Glacial mass balance of Austre Brøggerbreen (Spitsbergen), 1971–1999, modelled with a precipitation–run-off model

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An energy balance based HBV model was calibrated to the run-off from Bayelva catchment in western Spitsbergen, Svalbard. The model simulated the glacier mass balance, and the results were compared to observations at Austre Brøggerbreen for the period 1971–1997. Even though the model was optimized to observed run-off from a catchment in which the glaciers constitute 50 % of the area, and not to the observation of glacier mass balance, the model was able to reconstruct the trends and values of the mass balance found through observations. On average the simulation gave a negative net balance of 696 mm. The observed average is 442 mm. The simulated winter accumulation was in average for the same period 9 % lower and the summer ablation 17 % higher than the observed. The years 1994–96 show deviations between simulated and observed winter accumulation up to 160 %. This can probably be accounted for by extreme rainfall during the winter, leading to thick ice layers which make accurate observations difficult. The higher simulated summer ablation might indicate that the glaciers in the catchment as a whole have a larger negative mass balance than Austre Brøggerbreen. The simulations showed that the glacier mass-balance would be in equilibrium with a summer temperature 1.2 °C lower than the average over the last decades or with a 100 % increase in the winter (snow) precipitation. These are higher values than former estimates. A combined change of temperature and precipitation showed a synergic effect and thereby less extreme values.

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For operational hydrological purposes, techniques for predicting run-off from glacierized basins have been used for many years. Early models for predicting drainage of water from glaciers have been summarized by Fountain & Tangborne (1985). These modelling approaches have been further developed and have more recently been applied the mass balance of glaciers. Glacial mass balance simulations have been attempted by, for example, Tangborn (1999), Oerlemans et al. (1998) and Jóhannesson et al. (1993). Tangborn (1999) was able to reconstruct the observed mass balance of South Cascade Glacier, Washington State, for the period 1959–1996 using an ablation model based on melt indices for snow and ice, mean daily temperature and diurnal temperature range. A year by year comparison of the results showed substantial deviation between simulations and observations with a standard error of 0.75 m w.e. (water equivalent) in the annual balance. Moreover, Jóhannesson et al. (1993) found that the model was able only to explain 50 - 75 % of the year to year variation in the summer balance using a degree day model for calculating ablation. His model poorly predicted the interannual variations of the winter balance.

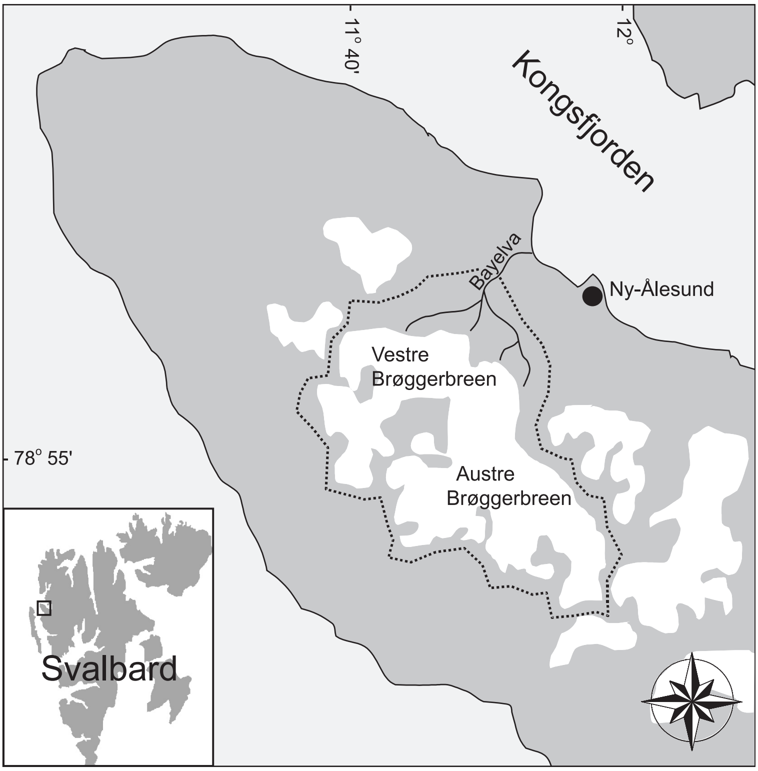

Fig. 1. Location of Ny-Ålesund and the study area

In this study a variant of the HBV model is applied to calculate summer and winter balances of Austre Brøggerbreen. The HBV model is by far the most commonly used precipitation–run-off model in Scandinavia, and is considered the "standard" run-off forecasting model for hydropower utilities in Norway. It is also widely used in other countries (Bergström 1992). The model has been calibrated and validated with run-off data from Bayelva, the outlet river from the two glaciers, Austre and Vestre Brøggerbreen, for the years 1974–78 and 1989–1997. It utilizes meteorological data from observations made in Ny-Ålesund for the period 1970–1997 by the Norwegian Meteorological Institute (DNMI). The calculated glacial mass balance of Austre Brøggerbreen has been compared to Norwegian Polar Institute's mass balance observations at Austre Brøggerbreen.

Since 1967 the Norwegian Polar Institute has measured the mass balance of Austre Brøggerbreen, a glacier in the vicinity of Ny-Ålesund, on the peninsula of Brøggerhalvøya, Svalbard (Fig. 1). With the exception of two years, the record shows that the glacier has decreased every year at an average of 423 mm w.e. averaged over the glacier surface (Hagen & Liestøl 1990; Lefauconnier et al. 1999). Lefauconnier & Hagen (1990) reconstructed the mass balance on Brøggerbreen since 1912 based on regression analysis with meteorological data and showed that the glacier has experienced a nearly constant negative mass balance since about 1920. They also showed that the negative balance was due to an increase in summer temperatures during the period 1910–1920. Most of the glaciers had their maximum extension in the period 1880–1900, the termination of the "Little Ice Age" in Svalbard. Since then the glaciers have generally retreated, probably due to higher summer temperatures.

Basal melt at Austre Brøggerbreen is negligible. The glacier is polythermal and the lowermost and thinnest part of the glacier is cold-based (Hagen & Sætrang 1991). As the run-off from the glacier is entirely from surface melting during the summer, it should therefore be possible to obtain good estimates of the glacier's mass balance through a precipitation–run-off model. In climate scenarios generated by Global Climate Models

(GCM), temperature and precipitation are the two parameters that are commonly estimated with some degree of reliability. Glacial retreat or growth depends strongly on changes in these but the response is generally very slow; glaciers are therefore often used as long-term climate indicators (Haeberli & Hoelzle 1995). If a precipitation–run-off model is able to reflect observations, sensitivity analyses could reveal the effects of summer temperature and winter precipitation and the model could be used to evaluate the effect of different climate change scenarios on a glacier's mass balance.

Site description and field measurements

Hydrological and meteorological data have been collected over a number of years below Austre Brøggerbreen in the Bayelva catchment area near Ny-Ålesund, in Svalbard, 78° 55' N, 11° 56' E (Fig. 1). Detailed measurements of snowmelt and snow properties in this catchment have been made since 1992 (see Bruland & Maréchal 1999).

The catchment is 30.8 km² and the relief ranges from 10 to 737 m.a.s.l with a mean of 253 m a.s.l. The catchment is 50% glaciated. The glacier Brøggerbreen, bounded by steep mountains along the watershed divide, covers most of the upper catchment area. Brøggerbreen is divided into Austre (Eastern) and Vestre (Western) Brøggerbreen; Austre Brøggerbreen alone constitutes about 50% of the glaciated area.

The mean altitude of the glaciers is approximately equal to the mean altitude for the catchment. The lower catchment consists of moraines, riverbed, and tundra with a uniform lichen cover with patches of rock sedge (*Carex rupestris*) and mountain avens (*Dryas octopetala*) (Hisdal 1993). There are no trees or tall shrubs to influence snow distribution or melt.

DNMI have made meteorological observations in Ny-Ålesund since 1961. The mean annual precipitation (1961–1990) at their station is 385 mm/year (Førland et al. 1997). Repp (1979) did discharge measurements for the Bayelva catchment over the period 1974–78. In 1989, the Norwegian Water Resources and Energy Directorate constructed a weir in Bayelva and resumed the time series. Run-off normally starts in the first week of June and lasts until mid-September. The average annual run-off from the catchment is 1020 mm. The large discrepancy between run-off and measured precipitation can be explained by glacial mass loss, precipitation gauge catch losses, and orographically induced precipitation gradients (Førland et al. 1997). A large portion of the precipitation falls as snow during high winds and the catch losses are high. Hansen-Bauer et al. (1996) suggest the following corrections of measured precipitation; 5 - 10% increase of liquid precipitation, 65 - 75% increase of solid precipitation (snow) and around 40% increase of sleet (or mixed precipitation). Killingtveit et al. (1994) found an increase in summer precipitation of 5 - 10% (of observed precipitation) for every 100 m increase in altitude. Based on snow surveys, Tveit & Killingtveit (1994) assumed a corresponding winter (snow) gradient of 14%. On Austre Brøggerbreen, Hagen & Liestøl (1990) found a fairly constant altitudinal increase of snow accumulation of 100 mm per 100 m; equivalent to a 25% increase per 100 m altitude. In a profile study, Førland et al. (1997) found that the total precipitation on Brøggerbreen during the 1994 and 1995 summer seasons, was 45% higher than recorded at the weather station in Ny-Ålesund. It was also found that precipitation in Ny-Ålesund was strongly dependent on the wind direction. Spillover and seeder/feeder effects probably cause high precipitation events on the glaciers during winds from the south and south-west (Førland et al. 1997). They estimate an increase in precipitation with altitude of 20% per 100 m up to around 300 m. Thirty to 40% of the total catchment area is above this elevation, and Førland et al. (1997) and Hagen & Lefauconnier (1995) point out that a linear gradient of 20 - 25% might produce inaccurately high estimates of precipitation in these uppermost areas.

Since 1992, snow conditions have been observed daily in several snowpits during the ablation period and snowmelt has been measured from three run-off plots, together with observations of albedo, solar radiation, temperature and relative humidity. The average snowmelt intensities were found to be 14 mm/day. Average air temperature and incoming solar radiation during the ablation periods (1992–98) were 2.1 °C and 230 W/m², respectively. The threshold temperature for snowmelt was 0° C.

Snow accumulation on Austre Brøggerbreen is measured at the end of the winter, usually in the beginning of May, before snowmelt. Snow

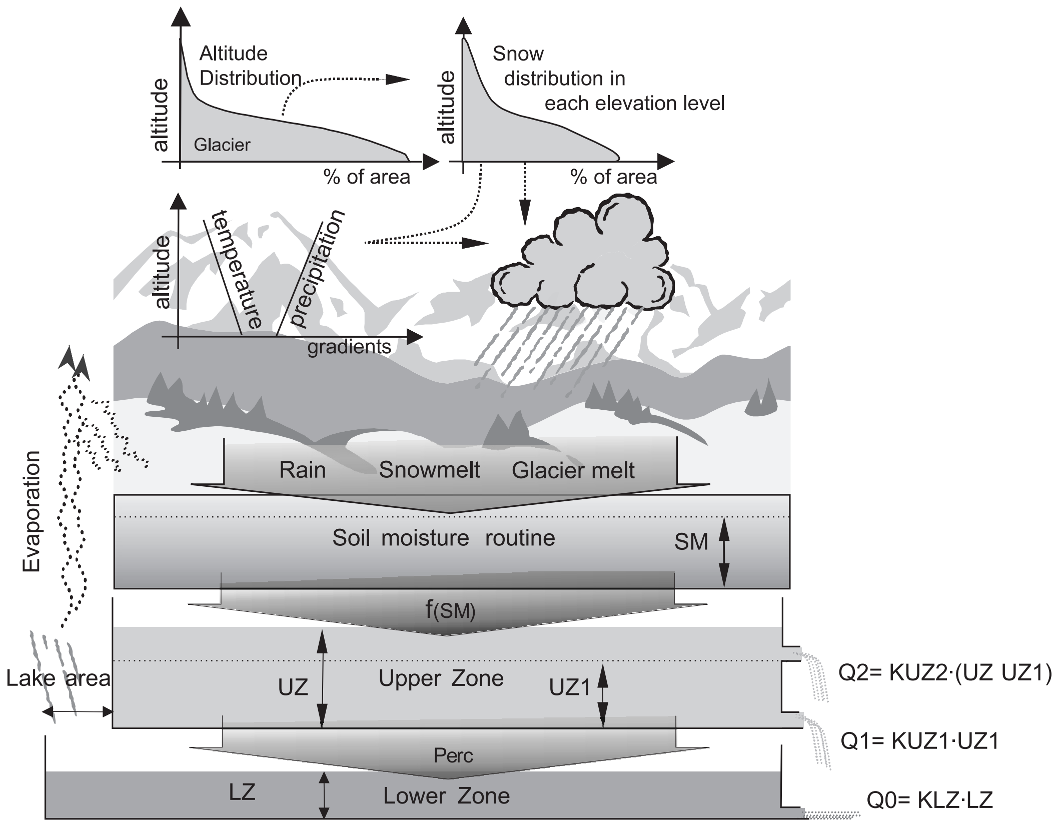

Fig. 2. The structure of the HBV model.

survey consists of point measurements every 100 m along transects across the glacier at about every 50 m height interval. Any later snowfall is not included. The main components of summer ablation are the snow and ice melt. It is measured as changes in surface level at stakes drilled into the snow and ice at about 50 m altitude intervals along a central flow line.

Model description

The HBV model is a conceptual precipitation–run-off model that uses precipitation, air temperature and potential evaporation data to compute snow accumulation, snowmelt, actual evapotranspiration, soil moisture storage, groundwater and run-off from the catchment (Fig. 2). The model was developed in the early 1970s at the Swedish Meteorological and Hydrological Institute and has been extensively documented (Bergström & Forsman 1973; Bergström 1975, 1976; Lindström et al. 1997). In addition to catchment characteristics, several parameters describing orographic precipitation gradient, precipitation gauge catch loss, temperature lapse rate, hydrographic characteristics and threshold temperatures for snow precipitation and melt have to be determined. In the original HBV model, the snow and glacial melt is calculated using a Temperature Index Melt (TIM) model. At higher latitudes and elevations, snowmelt gradually becomes the most important hydrological process, and the importance of solar radiation to melt grows, especially for the timing of run-off (Kane et al. 1997). Although the simple TIM model used in the traditional HBV model has been successfully tested and applied in several studies (Sand 1990; Hinzman & Kane 1991; Vehviläinen 1992; Hamlin et al. 1998), weaknesses have been evident in situations with low temperatures and high solar radiation (Bruland et al. 2001).

These situations are typical of Arctic and alpine areas in springtime. Bruland & Killingtveit (in press) substituted the TIM with a surface energy balance calculation. We also introduced an improved description of snow distribution on the glacier and a simulation of snow temperatures. This revised version of the HBV model, called E-bal HBV, is applied here.

Energy balance calculations

The energy balance model calculations in E-Bal HBV are based on the equations suggested by Harstveit (1984). Harstveit tested energy balance models and TIM against lysimeter snowmelt data collected during the period 1979–1982 from four locations close to Bergen, Norway. His study included testing and optimization of parameters used in the energy balance models; these optimal values are used here. Bruland & Killingtveit (in press) and Bruland et al. (2001) have previously applied Harstveit's energy balance equations to cases in Svalbard. Though Svalbard is geographically distant from western Norway, and its climate completely different, we concluded that the equations and Harstveit's empirical constants improved snowmelt calculations compared to simple temperature index calculations. The model is described by Eqs. (1)-(11). The energy balance for the snow-pack can be written:

$$Qm + Qi = Qs + Ql + Qh + Qe + Qg + Qr \quad (1)$$

where Qm is energy available for snow/ice melt, Qi is energy for internal heating and cooling of the snow-pack, Qs and Ql is net short- and long-wave radiation, respectively, and Qh and Qe is sensible and latent heat, respectively. During the ablation period in this region of Svalbard, with ground temperatures close to 0 °C and very low precipitation, ground heat flux (Qg) and heat from precipitation (Qr) can usually be neglected for practical computations. All energy terms in Eqs. (1)-(11) are expressed as W/m².

Net short-wave radiation (incoming−reflected) is computed as follows:

$$Qs = Qs_{in} - Qs_{out} \quad (2)$$

where Qs_{in} is extraterrestrial radiation corrected for atmospheric effects and Qs_{out} is reflected short-wave radiation given by the albedo of the surface.

$$Qs_{in} = Q_{ex} \cdot (s_1 \cdot C_s + s_2 \cdot C_s^{1/2} + s_3) \quad (3)$$

where Q_{ex} is extraterrestrial radiation given by the date and the latitude (W/m²), Cs is 1 − Cloudiness, varying from 0 at complete cloud cover to 1 at clear skies, s_1, s_2 and s_3 are empirical constants found in studies in Dyrdalen, western Norway, where $s_1 = -0.16$, $s_2 = 0.81$, $s_3 = 0.07$.

The calculation of the albedo in the model is based on Harstveit's (1984) regression models of albedo in Dyrdalen. He investigated correlation between observed albedo, age of snow (days), and cloud cover. His model, Eq. (4), gave a multiple coefficient of correlation for his data of $r = 0.77$.

$$A = a_1 \cdot (1 - C_s) + a_2 \cdot \ln(t) + a_3 \quad (4)$$

where t is age of snow in days, a_1, a_2 and a_3 are empirical constants found from studies in Dyrdalen; $a_1 = -0.13$, $a_2 = -0.05$, and $a_3 = 0.87$.

The cloudiness has large-scale variability and can be taken from the nearest meteorological observatory or calculated from observations of solar radiation. In our case we have available data of solar radiation back to 1980 and cloudiness observations back to 1970.

Several investigators have shown that estimates of incoming long-wave radiation from the atmosphere can be made from surface air temperature and vapour pressure, or surface air temperature and a cloud factor (US Army Corps of Engineers 1956; Swinbank 1963; Bengtsson 1976; Partridge & Platt 1976; Male & Gray 1981; Harstveit 1984; Ashton 1986). In this model the empirical formula suggested and tested by Harstveit (1984) is used (Eq. 5). His tests over 150 months of observation from Bergen gave a correlation coefficient of 0.95. The equation was in good agreement with Partridge & Platt's model during cloudy conditions, and Swinbank's model during clear sky conditions:

$$Ql_{in} = l_1 \cdot \sigma \cdot T_{air}^{\ 4} + l_2 \cdot C_s + l_3 \quad (5)$$

where σ is Stefan-Boltzmann constant, T_{air} is air temperature (K), l_1, l_2 and l_3 are empirical constants based on measurements in Bergen; $l_1 = 1.02$, $l_2 = 71$ and $l_3 = -92$. Outgoing long-wave radiation (Ql_{out}) is determined by Stefan Bolzmann's law:

$$Ql_{out} = \sigma\, T_{surf}^{4} \quad (6)$$

where T_{surf} is surface temperature (K). Sensible heat transfer (Qh) depends on the temperature difference between the air and snow surface, and wind speed. Usually empirical formulas are used to compute Qh, mostly in the form:

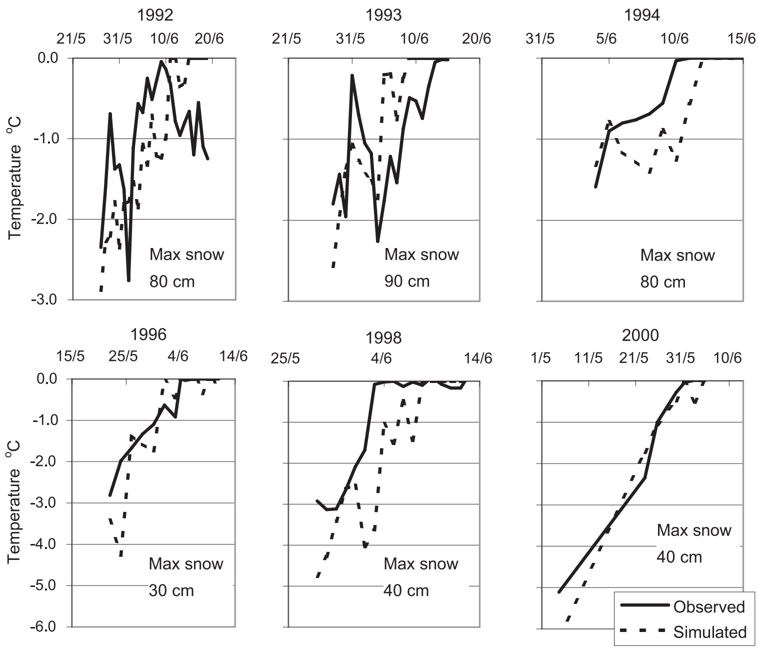

Fig. 3. Simulated average snowpack temperatures during the ablation compared with snowpack temperatures observed in snowpits. Initial snow depths (before snowmelt) are noted for each year.

$$Qh = f(u) \cdot (T_{air} - T_{surf}) \quad (7)$$

where u is wind speed (m/s) and f(u) is a function of the wind speed. Here an empirical model based on Harstveit's (1984) studies in Norway is used:

$$Qh = (w_1 \cdot u + w_2) \cdot (T_{air} - T_{surf}) \quad (8)$$

where $w_1 = 3.1$ and $w_2 = 2.3$.

Latent heat transfer (Qe) is either heat released from water vapour condensing on the snow (positive) or evaporation from the snow surface (negative). Latent heat transfer is computed in much the same way as sensible heat:

$$Qe = f(u) \cdot (e_{air} - e_{surf}) \quad (9)$$

where u is wind speed (m/s), e_{air} and e_{surf} is vapour pressure in the air and at the snow surface (mb), respectively.

Also based on Harstveit's studies, the following empirical model is used:

$$Qe = w_0 \cdot (w_1 \cdot u + w_2) \cdot (e_{air} - e_{surf}) \quad (10)$$

where $w_0 = 1.7$ and w_1 and w_2 are defined as in Eq. (8). During several occasions with high wind speeds (>5 m/s) simulated run-off was too high. At these wind speeds the linear wind function applied in Eq. (8) and Eq. (10) yielded too high turbulent heat fluxes. Different values for the coefficients w1 and w2 were tested, but the best approach was found when sensible heat was calculated using the following equation:

$$Qh = \ln(w_1' \cdot u + w_2') \cdot (T_{air} - T_{surf}) \quad (11)$$

w_1' and w_2' were selected to make Eq. (11) correspond to values found by Eqs. (8) and (10) at low wind speeds where these gave good results. At high wind speeds the logarithmic function prevents the estimated turbulent heat fluxes from becoming unreasonably large. The latent heat fluxes are so small that replacing the linear wind function with the logarithmic function did not have a significant influence on the simulations.

Since the energy balance calculations depend on the snow surface temperature, which in turn depends upon energy balance calculations, several iterations at each time step are necessary to balance the surface energy terms exactly. A simplification to this approach is required if the original HBV philosophy of simple computational approach with minimal data is to be maintained. Snow surface temperature is the key to solving the energy balance. Since the energy balance calculations in this model are most important during the ablation period, the surface temperature is set to 0 °C instead of being found through iterations. The simplified treatment of the surface temperature will have only minor effects on the performance of the model. This simplification is not applicable at sub-freezing air temperatures when the energy balance becomes negative. On these occasions, the snow surface and snow-pack cool and ablation ceases. Moreover, when the temperature in the snow-pack is below freezing, meltwater refreezes in the snow rather than causing run-off.

Snow surface temperature is also paramount to the calculation of snow-pack cooling and again iteration is necessary. Here, a simplified snow temperature calculation is suggested. The following equation was found to give snow temperatures (T_{snow}) with a fairly good fit to snow temperatures measured in snow pits (Fig. 3). The correlation coefficient was 0.77 between calculated and measured average snow temperatures.

$$T_{snow} = E_{refr} + T_n \tag{12}$$

E_{refr} is temperature increase due to energy released from refreezing of meltwater from the previous time step. The equation is based on the average air temperature (T_n) over the previous n days. As snow is a good insulator the average temperature of a thick snow-pack changes slower than for shallow snow pack and n is a function of the snow depth expressed as the remaining snow water equivalent (SWE in cm at the time step. n is set to have a maximum of 15 days.

$$n = (\mathrm{SWE})^{3/4} \tag{13}$$

T_{snow} is used to calculate the energy necessary to heat the snow-pack to isothermal condition at 0 °C. As long as the temperature in the snow-pack is below 0 °C, any snowmelt at the surface will refreeze in the snow-pack and no run-off occurs. This has a strong influence on the onset of snowmelt run-off.

Snow distribution

In most implementations of the HBV model a distributed snow routine is used to simulate the effect of an uneven or skew distribution of the snow at the defined elevation levels in the catchment (Killingtveit 1978; Killingtveit & Sælthun 1995). In this study, the effect of the skewed distribution is accounted for by reducing the snowmelt with a factor depending on snow-covered area:

$$SM = f_{snow} \cdot Q_m / L_f \tag{14}$$

where SM is snowmelt (kg/m²), f_{snow} is fractional snow covered area, Qm energy available for snowmelt (W/m²) and L_f is the latent heat of fusion (W/kg).

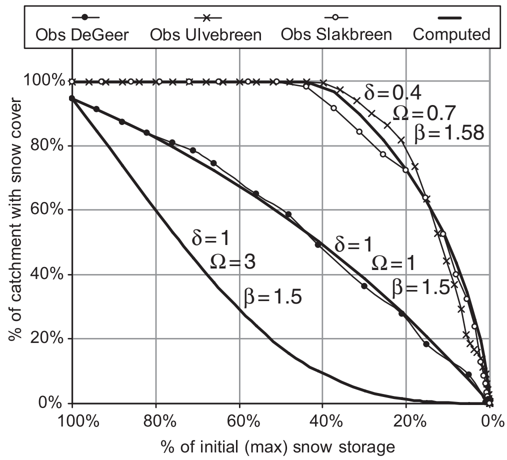

Fig. 4. Snow cover depletion curves (assuming even snowmelt) based on measured snow distribution in De Geerdalen and two glaciers in Svalbard compared to calculated depletion curves using Eq. (13)

Snow covered area, f_{snow}, will be a function of initial snow distribution and snow storage depletion:

$$f_{snow} = \beta - \frac{\beta}{\exp\left(\frac{SWE_{left}}{\delta \cdot SWE_{max}}\right)^{\Omega}},$$

$$\frac{SWE_{left}}{SWE_{max}} \geq \delta \Rightarrow f_{snow} = \text{initial snow coverage} \tag{15}$$

where SWE_{left} is snow water equivalent left at the time step, SWE_{max} is maximum snow water equivalent during the previous winter. β, δ and Ω are snow distribution skewness factors.

The values β, δ and Ω are found from the snow distribution in the catchment, and can be determined from snow survey data. Figure 4 illustrates how they influence the depletion of the snow storage. β is the initial snow coverage ranging from zero for no snow cover to 1.5 for complete snow cover. In areas with smooth surfaces, such as on glaciers, or low redistribution, such as in forests, the snow is more evenly distributed and it usually takes some time before the first snow-free patches appear. The δ value is the fraction of the total snow storage left when these patches appear and the depletion of the snow-covered area

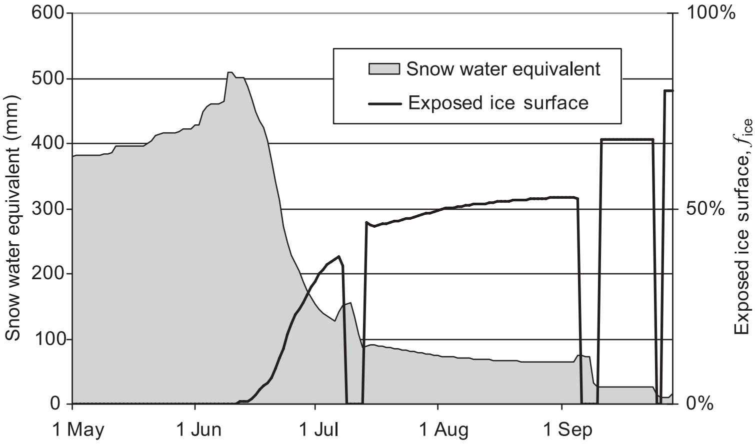

Fig. 5. Illustrated development of the glacial melt coefficient, f_{ice}.

begins. Ω reflects the skewness in the distribution of the remaining snow. With $\Omega=0$ the snow is said to be evenly distributed. Snow distribution is usually more skewed at higher elevations and δ and Ω can be set differently for each elevation band or expressed as a function of elevation. In the case of a new snowfall during the ablation season, a temporary new (redefined) SWE_{max} is used until this fresh snow has melted.

The equation was tested against data collected in May 2000 in DeGeerdalen (De Geer Valley), Spitsbergen, and data collected on several Svalbard glaciers by Winther et al. (1997). The DeGeer catchment has a relief ranging from 50 to 987 m a.s.l. Snow surveys were carried out along 5 carefully selected snow courses and the snow distribution was calculated. Assuming an even melt rate over the catchment the depletion curve will be as presented in Fig. 4. The computed depletion curve from Eq. (15), with β, δ and Ω values of 1.5, 1 and 1, respectively, has a high correlation ($r=0.99$) with the observed data. From the snow distribution data collected at the glaciers Slakbreen and Ulvebreen, representative values of β, δ and Ω were found for Svalbard glaciers (Fig. 4)

Glacial melt

In the Arctic and in several Norwegian drainage catchments, glaciation can be substantial. In the "Nordic" HBV model version (Sælthun 1996) glacial ice melt has been introduced accordingly. In the Nordic HBV model, melt is calculated the same way as for the snow with an increased rate due to the lower albedo of exposed glacier ice. Glacial ice melt at an elevation starts when the snow has completely melted. The ice thickness is assumed to be infinite, and the remaining snow at the end of the ablation is converted to ice the following year. The albedo for exposed ice ranges from 0.15 for dirty ice up to 0.51 for clean ice (Paterson 1994). A value of 0.4 is found reasonable at this location and is used as a constant in the model. The timing and rate of the glacial melt is also changed. At any elevation on the glacier the ice surface can be exposed long before all the snow at that elevation has melted. This is accounted for by letting the percentage of the glacier with exposed ice surface (f_{ice}) and thereby the glacial melt rate, be a function of the snow cover depletion given by f_{snow}:

$$f_{ice} = 1 - f_{snow} \tag{16}$$

On glaciers the snow cover is usually more evenly distributed and the snow distribution skewness factor, Ω, is lower than outside the glacier. For Ulvebreen and Slakbreen, the β, δ and Ω values were found to be 1, 0.4 and 0.7, respectively. Figure 5 illustrates how f_{ice} develops with the snow cover depletion and how it drops when a snowfall covers the previously exposed ice.

Table 1. R^2 values for the simulations.

Year	R^2	Year	R^2	Year	R^2
1974	0.76	1989	0.55	1994	0.81
1975	0.81	1990	0.67	1995	0.91
1976	0.74	1991	0.79	1996	0.87
1997	0.74	1992	0.89	1997	0.93
1978	0.75	1993	0.91	1998	0.93

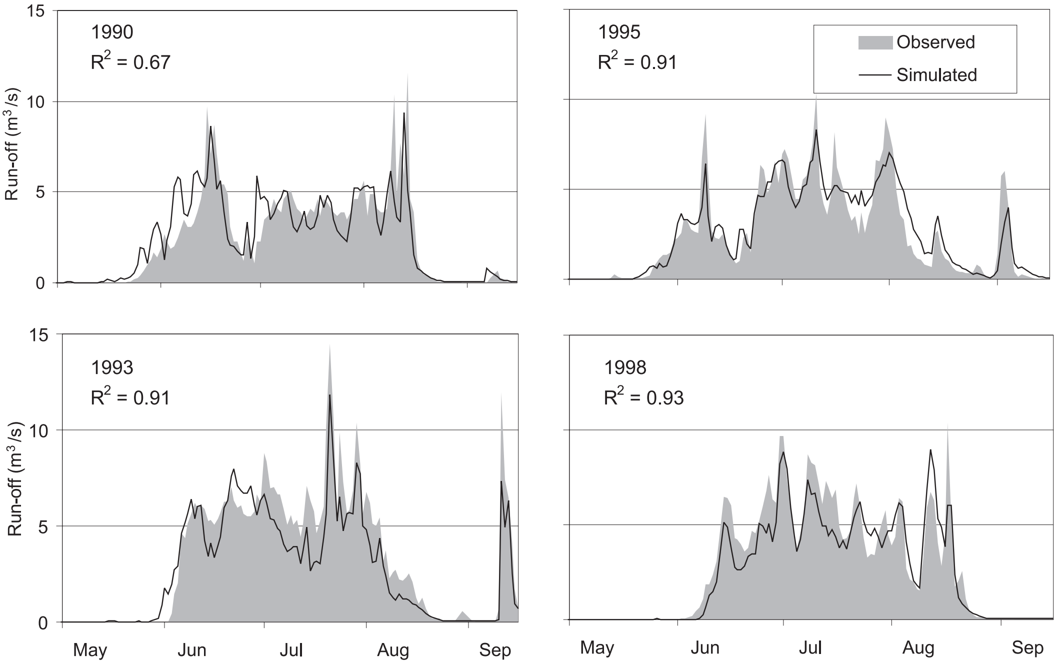

Fig. 6. Simulated run-off for the original and for the energy balance HBV model with logarithmic wind profile, compared to observed run-off.

Results and discussion

The model calculates the run-off from the entire catchment and is primarily calibrated to get the best fit to the run-off measurements in Bayelva. A selection of graphs is presented in Fig. 6. The Nash efficiency criterion, R^2 (Nash & Suthcliffe 1970), indicates the model efficiency, i.e. the agreement between the recorded and simulated hydrograph (Table 1). Compared to the original HBV model, the modified model gave a significantly better fit to observed run-off, especially at the onset of the snowmelt. The glacial mass balance calculations were also improved.

The snow accumulation on the glaciers is controlled by the correction factors for liquid and solid precipitation, the increase of precipitation with altitude, temperature lapse rate and the threshold value for solid precipitation. The values found optimal are consistent to those suggested in the literature (Table 2).

When using a value in the range suggested by Hansen-Bauer et al. (1996) for correction of solid precipitation, changing the altitude correction within the range 10 - 20 % did not significantly influence the average R^2 value. This insensitivity

Table 2. Optimal parameter values used in the model compared to values found in the literature.

Parameter	Optimal value	Values from literature
Precipitation corrections		
Liquid	1.15	1.05 - 1.10 [a]
Solid	1.65	1.65 - 1.75 [a]
Precipitation increase with altitude per 100 m	14 %	14 - 25 % [b, c]
Threshold temperature rain/snow	1 °C	1 °C [d]
Temperature lapse rate per 100 m	-0.65 °C	-0.6 °C to 1.0 °C [d]

[a] Hansen-Bauer et al. (1996).
[b] Tveit & Killingtveit (1994).
[c] Hagen & Liestøl (1990)
[d] Killingtveit & Sælthun (1995).

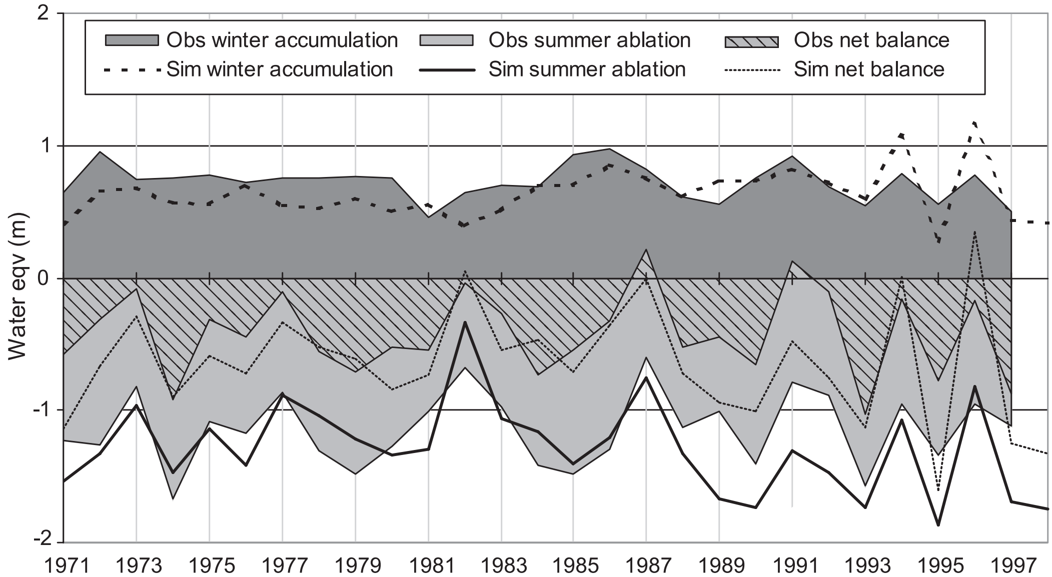

Fig. 7. Simulated and observed glacial mass balance.

can probably be explained by the importance of the cyclonic pattern. Førland et al. (1997) found this very important to the precipitation distribution on the Brøgger glaciers and it changes interannually and probably seasonally as well. At one station, located at 202 m a.s.l. on the glacier, they found that the precipitation was 33 % higher in 1994 and 78 % higher in 1995 than measured values in Ny-Ålesund. Tests in this study show that a large correction factor gives better results during the early years. From the late 1980s a lower correction is better. Here a value of 14 % is used for the entire period. The altitude correction of 20 - 25 % suggested by Hagen & Lefauconnier (1995) is based on measured precipitation (not corrected for catch loss) and is therefore higher than values used here.

The average winter accumulation in the period 1971–1998 is 698 mm SWE. For the same period the average of the modelled winter accumulation is 634 mm, which is 9 % lower. The years 1994–96 stand out (Fig. 7). In 1994 and 1996 the modelled winter accumulation are 37 % and 48 % higher, respectively, than observed at the glacier. This is probably due to heavy rainfalls in November the previous years. On 30 November 1993 the largest precipitation event ever observed in Ny-Ålesund dropped 57 mm of rain. The average temperature was 3.3 °C. Temperatures the previous days and the following days were down to -20 °C. This created a thick and solid ice layer over most of the peninsula. A near 20 cm thick solid ice layer was observed in snowpits at lower altitudes in May the following year. A thick ice layer cannot be penetrated by normal snow sounding and it could easily be mistaken for the surface from previous year. 86 % of the observed precipitation this winter fell before this date and accumulation measurements probably underestimated true snow depth. A similar event occurred in the winter of 1995–96. On two occasions, 3 December 1995 and 12 March 1996, heavy rainfall occurred. Subsequently, the May 1996 snowpits revealed a thick ice layer. In contrast, in 1995, the year in between these years, shows the opposite case. The observed accumulation is 20 % lower than the average accumulation while the observed precipitation and simulated accumulation is 51 % and 58 % lower than average, respectively. Snow sounding above the equilibrium line is difficult because it can be hard to distinguish between snow and firn at their interface. With abnormally shallow snow depths, even an experienced observer might believe that the first hard layer encountered had formed during the winter rather than identifying it as last year's surface. Hence, a larger snow accumulation in 1995 would be estimated. If made more forcibly, the sounding would penetrate into last year's snow, stopping at the solid ice layer in the 1993–94 snow-pack.

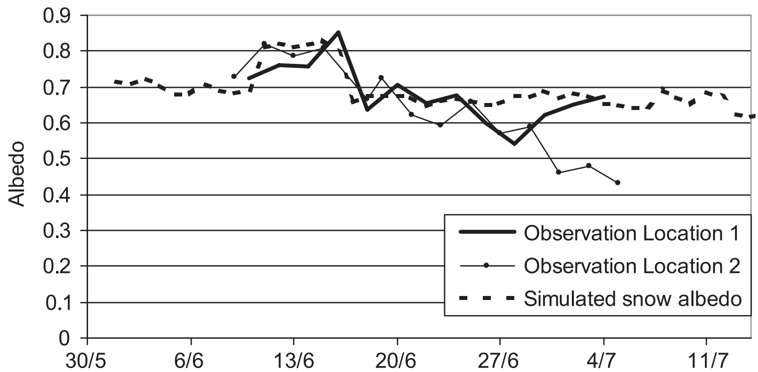

Fig. 8. Average of observed albedo values at two adjacent 100 m² large run-off plots (10 observations at each plot), compared to simulated albedo for snow during the ablation in 1998.

Hence, a larger snow accumulation would be estimated. These results show how the model could be used, retrospectively, to detect years for which observations are questionable, or in real time to obtain and ensure good measurements.

Albedo is a very important parameter in the energy balance calculations (Fig. 8). Due to a generally shallow snow cover, an uneven snow distribution and strong winds, substantial amounts of dust and organic matter from bare patches can be carried long distances by wind and deposited on the snow far from the source. Dust has been observed in snow pits at several layers. When exposed the dust significantly reduces the albedo, which is not accounted for in the model. At the higher elevations dust sources are reduced. The deviations seen at Fig.8 during the late ablation at location 2 can be explained by location of the plots. Both plots are located on the tundra below the glacier and as the snow-pack becomes shallow and translucent underlying dark soil reduces the snow albedo (Fig. 8).

The average observed summer ablation in the period 1971–1998 is 1140 mm w.e. The model calculates snow and ice melt separately; the sum of these at Austre Brøggerbreen is compared to the observed summer ablation. The modelled average ablation for the years 1971–1998 is 1330 mm w.e. or 17% higher than observed. Modelled average balance for the period was 696 mm while the observed was 442 mm. As mentioned, the model is not calibrated against the glacier mass balance but to the run-off from the entire catchment. This includes the run-off from both Austre and Vestre Brøggerbreen. Summer ablation and winter accumulation can vary over the area. The mean altitude of Austre Brøggerbreen is approximately 100 m higher than the mean of Vestre Brøggerbreen and the slope, aspect and surroundings are different. By modelling Austre Brøggerbreen separately the parameters could be further calibrated to optimize the mass balance calculations. Førland et al. (1997) found surprisingly high precipitation in the upper areas of the glaciers, concluding that spillover and "seeder" and "feeder" effects were the explanation rather than a generally high precipitation gradient. This effect, as well as snowdrift from the surroundings, will lead to the accumulation of more snow in the upper areas than the model with a linear precipitation gradient of 14%. Higher accumulation in the uppermost areas of the glacier would not affect the run-off intensities since in the model these areas are rarely free of snow and the ice surface, with lower albedo and higher melt intensity, is not exposed. The winter accumulation would, however, be improved. The map, from which the hypsographic curve of the glaciers is found, is more than 20 years old and the surface level is lower and the glaciers front has retreated since this map was created. If these changes were accounted for in the model, higher winter accumulation would probably be necessary to avoid changing the timing of the ice melt and simulated run-off. In turn this would reduce the summer ablation, but not necessarily the total run-off. This could be tested in a subsequent study.

Since the late 1960s, when mass balance observations began, Svalbard's glaciers have generally retreated. A higher average summer air temperature, less precipitation or a combination of these in the period since the "Little Ice Age", culminating just before the year 1900, might be the explanation. By manipulating the input data to the model these possible explanations can be tested. Looking at temperature alone, simulations indicate that the average summer temperature has to be 1.2 to 1.3 °C cooler on average to maintain Austre Brøggerbreen in balance with the climate for the years 1971–1998. The same result would be achieved by a 100 % increase in snow accumulation. These are substantial changes probably only possible in a completely different climatic regime than currently exists. A combination of lower air temperatures and more precipitation would to some extent give a synergistic effect since more precipitation would then fall as snow. With a 50 % increase of the precipitation the glaciers would be in balance with an only 0.3 °C lower summer temperature. Radiation, cloudiness, relative humidity, wind speed and number of days with precipitation have not been changed. It is likely that increased precipitation would increase cloud cover, reduce radiation and increase relative humidity. For example, reducing the incoming short-wave radiation by 10 % by increasing the cloudiness, but keeping both temperature and precipitation unchanged, the ablation was reduced on average by only 1.5 %.

Conclusions

Based on the available meteorological input data from Ny-Ålesund, the HBV energy balance model can satisfactorily reproduce the observed glacier mass balance at Austre Brøggerbreen glacier. The value of the parameters controlling snow accumulation and snowmelt were found through calibration and were consistent with values reported in the literature. Over the 26-year simulation period a negative net balance of 696 mm was calculated compared to the observation of 442 mm water equivalent. The simulated winter accumulation for the same period was 9 % lower and the summer ablation 17 % higher than the observed. These differences could be reduced if calibration of the mass balance calculation was given priority over calibration of the run-off. Higher accumulation rates than used in the upper areas of the glaciers would improve the winter accumulation calculation. Processes other than direct snowfall are important. Furthermore, updated maps of glacier geometry would probably improve the calculations. The glacier surface and extent has changed since the map was created. A lower altitude of the glacier would require higher winter accumulation, which in turn would reduce the summer ablation. The years of significant deviation between the modelled and observed results can be explained by observation errors resulting from thick ice layers in the snow-pack caused by events of heavy winter rain. Simulated changes on the order of -1 to -1.5 °C of the mean summer temperature or 100 % increase of the winter (snow) precipitation is necessary for Austre Brøggerbreen to be in balance with the local climate. A combined change in temperature and precipitation induces a synergistic effect and would also achieve glacier balance.

Tangborn (1999) and Jóhannesson et al. (1993) both experienced large errors when comparing the simulated interannual variation of the glacial net balance with observations. In this study, the temperature index calculations is replaced with an energy balance calculation for snow and ice melt. As an energy balance model better accounts for the meteorological conditions than a temperature index model, it is also more capable of simulating the year to year variations in the glacial net balance.

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References

Ashton, G. D. (ed.) 1986: *River and lake ice engineering*. Littleton, CO: Water Resources Publications.

Bergström, S. 1975: The development of a snow routine for the HBV-2 model. *Nord. Hydrol. 6*, 73–92.

Bengtsson, L. 1976: Snowmelt estimates from energy budget studies. *Nord. Hydrol. 7*, 3–18.

Bergström, S. 1976: *Development and application of a conceptual runoff model for Scandinavian catchments. SMHI Rep. Hydrol. 7.* Norrköping: Swedish Meteorological and Hydrological Institute.

Bergström, S. 1992: *The HBV model—its structure and applications. SMHI Rep. Hydrol. 4.* Norrköping: Swedish Meteorological and Hydrological Institute.

Bergström, S. & Forsman, A. 1973: Development of a conceptual deterministic rainfall–runoff model. *Nord. Hydrol. 16* , 89–104.

Bruland, O. & Killingtveit, Å. in press: An energy balance based HBV-model applied in an Arctic watershed at Svalbard, Spitsbergen. *Nord. Hydrol.*

Bruland, O. & Maréchal, D. 1999: *Energy and water balance* *of active layer 1991–1994. Understanding land Arctic physical properties 1996–1998. Project data report. SINTEF Rep. STF22 A98417.* Foundation for Scientific and Industrial Research, Norwegian Institute of Technology.

Bruland, O., Maréchal, D., Sand, K. & Killingtveit, Å. 2001: Energy and water balance studies of a snow cover during snowmelt period at a high Arctic site. *Theor. Appl. Climatol. 70*, 53–63.

Fountain, A. G. & Tangborn, W. V. 1985: Overview of contemporary techniques for predicting runoff from glacierized areas. *Hydrol. Sci. 149*, 27–41.

Førland E. J., Hansen-Bauer, I. & Nordli, P. Ø. 1997: *Orographic precipitation at the glacier Austre Brøggerbreen. DNMI Rep. 2/97 Klima.* Norwegian Meteorological Institute.

Haeberli, W. & Hoelzle, M. 1995: Application of inventory data for estimating characteristics of and regional climate-change effects on mountain glaciers: a pilot study with the European Alps. *Ann. Glaciol. 21*, 206–212.

Hagen, J. O. & Lefauconnier, B. 1995: Reconstructed runoff from the high Arctic basin Bayelva based on mass-balance measurements. *Nord. Hydrol. 26*, 285–296.

Hagen, J. O. & Liestøl, O. 1990: Long term glacier mass balance investigations in Svalbard 1950–1988. *Ann. Glaciol. 14*, 102–106.

Hagen, J. O. & Sætrang, A. 1991: Radio-echo soundings of sub-polar glaciers with low-frequency radar. *Polar Res. 9*, 99–107.

Hamlin, L., Pietroniro, A., Prowse, T., Soulis, R. & Kouwen, N. 1998: Application of indexed snowmelt algorithms in a northern wetland regime. *Hydrol. Process. 12*, 1641–1657.

Hansen-Bauer, I., Førland, E. J., Nordli, P. Ø. 1996: *Measured and true precipitation at Svalbard. DNMI Rep. 31/96 Klima*. Oslo: Norwegian Meteorological Institute.

Harstveit, K. 1984: *Snowmelt modelling and energy exchange between the atmosphere and a melting snow cover. Scientific Report 4.* Bergen: Geophysical Institute, University of Bergen.

Hinzman, L. D. & Kane, D. L. 1991: Snow hydrology of a headwater Arctic basin. 2: conceptual analysis and computer modelling. *Water Resour. Res. 27*, 1111–1121.

Hisdal, V. 1993: *Svalbards nature og historie. (Svalbard's nature and history.) Polarhåndb. 11.* Oslo: Norwegian Polar Institute.

Jóhannesson, T., Sigurdsson, O., Laumann, T. & Kennett, M. 1993: *Degree–day glacier mass balance modelling with application to glaciers in Iceland and Norway. Nordic Hydrological Program Rep. 33.* Reykjavík: Norwegian Hydrology Council.

Kane, D. L., Gieck, R. E. & Hinzman, L. D. 1997: Snowmelt modelling at Small Alaskan Arctic Watershed. *J. Hydrol. Eng. 2*, 204–210.

Killingtveit, Å. 1978: A new approach to snow accumulation modelling. *Proceedings of the Nordic Hydrological Conference. Papers of sessions. NHP Rep. 10.* Pp. 1–12. Otaniemi, Finland: Helsinki University of Technology, Division of Water Engineering.

Killingtveit, Å., Petterson, L. E. & Sand, K. 1994: Water balance studies at Spitsbergen, Svalbard. *Proceedings, Tenth International Northern Research Basins Symposium and Workshop, Spitsbergen, Norway. SINTEF Rep. STF22 A96415*, 77–94. Foundation for Scientific and Industrial Research, Norwegian Institute of Technology.

Killingtveit, Å. & Sælthun, N. R. 1995: *Hydrology. Hydropower Development 7.* Trondheim: Norwegian University of Technology and Science, Division of Hydraulic Engineering.

Lefauconnier, B. & Hagen, J. O. 1990: Glaciers and climate in Svalbard: statistical analysis and reconstruction of the Brøggerbreen mass balance for the last 77 years. *Ann. Glaciol. 14*, 148–152.

Lefauconnier, B., Hagen, J. O., Ørbæk, J. B., Meldvold, K. & Isaksson, E. 1999: Glacier balance trends in the Kongsfjord area, western Spitsbergen, Svalbard, in relation to the climate. *Proceedings of the International Symposium on Polar Aspects of Global Change, Tromsø, Norway, 24–28 August 1998. Polar Res. 18*, 113–400.

Lindström, G., Johansson, B., Persson, M., Gardelin, M. & Bergström, S. 1997: Development and test of the distributed HBV-96 hydrological model. *J. Hydrol. 201*, 272–288.

Male, D. H. & Gray, D. M. 1981: Snow ablation and runoff. In D. H. Male & D. M. Gray (eds.): *Handbook of snow.* Pp. 360–436. New York: Pergamon Press.

Nash, J. E. & Suthcliffe, J. V. 1970: River flow forecasting through conceptual models. Part 1: a discussion of principles. *J. Hydrol. 10*, 282–290.

Oerlemans, J., Anderson, B., Hubbard, A., Huybrechts, P., Jóhannesson, T., Knap, W. H., Schmeits, M., Stroeven, A. P., van de Wal, R. S. W., Wallinga, J. & Zuo, Z. 1998: Modelling the response of glaciers to climate warming. *Clim. Dyn. 14*, 267–274.

Partridge, G. W. & Platt, C. M. R. 1976: *Radiative processes in meteorology and climatology.* Amsterdam: Elsevier.

Paterson, W. S. B. 1994: *The physics of glaciers.* 3rd edition. Oxford: Elsevier.

Repp, K. 1979: *Breerosjon, glaciohydrology og materialtransport i et høyartktisk miljø, Brøggerbreen, Vest Spitsbergen. (Glacier erosion, glaciohydrology and material transport in a high Arctic environment, Brøggerbreen, western Spitsbergen.)* Thesis, Dept. of Geography, University of Oslo.

Sand, K. 1990: *Modelling snowmelt runoff processes on temperate and Arctic environments. IVB Report b-1990-1.* Thesis, Norwegian University of Technology and Science, Division of Hydraulic Engineering. Trondheim.

Swinbank, W. C. 1963: Longwave radiation from clear skies. *Q. J. R. Meteorol. Soc. 89*, 339–348.

Sælthun, N. R. 1996: *The "Nordic" HBV model. Description and documentation of the model version developed for the project Climate Change and Energy Production. NVE Publ. 7.* Oslo: Norwegian Water Resources and Energy Administration.

Tangborn, W. 1999: A mass balance model that uses low-altitude meteorological observations and the area-altitude distribution of a glacier. *Geogr. Ann. 81*, 753–756.

Tveit, J. & Killingtveit, Å. 1994: Snow surveys studies of water budget on Svalbard 1991–1994. *Proceedings, Tenth International Northern Research Basins Symposium and Workshop, Spitsbergen, Norway. SINTEF Rep. STF22 A96415*, 77–94. Foundation for Scientific and Industrial Research, Norwegian Institute of Technology.

US Army Corps of Engineers 1956: *Snow hydrology. Summary report of snow investigations.* Portland: Corps of Engineers, North Pacific Division.

Vehvilänen, B. 1992: *Snowcover models in operational watershed forecasting. Publications of the Water and Environment Research Institute 11.* Helsinki: National Board of Waters and the Environment.

Winther, J. G., Bruland, O., Sand, K., Killingtveit, Å. & Maréchal, D. 1997: Snow accumulation distribution on Spitsbergen, Svalbard, in 1997. *Polar Res.* 17, 155–164.