by the model, but these albedo fluctuations have little impact on the net balance as the energy available for melt is seldom positive in the winter months. Ice albedo was assumed to be constant, a value of 0.30 was chosen. Measurements from the AWS as well as the point measurements showed, however, that ice albedo varies from *c.* 0.2–0.4, but it is difficult to account for such spatial and temporal variation in the model. Firn albedo was chosen to be 0.4.

Accumulation

Precipitation was considered to fall as snow (P_{solid}) if T_{air} was < 1°C. The average daily precipitation data from two stations, Bulken and Røldal, were used as input (weighted average). To extrapolate the precipitation measured at the station to the glacier, the amount of precipitation was multiplied by a tuning factor, P_{corr}, determined by comparing measured and modelled b_w. Storbreen has an undulating surface, and snow depth varies both with elevation and topography. Published b_w-profiles over 1987–2006 include these non-linear patterns (Fig. 5). To account for this in the model, we applied a

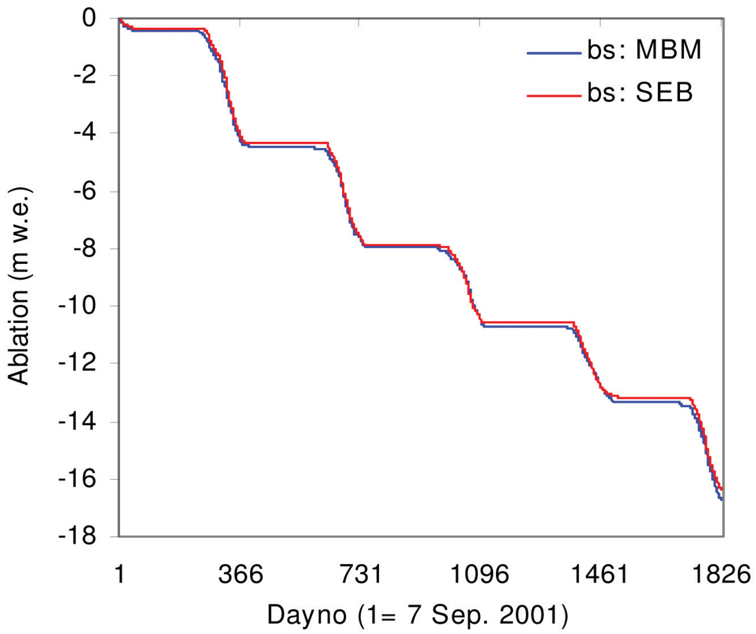

Fig. 6. Modelled ablation (runoff) at the AWS location at Storbreen (*c*. 1570 m a.s.l.) over the period 2001–2006 for the mass balance model used in this paper (b_s: MBM) and using an energy balance model (b_s:SEB, Andreassen *et al*. 2008)

correction factor (prate(i)) for each altitudinal interval so that the mean modelled b_w-profiles at the end of the accumulation season were similar to the mean of the published b_w-profiles over 1987–2006.

Model calibration and runs

To calibrate and validate the model we compared the model output at the AWS altitude interval (1575 m a.s.l.) with the observed and calculated data from the AWS as previously described for air temperature (Fig. 3) and albedo (Fig. 4). We also compared the cumulative melt (as runoff) from the mass balance model used in this paper (MBM) with melt calculated from the surface energy balance (SEB) model (Fig. 6). For a proper interpretation of the result we point out once more that the input for the SEB consists of the detailed measurements of the AWS, whereas the MBM is driven by meteorological data from the regular weather stations as described above. The two models compare well and the melt curves follow each other closely, the difference in melt is 0.3 m w.e. for the five year period 2001–2006 (MBM: –16.7 m w.e, SEB: –16.4 m w.e.). The results from the SEB-model are previously found to compare well with recordings from stakes and elevation lowering of the surface (Figs 3 and 4 in Andreassen *et al*. 2008). Therefore, good agreement in all five years between the two models indicates that the snow albedo parameterisation, the prescription of air temperature and other model parameters work well.

We also compared the modelled profiles for 1987–2006 with measured profiles for the same period and adjusted the weighted precipitation profile (prate(i)) and the temperature lapse rate so it fitted the measured mean profiles over this period (Fig. 5). Finally, we ran the model for the period 1923–2006 starting 1 January 1923. The area distribution from 1984, which is nearly identical to 1997 (Andreassen 1999; Table 2) was used as a reference state; any changes in glacier area were neglected. To compare measured and modelled values of b_w, the model produced output b_w from the day of the ablation measurements in the autumn to the day of accumulation measurements in the spring when these dates were known. Similarly output b_s was derived from the day of accumulation measurements ($b_w$day) to the day of ablation measurements ($b_s$day). For the years between 1923 and 1985 the mass balance was modelled using the dates 10 September to 5 May as start and end dates for the b_w and b_s output. The b_s will be influenced by the initial snow depth at the start of the ablation season due to the difference in albedo for snow and ice. To study the performance of b_s

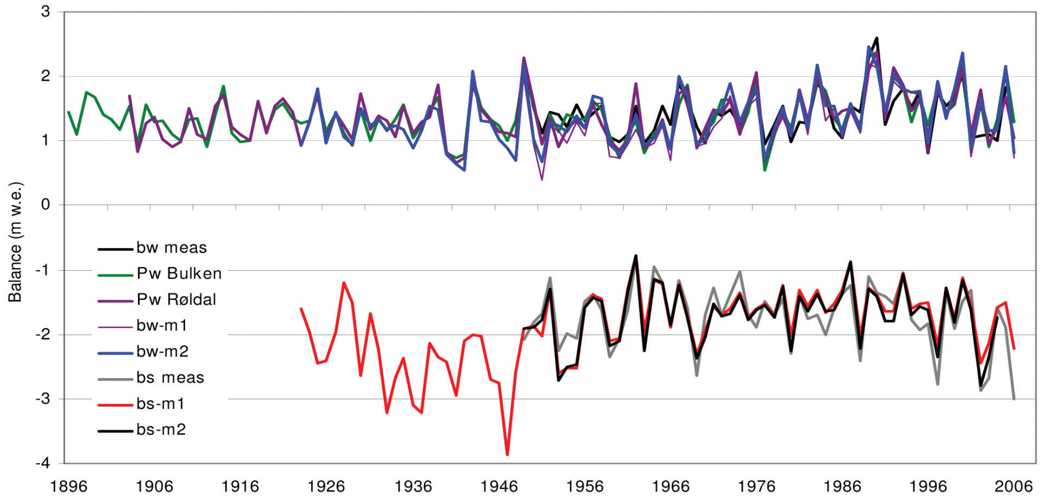

Fig. 7. Reconstructed winter and summer balance of Storbreen (1896) for years 1924–1948 and modelled and measured winter balance and summer balance for Storbreen for years1949–2006. P_w Bulken and P_w Røldal represent annual winter precipitation sums calculated from the stations and multiplied by a constant tuning factor. See Table 3 for explanation of the other runs

we therefore modelled b_s in two ways, in the reference run ($b_{s\text{-}m1}$) we used modelled snow depth (in m w.e.) as input. In the second run ($b_{s\text{-}m2}$), the modelled snow depth was replaced with measured b_w at the day of accumulation measurements (for the period 1949–2006). The snow depth was calculated by multiplying the measured specific b_w for that year with a weighted factor at each altitude (determined as the mean over the period 1987–2006 as for this period winter balance profiles are available). In this way we could approximate the measured snow depth (in m w.e.) for each altitudinal interval for the entire series of measurements 1949–2006.

Model Results

Modelled winter balance (bw) and summer balance (bs), 1949–2006

In general, modelled winter and summer balance over 1949–2006 correlated well with observed values (Table 3). Correlation (r) for winter balance was 0.84 and the resulting root mean square error (rmse) was 0.25 m w.e. for the reference run ($b_{w\text{-}m1}$). Separating the period in two to check the possible effect on relocation of the air temperature station from Fokstua to Fokstugu in 1968 reveals that the model underestimates winter balance in the first period 1948/49–1967/68, the rmse is noticeably higher in this period (0.33 m w.e.), than in the sec-

Table 3. Correlation (r) and root mean square error (rmse) between modelled and measured winter (b_w), summer (b_s) and net (b_n) balances at Storbreen. $b_{w\text{-}m1}$: reference modelling b_w where P_{solid} when $T_{air} < 1$ºC, $b_{w\text{-}m2}$: P_{solid} when $T_{air} < 3$ºC, $b_{s\text{-}m1}$: b_s modelled using modelled b_w ($b_{w\text{-}m1}$) as snow depth, $b_{s\text{-}m2}$: b_s modelled using measured b_w as snow depth, $b_{n\text{-}m1}$: b_n=$b_{w\text{-}m1}$+$b_{s\text{-}m2}$, $b_{n\text{-}m2}$: b_n=b_w+$b_{s\text{-}m2}$. *Notes*: n: number of years.

	Period	n	$b_{w\text{-}m1}$	$b_{w\text{-}m2}$	$b_{s\text{-}m1}$	$b_{s\text{-}m2}$	$b_{n\text{-}m1}$	$b_{n\text{-}m2}$
	1949–2006	58	0.84	0.86	0.83	0.87	0.87	0.94
r	1949–1968	19	0.85	0.89	0.89	0.92	0.92	0.95
	1968–2006	39	0.86	0.87	0.87	0.88	0.90	0.94
rmse	1949–2006	58	0.25	0.22	0.28	0.24	0.30	0.22
(m w.e.)	1949–1968	19	0.33	0.23	0.24	0.22	0.45	0.22
	1968–2006	39	0.21	0.23	0.29	0.24	0.32	0.24

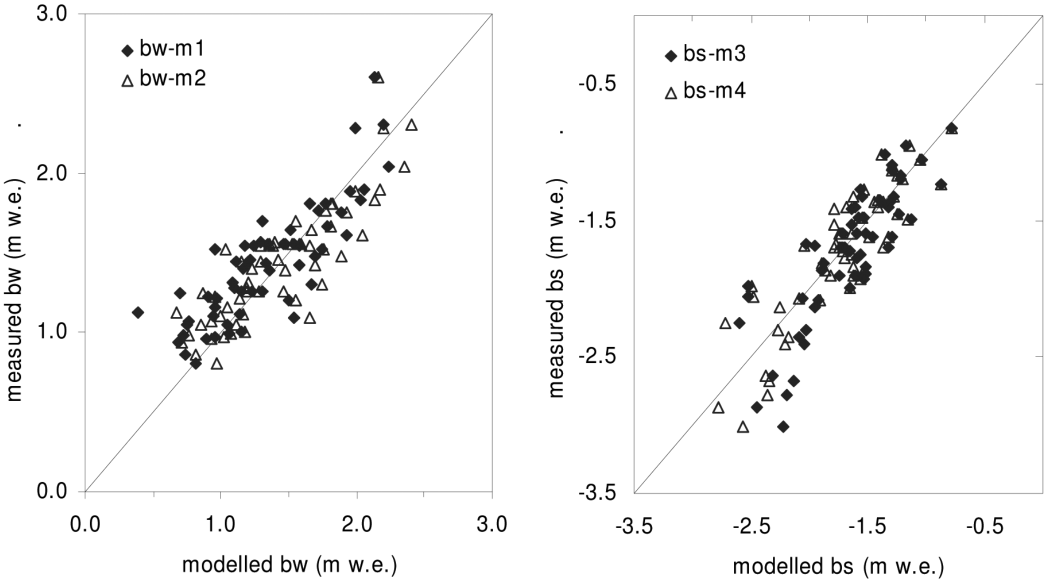

Fig. 8. Scatter plots of measured versus modelled specific winter balance (b_w) and summer balance (b_s). $b_{w\text{-}m1}$: reference run modelling b_w where P_{solid} when T_{air}<1ºC and T_{air}<3ºC, $b_{s\text{-}m1}$: b_s modelled using modelled b_w ($b_{w\text{-}m1}$) as snow depth; $b_{s\text{-}m2}$: b_s modelled using measured b_w as snow depth

ond period 1968/1969–2005/2006 (0.21 m w.e.). The rmse values calculated from precipitation sums (P_w) for Bulken and Røldal do not show such a difference in rmse for the two periods, the difference in rmse is within 0.02 m w.e. (not shown). Increasing the threshold for snow (P_{solid}) from 1 to 3°C (model run $b_{w\text{-}m2}$) resulted in a higher overall correlation (r=0.86) and lower rmse for the 1949–2006 (0.22 m w.e.) and the 1949–1968 periods (0.23 m w.e). Hence, results suggest that the inhomogeneous winter temperatures between Fokstua and Fokstugu stations are causing this underestimation and that the monthly lapse rates calculated from Fokstugu are overestimating the air temperature in winter in this period. Summer balance was better modelled using measured b_w ($b_{s\text{-}m2}$: r=0.87) instead of modelled b_w ($b_{s\text{-}m1}$: r=0.83) as initial snow condition at the start of the ablation season (Table 3). The corresponding rmse values are 0.24 and 0.28 m w.e.

The b_s was better modelled for the first period than for the second period. However, on the whole the model satisfactorily captured the main characteristics and the inter-annual variability in summer and winter balances over the whole period 1949–2006 (Fig. 7) and for high and low values of b_w and b_s (Fig. 8). The model also showed good performance in years with high winter balances as in the period 1989–1995. Resulting net balances are well correlated and (obviously) best modelled as the sum of b_w measured and b_s modelled ($b_{s\text{-}m2}$). The best obtained correlations of modelled b_w (0.86), b_s (0.87) and b_n (0.94) for 1949–2006, show that the model could explain 73, 76 and 88% of the variance over this period.

From each year we also extracted the day with maximum b_w (b_wd) and minimum b_s (b_sd, minimum values as b_s is negative), at the approximate position of the **equilibrium line altitude (ELA)** of 1775 m a.s.l. as an indication of the length of the ablation and accumulation season (Fig. 9). Comparing the available dates of accumulation and ablation measurements since the 1980s reveals that maximum b_w (accounting for melt) often occurs after the b_w measurements and that minimum b_s often occurs before the observation day at this altitude. These results must be interpreted with care as, for instance, one single day with melt late in the season (with no accumulation in between) may result in a high value of b_sd. Nevertheless, calculated values as well as nine year running means indicate that the ablation season has become longer since *c*.1990 due to extended melt in fall.

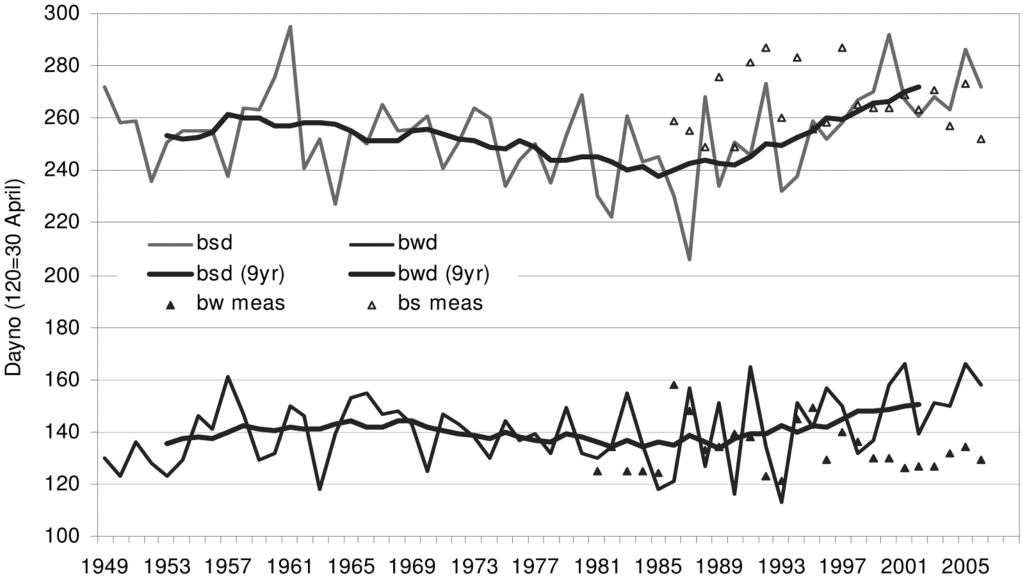

Fig. 9. Day of maximum winter balance (b_wd) and minimum summer balance (b_sd) for altitude 1775 m a.s.l. Day no.: 120 = 30 April, 160=9 June, 220=8 August, and 280=7 October. A nine year running average filter is included for both series

Reconstruction of mass balance prior to 1949

The mass balance of Storbreen prior to 1949 was modelled using $b_{w\text{-}m2}$ (P_{solid} when $T_{air} < 3°C$ in winter) and $b_{s\text{-}m1}$. Reconstructed specific mass balance over 1923/24–1948/49, representing 25 mass balance years, shows highly negative summer balances (Fig. 7). The resulting cumulative mass deficit is *c*. –30 m w.e., nearly twice as much as for the 58 year period 1949–2006 (–17 m w.e.). The mean specific winter, summer and net balance was 1.2, –2.4 and –1.2 m w.e., respectively for 1924–1948, compared to the 1.4, –1.7 and –0.30, respectively, measured for the period 1949–2006. In perspective, the highest b_s measured at Storbreen was –2.6 m w.e. in 2006. Modelled b_w show low winter balances in the 1940s. In the same period b_s is highly negative in several years, the most negative b_s was modelled for 1947 (b_s =–3.9 m w.e.).

The winter balance was also reconstructed from precipitation data back to 1896 (Bulken) and 1903 (Røldal), respectively. The b_w was calculated by multiplying the winter precipitation sum, P_w, for each year by a tuning factor, P_{corr}, derived over the period 1949–2006, so that:

$$b_{w(1949–2006)} = P_{corr} * P_{w(1949–2006)} \quad (8)$$

The b_w reconstructions from P_w at Bulken and Røldal indicated a slightly higher mean b_w for the period 1923/24–1948/49, 1.3 m w.e., than calculated by the model using a temperature threshold of 3°C for P_{solid}. The b_w for the period prior to 1895/96–1922/23 was also 1.3 m w.e. using precipitation data from Bulken. The reconstructed b_w further indicated that the high winter balances measured in the period 1989–1995 with a maximum in 1989 (2.35 m w.e.) were highest in the entire period 1896–2006 (Fig. 7). Only two years, 1941 and 1949, had comparable, but lower, values. A previous reconstruction of winter balance modelled by circulation indices back to 1781 showed that the high winter accumulation in the early 1990s was unprecedented throughout the entire series (Nordli *et al*. 2005).

Climate and model sensitivity

The ratio of a change in the specific mass balance of a glacier (the mass balance averaged over the surface area of the glacier) to a small change in a climatic parameter is termed the static sensitivity (e.g. Church *et al*. 2001). The static sensitivity ignores retreat changes in glacier geometry and other dynamic effects (Jóhannesson 1997). A previous

Table 4. Modelled change in specific winter (Δb_w), summer (Δb_s) and net balance (Δb_n) (in m w.e.) and in days of maximum b_w (b_wd) and minimum b_s (b_sd) using uniform perturbations in climate parameters. The period 1969–2006 is used as reference.

Change	Δb_w	Δb_s	Δb_n	b_wd	b_sd
–1°C	0.09	0.38	0.47	5	–13
+1°C	–0.14	–0.49	–0.63	–9	11
+2°C	–0.34	–1.06	–1.40	–13	16
+3°C	–0.60	–1.66	–2.26	–21	27
–10% P	–0.16	–0.05	–0.21	–2	0
+10% P	0.15	0.05	0.19	1	0
+20% P	0.31	0.07	0.38	2	–2
+30% P	0.47	0.09	0.56	2	–3
+2°C/+10%P	–0.19	–0.98	–1.18	–13	16
+2°C, +10%P, α_{ice} 0.2	–0.21	–1.09	–1.29	–13	16
+2°C, +10%P, α_{firn} 0.3	–0.20	–1.01	–1.21	–13	16

study of three glaciers in southern Norway demonstrated that maritime glaciers are more sensitive to climate change than continental glaciers (Oerlemans 1992). We analysed the static climate sensitivity of Storbreen by applying perturbations to the input data from Fokstua/Fokstugu (T_{air}) and Røldal/Bulken (P) respectively using the period 1968/69–2005/2006 as the reference period. The measured mean b_w, b_s and b_n in this period are 1.44, –1.61 and –0.14 m w.e., respectively. This period was chosen to avoid uncertainties due to inhomogeneous winter temperatures between Fokstua and Fokstugu. However, we also derived sensitivities for three other periods (1971–2000, 1987–2006 and 1949–2006) to test whether the calculations were sensitive to the choice of reference period, but the results were only slightly influenced by the chosen period (typically b_n ±0.00–0.02 m w.e.). To calculate mass balance changes b_s was modelled using snow depths calculated by the model ($b_{s\text{-}m1}$). The area distribution from 1984 was used; any changes in area were neglected. We calculated the change in b_w, b_s and b_n as well as change in days of maximum b_w (b_wd) and minimum b_s (b_sd) for a –1°C, +1°C, +2°C and +3°C change in temperature, for a –10%, +10%, +20% and +30% change in precipitation, and for a combination of +2°C and +10% P. To this combination we also tested the sensitivity of changing the ice and firn albedo.

Calculated sensitivity to a 1°C warming was –0.63 m w.e. (ranging from –0.61 to –0.66 m w.e. for the three other reference periods), whereas the sensitivity for a 10% increase in precipitation was +0.19 m w.e. (ranged between +0.19–0.20 m w.e. for the three other reference periods). This implies that a 1°C warming must be compensated by a 30% increase in precipitation to avoid mass deficit. The sensitivity to a 1°C decrease in temperature was +0.47 m w.e., thus the climate sensitivity of ±1°C is 0.55 m w.e. A decrease in precipitation of 10% reduced the b_n by 0.21 m w.e, the resulting climate sensitivity to ±10% P is 0.20 m w.e. Model results revealed that the day of maximum b_w and minimum b_s will be greatly influenced by a warming, but only weakly influenced by increased precipitation (as the b_w increases, but not so much the length of the season). A warming of 1°C will decrease (increase) the average day of maximum b_w (minimum b_s) by 9 and 11 days compared to the reference period (Table 4). For a 3°C warming the length of the ablation season may increase by *c*. 50 days. Recent climate projections suggest an accelerated increase in global temperatures, but also increased precipitation (Meehl *et al*. 2007). The change in b_n for a combination of a +2°C warming and a +10% increase in temperature is –1.18 m w.e. compared to the reference period 1969–2006 where b_n was –0.14 m w.e.

Finally we analysed how sensitive the model was to alterations in input parameters (Table 5). Changing the albedo of ice from 0.30 to 0.20 will result in a more negative b_n, –0.06 m w.e. The change will have an effect on both b_w and b_s as melting after the ablation measurements in fall will effect b_w. Changing the albedo of firn from 0.4 to 0.3 will have little effect on the results in the reference period. An increase (decrease) in transmissivity of 0.02 results in a change in b_n of –0.09 (+0.09) m w.e. The model is (of course) highly sensitive to the chosen values of c_0 and c_1 (Table 5). The model sensitivity to a decrease in ice and firn albedo is higher when the perturbation is applied in a warmer climate. Reducing the ice albedo from 0.3 to 0.2 to the combined +2°C and

Table 5. Change in specific winter (Δb_w), summer (Δb_s) and net balance (Δb_n) (in m w.e.) toalterations in values of input parameters.

Period 1969–2006	Original value	New value	Δb_w (m w.e.)	Δb_s (m w.e.)	Δb_n (m w.e.)
Albedo, ice	0.30	0.20	–0.005	–0.054	–0.059
Albedo, firn	0.40	0.30	0.000	–0.006	–0.007
Transmissivity	0.43	0.45	–0.007	–0.087	–0.094
Transmissivity	0.43	0.41	0.007	0.085	0.092
c1 (W $m^{-2}K^{-1}$)	8.5	9.0	–0.005	–0.054	–0.059
c1 (W $m^{-2}K^{-1}$)	8.5	8.0	0.005	0.054	0.059
c0 (Wm^{-2})	–20	–18	–0.010	–0.065	–0.075
c0 (Wm^{-2})	–20	–22	0.010	0.064	0.074

+10% P perturbations, results in a Δb_n of –0.09 m w.e. (Table 4). Similarly, a reduction in firn albedo from 0.4 to 0.3 will result in Δb_n of –0.03 m w.e. The reason for this increase in sensitivity is that the period of ice and firn melt will increase in a warmer climate. As an example, model calculations show that the number of days where the surface is ice at the 1575 m a.s.l. altitude (AWS-location) is 41 days in the reference period 1969–2006. Increasing the temperature by +2°C will result in more than a doubling to 89 days, and for the combination of +2°C and +10% P perturbations the length of the ice melt period is 83 days, slightly lower due to the 10% increase in P.

Discussion

Selection of input data

The good correlation between precipitation measured west of the water divide and b_w at Storbreen reveals that many stations are suitable to be used as input data. The poor correlation with stations east of the glacier points to the differences in precipitation regime between the western and eastern regions in southern Norway and that b_w at Storbreen is generally influenced by the same flow patterns as areas west of the glacier. Scrutinising data from several stations is worthwhile before choosing which station to use for input data. In a study of a total of 42 Arctic glaciers, the b_w of Storbreen was modelled with precipitation data from Fokstua as input and a much lower correlation (r^2=0.48) was obtained (De Woul and Hock 2005; Table 2) than the r^2 of 0.75 we obtained by using a combination of Bulken and Røldal. Their observation period was shorter (45 years) than the period used in this study (58 years). Rasmussen and Conway (2005) obtained r^2=0.63 using an upper-air model for the period 1948/49–1998/99. For temperature fewer long-term stations were available. The lack of continuous and homogeneous temperature series in high elevation areas is a challenge in long-term modelling. This study has illustrated problems with inhomogeneous winter temperatures when stations are relocated. Although winter and summer balance can also be modelled well using simple linear regression models, as revealed in this study for P_w and T_{JJA}, more complex models have many more opportunities to study impact of glacier change, such as decrease in ice albedo or change in ice melt season.

Model performance

Correlations between modelled and measured b_w and b_s were high and show that the mass balance of Storbreen can be modelled satisfactorily using data from weather stations outside the glacier. The obtained rmse values for b_w and b_s of 0.22–0.25 and 0.24–0.28 m w.e. respectively in the period 1949–2006 are acceptable considering the uncertainties in both modelled and measured values. Although discrepancies occurred in some years, the mass balance model was able to reproduce the main characteristics of annual and seasonal mass balance of Storbreen. Close agreement between modelled melt using a detailed energy balance model and a simplified mass balance model suggests that air temperature and albedo is satisfactorily parameterised, at least for the AWS altitude. However, the compared period covers only 5 of the 58 modelled years in this study. Furthermore, the specific winter, summer and net balances measured at Storbreen in 2001–2006 deviate markedly from the mean for the whole observation period. A combination of smaller winter balances than normal and high summer balances resulted in strong deficits for the glacier in this period. The b_s measured in 2006, 2002 and 2003 are the largest, second largest and fourth largest negative balances ever measured at the glacier, coinciding with three of the warmest summers measured in Norway since air temperature measurements began in 1876 (Andreassen *et al.* 2007; 2008). Nevertheless, the good agreement over the whole period 1949–2006 as well as for modelled and measured profiles over 1987–2006 indicates that the model set up and choice of parameters also works well for the other years in the series.

It should be emphasised that the modelled b_w and b_s will not reflect annual variations in the incoming radiation since the transmissivity is kept constant in the model. Accounting for this in the model would be difficult as long-term observations of radiation are generally rare in Norway. The closest long-term radiation series is from Bergen. These measurements indicate that there was a small decrease in incoming radiation in the 1960s and a small increase in the 1990s (Olseth 2005), such a reduction and subsequent increase in radiation is also observed at many other stations in the world, and is referred to as global dimming and brightening (Wild *et al.* 2005). A change in incoming radiation will directly influence the surface energy balance and thus the energy available for melt as illustrated by changing the transmissivity values (Table 5). It should be noted though that a reduction in shortwave incoming radiation due to changes in cloud conditions instead of atmospheric aerosol content will increase the longwave incoming radiation and to some extent compensate for the change in shortwave incoming radiation (e.g. Greuell and Genthon 2004). Moreover, the model will not reflect any spatial or temporal variations in ice albedo since ice albedo is kept constant. Despite these limitations, the good model performance in our study seems to support the finding by Andreassen *et al.* (2008) that the inter-annual variability in melt is mainly caused by variations in air temperature and albedo, although radiation is the most important contributor to melt. Nonetheless, a model is not expected to have a 1:1 correlation with observed values as there will be uncertainties in meteorological observations and model parameters as well as in the mass balance calculations. Measured mass balance does not necessarily represent the 'true values' of a glacier's mass balance. Uncertainties in the mass-balance measurements are dependent on both the accuracy of the point observations and conversion of point values to area-averaged values. As described in the data section and also discussed by Andreassen (1999), the extent of the monitoring programme at Storbreen has varied and in some years the measurements were rather sparse. Long-term series are likely to be inhomogeneous due to changes in personnel, monitoring programmes and calculation methods (Braithwaite 2002). A previous comparison of the measured cumulative mass balance of Storbreen with mass balance calculated from maps revealed that the geodetic mass balance was in reasonable agreement with the direct measurements and concluded that neither the direct nor the geodetic method could be used to verify the other method as there are uncertainties in both methods (Andreassen 1999). Furthermore, a true comparison between the measured and modelled mass balance can not be undertaken due to uncertainties in the observation dates before 1985. We also neglect geometric changes by keeping the geometry constant using the 1984-area-altitude distribution, whereas a series of maps have been used to produce the mass balance results of Storbreen (Liestøl 1967; Andreassen 1999). Ideally, the mass balance series of Storbreen should be converted to a reference-surface balance as proposed by Elsberg *et al.* (2001) which omits the influences of changes in area and surface elevation. However, Rasmussen *et al.* (2007) found that for Storbreen, when the mean 1950–2002 $b_n(z)$ profile is integrated over hypsometries for 1997 and 1951, the effect of the changing geometry is 0.2 m w.e. over 1949–2005, which is small compared with the total –15 m w.e. change for this period.

The reconstructed values prior to 1949 must be considered estimates, rather than the constructed specific values, as they relate to the 1984-geometry, not the actual geometry of Storbreen in this period. Little information is available to validate the results, some older maps exist, but these map surveys are generally not accurate enough to be used for volume change calculations. The error in the reconstructed mass balance is the sum of random and systematic errors. Assuming that the error for each year is truly random, the standard error for the cumulative period, *T*, can be calculated as:

$$T = \sqrt{xt^2} \tag{9}$$

where *x* is number of years with measured values and *t* is the average standard error for each year with measured mass balance (following Andreassen *et al.* 2002). The average accuracy of the annual net balance is estimated to be ±0.3 m w.e. for Storbreen (e.g. Kjøllmoen *et al.* 2007). Subjectively assuming a higher uncertainty in the reconstructed annual b_n of ±0.6 m w.e., the errors in the cumulative b_n of the 25 reconstructed years are calculated to be ±3.0 m w.e. Systematic errors, however, will cause larger errors in the cumulative estimates, for example a systematic error of +(–) 0.2 m w.e. per year will underestimate (overestimate) the cumulative balance by 5 m w.e. in the reconstructed period.

Climate sensitivity

The calculated climate sensitivity of Storbreen (using the 1969–2006 reference period) was –0.63 m w.e. for a 1°C warming and +0.20 m w.e. for a 10% increase in precipitation. Compared to other studies of Storbreen the precipitation sensitivity is higher in our study than the +0.12 m w.e. obtained by De Woul and Hock (2005) and +0.15 m w.e. obtained by Rasmussen and Conway (2005). Our sensitivity to a 1°C warming is similar to the –0.65 m w.e. calculated by De Woul and Hock (2005), but significantly higher than the –0.42 m w.e. calculated by Rasmussen and Conway (2005). A direct comparison cannot be made, however, as the sensitivity numbers will vary according to the way they are computed (model formulation, input data, time periods, etc.).

Conclusions

A mass balance model using daily values of temperature and precipitation from meteorological stations outside the glacier has been applied and tested for Storbreen. Measurements from an AWS operating in the ablation zone were used to calibrate and validate the model. A major effort was needed to deal with the lack of continuous series of meteorological input data in the vicinity of the glacier, and to find the best stations to use as input data. Relocation of stations is a problem that caused uncertainties due to inhomogeneous winter temperatures. Data from precipitation stations southwest of the glacier were best correlated with the observed winter balance at Storbreen. Modelled winter and summer balances compared well (r=0.83–0.87) with observed values for the period 1949–2006. Although discrepancies occurred in some years, the mass balance model was able to reproduce the main characteristics of winter and summer balance of Storbreen. Close agreement between modelled melt using a detailed energy balance model and the simplified mass balance model suggested that air temperature and albedo was satisfactorily parameterised. Climate sensitivity calculations suggested that a 1°C warming must be compensated by a 30 % increase in precipitation to avoid mass deficit and that the day of maximum b_w and minimum b_s will be greatly influenced by a warming. Model results indicated that warming of 1 (3)°C will increase the length of the ablation season by *c.* 20 (*c.* 50) days. The model sensitivities to ice and firn albedo will increase in a warmer climate due to earlier removal of the snow pack and thus extension of the ice and firn melt periods. The model was used to extend the mass balance series 25 years back in time. Results suggested a cumulative mass deficit of Storbreen in the order of 30 m w.e., mainly due to highly negative summer balances, but also partly due to smaller b_w than for the period 1949–2006.

As the simplified mass balance model used for Storbreen needs only daily precipitation and temperature as input (observations that can be downloaded in near real-time) it can be run on an operational basis. The model also has potential for being applied to other mass balance glaciers in Norway. Information about the local climate and mass balance is mandatory for accurate results.

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