Journal of Glaciology, Vol. 46, No. 152, 2000

Sensitivity of mass balance of five Swiss glaciers to temperature changes assessed by tuning a degree-day model

Roger J. Braithwaite, Yu Zhang
School of Geography, University of Manchester, Manchester M13 9PL, England

ABSTRACT. A degree-day model is used to assess the sensitivity of the mass balance of five Swiss glaciers to temperature changes. The model uses temperature data extrapolated from nearby climate stations, and is tuned by varying precipitation to make the model fit the observed distribution of mass balance with altitude. Once the model is tuned, the effect of temperature change is simulated by recalculating the mass balance with the same parameters as before, but with a temperature increase of 1°C throughout the year. The largest mass-balance changes, involving increased ablation of > 1 m w.e. a^{-1} $°C^{-1}$, occur at the snout, with a progressively smaller increase with altitude. The area-averaged sensitivities for the five glaciers are −0.7 to −0.9 m w.e. a^{-1} $°C^{-1}$. If annual precipitation also increased by 20% it would partly offset the effect of the 1°C higher temperatures but could not compensate for it.

INTRODUCTION

In this paper we estimate the sensitivity of glacier mass balance to temperature changes using a degree-day model. The model is applied to five Swiss glaciers (Fig. 1) for which detailed mass-balance data are available. We did this work as a pilot study for a new assessment of world sea-level rise from increased melting of glaciers, following up earlier work by Meier (1984), Oerlemans and Fortuin (1992) and Kuhn (1993). Alpine glaciers are too small to have any direct effect on world sea level, but in the present paper we describe a simple methodology and test it on some well-documented glaciers; we will apply it later to larger but less known ice masses.

The present study uses the mass balances of five Swiss glaciers that have been measured over a series of years by scientists from the Swiss Federal Institute of Technology (ETH) in Zürich, i.e. Griesgletscher (1961–95), Limmerngletscher (1947–85), Plattalvagletscher (1947–85), Rhonegletscher (1979–83) and Silvrettagletscher (1959–92). These data refer to the direct glaciological method whereby the mass balance is measured as a function of altitude by a combination of ablation stakes and accumulation pits. Mass-balance data from two other Swiss glaciers are not used because they do not give the mass-balance vs altitude relation: Aletschgletscher where the hydrological and "index" stake methods are used, and Claridenfirn where the "index" stake method is used. Data for the five glaciers are summarized in the publications of the Permanent Service on the Fluctuations of Glaciers (Kasser, 1967, 1973; Müller, 1977) and the World Glacier Monitoring Service (Haeberli, 1985; Haeberli and Müller, 1988; Haeberli and Hoelzle, 1993). The most recent balance–altitude data for Griesgletscher are given by Funk and others (1997). The most detailed mass-balance study, including separate measurements of winter and summer balances, is for Rhonegletscher (Funk, 1985), while only annual balances are available for the other glaciers.

For glaciological purposes, the best climate stations are at relatively high elevations, where essentially the same climatic conditions are sampled as those at the glaciers. For present purposes, the stations at Gütsch ob Andermatt and Weissfluhjoch (Table 1; Fig. 1) are relatively close to the five glaciers, and their data are used in the present study. The data for these and other Swiss stations are given in annual publications (*Annalen der Schweizerischen Meteorologischen Zentralanstalt*) from the Swiss Meteorological Office in Zürich. Climate obser-

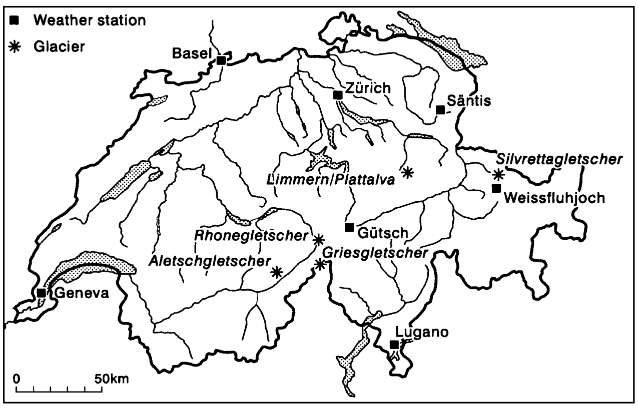

Fig. 1. Locations of Swiss glaciers with long mass-balance records, together with nearby high-elevation weather stations.

Table 1. Locations of five Swiss glaciers with mass-balance records, together with locations of two high-lying climate stations

Glacier	*Lat.* (N)	*Long.* (E)	*Climate station*	*Lat.* (N)	*Long.* (E)	*Altitude* m a.s.l.
Gries	46°26′	8°20′	Gütsch ob Andermatt	46°39′	8°37′	2287
Limmern	46°49′	8°59′	Gütsch ob Andermatt	46°39′	8°37′	2287
Plattalva	46°50′	8°59′	Gütsch ob Andermatt	46°39′	8°37′	2287
Rhone	46°37′	8°24′	Gütsch ob Andermatt	46°39′	8°37′	2287
Silvretta	46°51′	10°05′	Weissfluhjoch	46°50′	9°49′	2690

https://doi.org/10.3189/172756500781833511 Published online by Cambridge University Press

vations have also been made for more than a century at Säntis at 2500 m a.s.l. just to the northeast of the study area (Fig. 1).

The simplest, and possibly most obvious, way to link mass-balance data to climate is by correlation of mass-balance measurements with climate data. For example, Liestøl (1967), Martin (1975), Braithwaite (1977), Laumann and Tvede (1989), Chen and Funk (1990) and Müller-Lemans and others (1995) have all calculated multiple regression models for mean specific mass-balance data in terms of summer mean temperature and annual precipitation at some suitable nearby climate station. According to this approach, the sensitivity of the mass balance to temperature change would be represented by the temperature coefficient in the regression equation. Although we prefer to calculate the mass-balance sensitivity with the degree-day model, we first calculate multiple regression models so that we can compare the two approaches.

REGRESSION MODELS

The first problem of the regression approach is that it is difficult to prescribe the length of summer for temperature averages, or the choice of hydrological year for the annual precipitation. This point is illustrated by correlating the mean specific mass balance of each glacier with mean temperatures for different periods, i.e. T_7 for July temperatures only, T_{7-8} for the July–August average, T_{6-8} for June–August, T_{6-9} for June–September, T_{5-9} for May–September and T_{5-10} for May–October.

There are only 3 years of data for Rhonegletscher, so no regression models are calculated for that glacier. The highest correlations for the other four glaciers (Table 2) appear for T_{6-8}, suggesting that the June–August period (92 days) is the best choice for the summer period for the whole glacier. This seems a little too short, however. For example, the equilibrium-line altitude (ELA) is generally regarded as the most representative altitude on the glacier (Ohmura and others, 1992), and extrapolation of climate-station temperatures to the ELAs of the four glaciers suggests a longer melting season at this altitude: from 112 days at Griesgletscher to 147 days at Silvrettagletscher. There is also a hint in Table 2 of a secondary maximum correlation coefficient for mass balance and temperature for May–September (153 days), but this seems much too long a period to represent the melting season at the ELA.

The mean specific mass balances were also correlated with annual precipitation for different choices of hydrological year: P_{8-7} for the total precipitation from August in one calendar year to July in the following year, P_{9-8} for September–August, P_{10-9} for October–September and P_{11-10} for November–October. The correlations between mass balance and annual precipitation (Table 3) are generally weaker than those for temperature. There is a very slight maximum for correlations between mass balance and P_{9-8}, but other choices of hydrological year cannot be precluded.

Table 2. Correlations between mean specific mass balance and summer mean temperature, assuming various lengths of summer

Glacier	N	T_7	T_{7-8}	T_{6-8}	T_{6-9}	T_{5-9}	T_{5-10}
Gries	33	−0.43	−0.58	−0.70	−0.66	−0.70	−0.59
Limmern	32	−0.52	−0.60	−0.72	−0.57	−0.64	−0.47
Plattalva	32	−0.42	−0.51	−0.67	−0.51	−0.61	−0.45
Rhone	4	–	–	–	–	–	–
Silvretta	34	−0.62	−0.65	−0.71	−0.55	−0.58	−0.51
Mean		−0.50	−0.59	−0.70	−0.57	−0.63	−0.51

Table 3. Correlations between mean specific mass balance and annual precipitation for various choices of hydrological year

Glacier	N	P_{8-7}	P_{9-8}	P_{10-9}	P_{11-10}
Gries	33	0.45	0.49	0.52	0.51
Limmern	31	0.59	0.63	0.58	0.52
Plattalvag	31	0.62	0.65	0.61	0.55
Rhone	4	–	–	–	–
Silvretta	34	0.37	0.40	0.39	0.35
Mean		0.51	0.54	0.53	0.48

The patterns of correlations in Tables 2 and 3 suggest the following multiple regression model:

$$\underline{b} = \sigma + \beta T_{6-8} + \gamma P_{9-8}, \tag{1}$$

where $\underline{b}$ is the mean specific balance of the glacier, α is the intercept in the multiple regression equation and β and γ are regression coefficients for summer temperature (June–August) and annual precipitation (September–August), respectively. The intercept and regression coefficients in Equation (1) are readily determined for a matrix of mass-balance, temperature and precipitation data using a commercial personal computer data package.

MASS BALANCE AND DEGREE-DAY MODEL

The model used here is based on that developed by Braithwaite (1977, 1980, 1985) and used by Braithwaite and Thomsen (1989). The earlier model attempted to take account of the refreezing of meltwater and rainfall in subpolar glaciers, which is assumed to be negligible for the temperate glaciers considered in this paper.

The mass balance of the glacier is characterized by observed mass-balance values at regular altitude intervals on the glacier, i.e. 100 m intervals. The observed balance at the jth altitude is b_j, and the corresponding mean specific balance for the whole glacier is $\underline{b}$ defined by:

$$\underline{b} = (1/S) \sum_{j=1}^{j=J} s_j b_j, \tag{2}$$

where s_j is the area of the jth altitude band and S is the total area of the glacier. The corresponding modelled balances are $b_j{}^*$ and $\underline{b}^*$, respectively. The modelled balance at jth altitude is given by:

$$b_j{}^* = c_j{}^* - a_j{}^*, \tag{3}$$

where $c_j{}^*$ and $a_j{}^*$ are the annual accumulation and annual ablation at the jth altitude in the model.

The annual ablation a^* is generally made up of a sum of ice ablation $a_\mathrm{i}{}^*$ and snow ablation $a_\mathrm{s}{}^*$, and is calculated by the degree-day model (Braithwaite and Olesen, 1989):

$$a_j{}^* = k_\mathrm{i}\mathrm{PDD_i} + k_\mathrm{s}\mathrm{PDD_s}, \tag{4}$$

where $\mathrm{PDD_i}$ and $\mathrm{PDD_s}$ are the annual positive degree-day sums (sum of positive air temperatures measured at screen height above the glacier) at the altitude in question for the periods of ice melt and snowmelt, respectively. The parameters k_i and k_s are the degree-day factors for ice melt and snowmelt. (To avoid a cluttered notation, the subscript j is implicit for PDD variables.)

https://doi.org/10.3189/172756500781833511 Published online by Cambridge University Press

The degree-day model has already been used to calculate the sea-level rise from the increased melting of the Greenland ice sheet (Huybrechts and others, 1991) and can be applied to other glacier areas if the appropriate k values are known. There is ample evidence from both fieldwork and modelling studies that snowmelt generally has a lower degree-day factor than ice melt, but there is otherwise no evidence of any single set of universal factors that can be applied to all situations. A summary of positive degree-day factors from the literature is given in Table 4, together with values for ice ablation on Griesgletscher (at 2510 and 2560 m a.s.l.) measured by the authors in July–August 1996 and August 1997.

The reasons for the different degree-day factors in Table 4 are not immediately obvious. The three highest values in Table 4 for Spitsbergen and Greenland reflect rather cold low-ablation situations where higher degree-day factors may be expected from energy-balance considerations (Braithwaite, 1995b). The greater altitude of Griesgletscher compared with the Norwegian and Greenland glaciers in Table 4 would suggest that Griesgletscher ought to have lower turbulent fluxes, and therefore a lower degree-day factor for ice (Braithwaite, 1995a), in agreement with, for example, Kasser (1959), but this is contradicted by the field data.

Table 4. Positive degree-day factors at various locations. Units are mm d^{-1} °C^{-1}

Ice	*Snow*	*Location*	*Source*
	4.5	Weissfluhjoch, Switzerland	Zingg (1951)
5.0–7.0		Various Swiss glaciers	Kasser (1959)
18.6		EGIG* Camp IV, Greenland	Ambach (1963, 1988)
13.8		Spitsbergen	Schytt (1964)
6.3		St. Supphellebreen, Norway	Orheim (1970)
	5.5	Gr. Aletschgletscher, Switzerland	Lang and others (1977)
5.5 ± 2.3		Various glaciers, Norway	Braithwaite (1977)
	4.5	Weissfluhjoch, Switzerland	De Quervain (1979)
6.3 ± 1.0		Arctic Canada	Braithwaite (1981)
20.1–22.2		GIMEX† profile, West Greenland	Van de Wal (1992)
6.0	3.0	Franz Josef Glacier, New Zealand	Woo and Fitzharris (1992)
7.7	5.7	Sátujökull, Iceland	Jóhannesson and others (1993)
6.4	4.4	Nigardsbreen, Norway	Jóhannesson and others (1993)
7.3	2.8	Qamanârssûp sermia, West Greenland	Jóhannesson and others (1993)
6.0	4.5	Ålfotbreen, Norway	Laumann and Reeh (1993)
5.5	4.0	Nigardsbreen, Norway	Laumann and Reeh (1993)
5.5	3.5	Hellstugubreen, Norway	Laumann and Reeh (1993)
8.1	2.9	Nordbogletscher, West Greenland	Braithwaite (1995b)
8.3	3.7	Qamanârssûp sermia, West Greenland	Braithwaite (1995b)
6.9–7.1		Patagonia, Argentina	Takeuchi and others (1996)
5.9–9.8		North Greenland	Braithwaite and others (1998)
6.2	3.8	Glacier de Sarennes, France	Vincent and Vallon (1997)
8.3–9.4		Griesgletscher, Switzerland	This study
6.3	4.4	Storglaciren, Sweden	Hock (1999)

* Expédition glaciologique internationale au Groenland.

† Greenland Ice Margin Experiment (Universities of Utrecht and Amsterdam).

The modelled balance is then given by:

$$b_j^* = c_j^* - k_i\mathrm{PDD_i} - k_s\mathrm{PDD_i} - k_s\mathrm{PDD_s}. \qquad (5)$$

The degree-day sum for the snowmelt period $\mathrm{PDD_s}$ is c_j^*/k_s, so the degree-day sum for the ice-melt period is given by:

$$\mathrm{PDD_i} = \mathrm{PDD} - c_j^*/k_s \qquad \mathrm{PDD} > c_j^*/k_s \qquad (6a)$$

and

$$\mathrm{PDD_i} = 0 \qquad \mathrm{PDD} \leq c_j^*/k_s, \qquad (6b)$$

where PDD is the annual positive degree-day total equal to $\mathrm{PDD_i} + \mathrm{PDD_s}$. Equation (6a) applies to the ablation area, and Equation (6b) to the accumulation area.

Snow accumulation at any particular altitude is estimated for each month from monthly mean temperature and monthly precipitation by assuming that precipitation is split between rain and snow according to the probability that air temperatures within the month lie above or below 0°C. This probability is estimated from monthly mean temperature (Fig. 2) by assuming that temperatures within the month are distributed according to the normal distribution with standard deviation $\sigma = 4$°C (Braithwaite, 1985). The estimation of accumulation here implicitly assumes that (1) precipitation falls as snow if air temperature is below 0°C, and (2) the probability of precipitation occurring is independent of temperature. Neither of these assumptions is exactly correct, but they may compensate (e.g. snow often falls when surface temperatures are above the freezing point, but precipitation in the Alps also generally occurs in the colder part of a summer month).

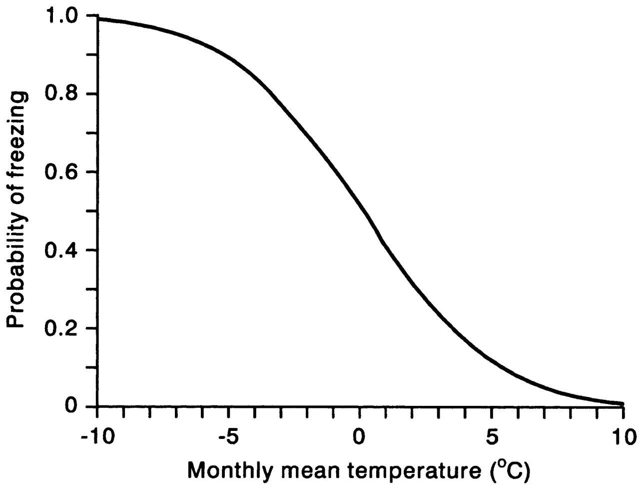

Fig. 2. Probability of freezing vs monthly mean temperature assuming that temperatures are normally distributed within the month, with $\sigma = 4$°C.

Rain in the model is assumed to run off the glacier and not to contribute to the mass balance, and the annual snow accumulation c_j^* in the model is found by summing the calculated accumulation for each month.

The degree-day total is calculated for each month from the monthly mean temperature according to the Braithwaite (1985) model which assumes that temperatures are normally distributed within the month. Annual degree-day total is then found by summing the monthly totals, and the mass balance is obtained from Equations (5), (6a) and (6b) using the appropriate degree-day factors. For example, degree-day factors of 8 and 4.5 mm d^{-1} °C^{-1} are used for ice and snow ablation, respectively, in Figure 3. The former is

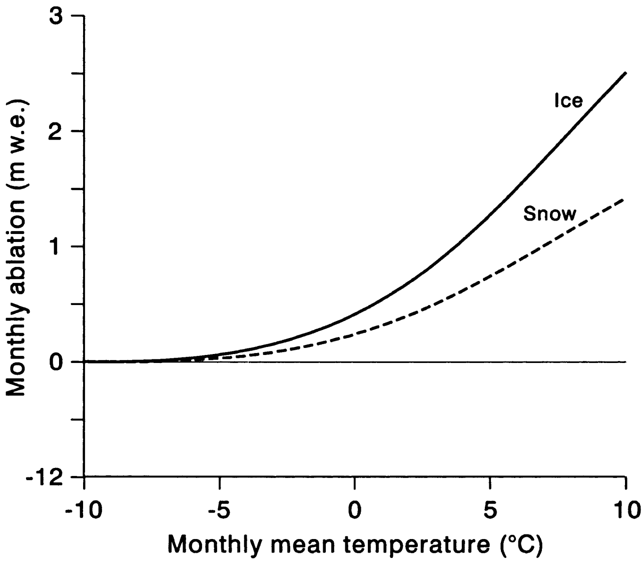

Fig. 3. Monthly ablation for snow and ice vs monthly mean temperatures assuming that temperatures are normally distributed within the month, with $\sigma = 4°C$. Degree-day factors for ice and snow are taken to be 8.0 and 4.5 mm d^{-1} °C^{-1}, respectively.

close to that found for ice in Greenland (Braithwaite and Olesen, 1989) and on Griesgletscher, while the latter is reported for seasonal snow in Switzerland (Zingg, 1951; De Quervain, 1979). Whenever other values of the degree-day factor for ice were used in the present study, the same ratio of snow to ice factors, i.e. 4.5/8.0 = 0.6, was maintained.

The monthly mean air temperatures for the accumulation and ablation calculations are estimated at each altitude by extrapolating from a suitable weather station below the glacier using a standard lapse rate (e.g. 0.006–0.008°C m^{-1}).

MODEL TUNING

As the annual precipitation for each glacier is not well known, precipitation is treated as a tuning variable in the model and varied to make the model fit the observed data. The time distribution of precipitation over the glacier is assumed to be the same as at the climate station, although partition between rain and snow varies according to temperature. For Griesgletscher it is sufficient to use the same precipitation for all altitudes, although the vertical lapse of temperature ensures an increase of accumulation with altitude.

At the beginning of the study, which started with Griesgletscher, the temperature lapse rate and degree-day factor for ice melt were also regarded as uncertain, and model calculations were made for a range of values. In all, 96 different runs of the model were made for three values of lapse rate (0.006–0.008°C m^{-1}), two values of degree-day factor (8 or 9 mm d^{-1} °C^{-1}) and 16 values of precipitation (1.0–2.5 m w.e. a^{-1} in intervals of 0.1 m w.e. a^{-1}). The accuracy of any model run is expressed by the error $\underline{e}$ between mean specific values of modelled balance $\underline{b}^*$ and observed balance $\underline{b}$.

$$\underline{e} = \underline{b}^* - \underline{b} = (1/S)\sum_{j=1}^{j=J} s_j\left(b_j^* - b_j\right). \qquad (7)$$

For many of the model runs, there is little resemblance between the modelled and observed mass balance, but a

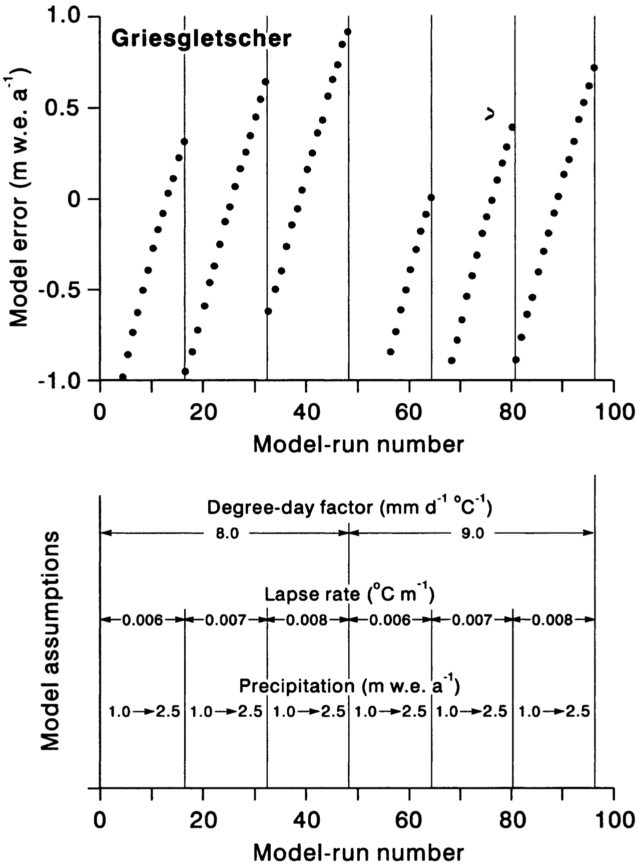

Fig. 4. Model errors and assumptions for 96 runs of the degree-day model for Griesgletscher.

number of runs appear to give a reasonable fit between model and data. For example, the model error in Figure 4 is small for a number of different combinations of the model variables. This shows that tuned models are not unique. In the present approach, increased precipitation can usually be offset by increasing the degree-day factor to melt more snow, or by reducing the lapse rate to give higher temperature (and ablation) at the altitude in question. Presumably, more sophisticated models like the energy-balance model of Oerlemans and Fortuin (1992) will also suffer from this problem of non-uniqueness.

From a visual examination of the balance–altitude relations for the different model runs, the results of model run 26 appear to give the best overall agreement between model and data for Griesgletscher. The parameters for this model are annual precipitation = 1.9 m w.e. a^{-1}, temperature lapse rate = 0.007°C m^{-1} and degree-day factor for ice = 8.0 mm d^{-1} °C^{-1}, which are all plausible values. However, it is clear from Figure 5 that the model cannot exactly reproduce the observed data, which show a distinct "kink" between about 2950 and 3150 m a.s.l. Presumably this is caused by a local precipitation anomaly. The results of model 92 are also shown in Figure 5, to illustrate the fact that some models can fit the data very well in one altitude range (near the snout in the present case) whilst deviating greatly elsewhere (in the accumulation area).

The tuning is repeated in a similar way for the other four Swiss glaciers: Limmerngletscher, Plattalvagletscher, Rhonegletscher and Silvrettagletscher. We became confident enough to prescribe the lapse rate and degree-day factor for ice, with values of 0.007°C m^{-1} and 8.0 mm d^{-1} °C^{-1}, respectively, rather than treat them as parameters. For the first three glaciers it was also found necessary to vary pre-

https://doi.org/10.3189/172756500781833511 Published online by Cambridge University Press

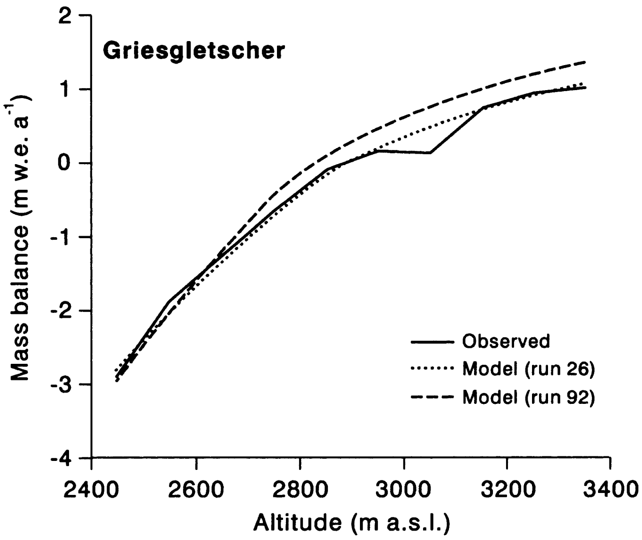

Fig. 5. Observed and modelled mass balances vs altitude for Griesgletscher.

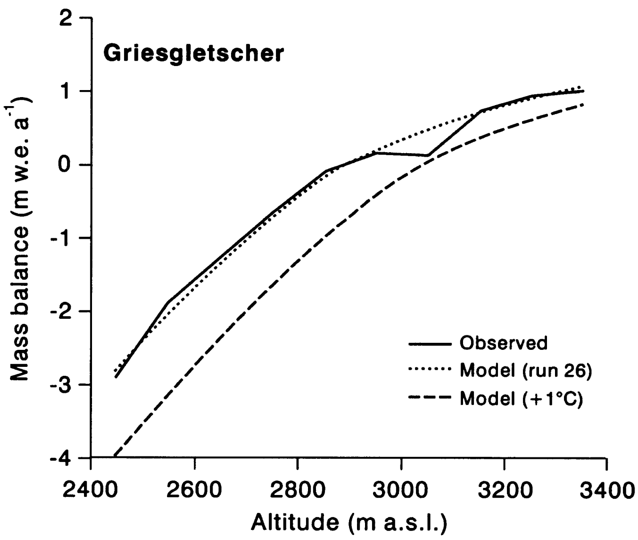

Fig. 6. Modelled mass balance for Griesgletscher for present climate and for a +1°C temperature rise.

cipitation with altitude to fit the balance–altitude models to data. For Limmerngletscher the modelling indicated particularly heavy accumulation on the snout, while Rhonegletscher needs a strong precipitation increase with altitude to reproduce its vertical mass-balance curve. In the case of Rhonegletscher, the modelled accumulation of 2.09 m w.e. a^{-1} can be compared with the measured winter balance of 1.89 m w.e. a^{-1} for the 4 years of record. In the nature of things, annual accumulation (whole year) must be somewhat higher than winter balance (net balance for part of the year), so the modelled and observed data are in reasonable agreement.

The modelled and observed averages for mean specific balance are given in Table 5 for the five glaciers. The periods of record are incomplete in a couple of cases because balance–altitude data could not be found for all years. In all cases, however, the model uses temperature data for the same period as covered by the mass-balance data.) The tuning process naturally involves reducing the error between model and data, but over-tuning was avoided. For example, model run 26 was judged to give the best overall agreement with observed altitude distribution of mass balance for Griesgletscher, although it did not give the lowest error for mean specific balance.

EFFECT OF TEMPERATURE CHANGE

Once the mass-balance model has been satisfactorily tuned, the effect of temperature change is simulated by recalculating the mass balance with the same parameters as before,

Table 5. Comparison of modelled and observed mean specific balance for five Swiss glaciers. Figures are averages for the given periods

Glacier	*Period*	*Modelled* m w.e. a^{-1}	*Observed* m w.e. a^{-1}	*Error* m w.e. a^{-1}
Gries	1961–90	−0.31	−0.38	+0.07
Limmern	1976–85	−0.04	−0.02	−0.02
Plattalva	1976–85	+0.28	+0.28	0.00
Rhone	1980–82	+0.20	+0.28	0.00
Silvretta	1960–90	+0.07	+0.08	−0.01

but with a changed temperature. In the present paper we are only concerned with a uniform change of 1°C throughout the year. This is purely an illustration, and different temperature changes can easily be applied to different months when plausible climate-change scenarios become available for the Swiss Alps.

The effect of a 1°C temperature change on the mass balance of Griesgletscher is illustrated in Figure 6 for model run 26. It is noteworthy that the largest change, involving an increased ablation of > 1 m w.e. a^{-1}, occurs at the snout, with a progressively smaller increase with greater altitude on Griesgletscher, in agreement with Oerlemans and Hoogendoorn (1989), Jóhannesson and others (1993) and Laumann and Reeh (1993). There are probably two reasons for this: (1) variations in ablation increase with the magnitude of the ablation itself, and (2) snow ablation, with a lower degree-day factor than ice ablation, becomes more common with increasing altitude. The first point is illustrated by the ablation curves in Figure 3 where the ablation–temperature gradients increase as ablation increases.

The overall change in the mass balance of Griesgletscher as a result of the increased temperature, i.e. the sensitivity of the mean specific mass balance to temperature change, is −0.69 m w.e. a^{-1} °C^{-1} for model run 26. This represents an area average of the balance changes at different altitudes. The same procedure can also be applied to the other model runs that give a reasonable fit to the data (e.g. the 11 cases in Figure 4 where the model error is less than ±0.1 m w.e. a^{-1}, and a range of sensitivities results. The sensitivity is strongly correlated with the annual precipitation (Fig. 7), in agreement with Oerlemans and Fortuin (1992) who proposed an explicit functional relation between mass-balance sensitivity and precipitation. A more obvious explanation is that where there is high precipitation there is also high ablation and therefore greater negative sensitivity.

The range of sensitivities in Figure 7 casts a new light on the problem of uniqueness for tuned models. In this case, we have a number of model runs with a similarly small error, as shown in Figure 4, and it may seem unimportant which one we choose, but they do have different sensitivities.

The mass-balance sensitivities calculated for the five glaciers by the degree-day model are considerably larger than those for the regression model (Table 6). This may be partly a statistical artefact, i.e. regression models "smooth"

https://doi.org/10.3189/172756500781833511 Published online by Cambridge University Press

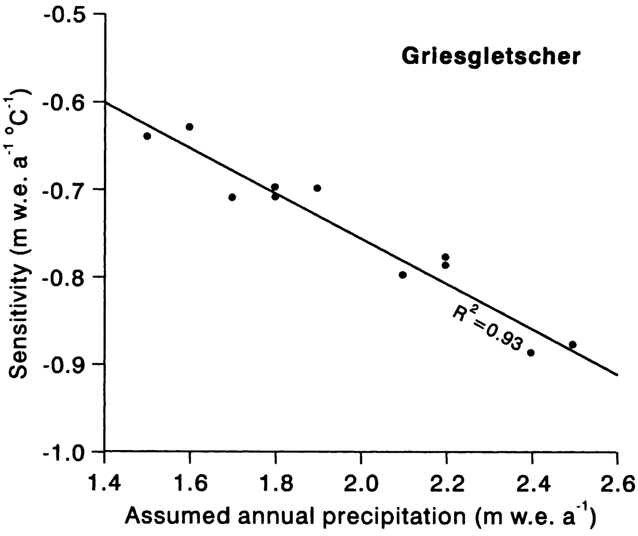

Fig. 7. Sensitivity vs assumed annual precipitation for Griesgletscher.

the processes that they purport to describe, but it may also be because the degree-day model is physically more realistic. For example, in the degree-day model, higher temperature will explicitly reduce the model accumulation (by reducing the proportion of precipitation falling as snow) even if the precipitation remains the same, while this effect can only be implicit, at best, in the temperature coefficient in the multiple regression model. Also, the regression model cannot distinguish between ice- and snowmelting (Vincent and Vallon, 1997).

Table 6. Changes in mass balance of five Swiss glaciers due to a 1°C rise in temperature, estimated by two different models

Glacier	*Regression model* m w.e. a^{-1} °C^{-1}	*Degree-day model* m w.e. a^{-1} °C^{-1}
Gries	-0.44 ± 0.09	−0.69
Limmern	-0.53 ± 0.09	−0.79
Plattalva	-0.50 ± 0.10	−0.82
Rhone	–*	−0.68
Silvretta	-0.64 ± 0.10	−0.89
Mean	−0.53	−0.77

* Not enough data.

More fundamentally, the regression model explicitly assumes a linear relation between mean specific balance and temperature, while the degree-day model implies no such relation. The mass-balance sensitivity quoted here refers to the mass-balance change for the first increase of 1°C because this is the convention in discussions of glaciers and sea level (e.g. Meier, 1984; Oerlemans and Fortuin, 1992; Kuhn, 1993), although most workers agree that the relation between mass balance and mean temperature must be somewhat non-linear.

DISCUSSION

The temperature-change experiments (above) were carried out holding precipitation constant, but it was found that model accumulation fell by 5–8% with a temperature rise of 1°C. This reflects the reduced proportion of snowfall at higher temperatures in the model. As future climate changes could involve changes in precipitation as well as temperature, a further experiment was made for Griesgletscher (model run 26) where precipitation was increased or decreased by 20% to compare with the effect of raising or lowering temperatures by 1°C. The results (Table 7) show that increased precipitation would partly offset the effect of higher temperatures, but even a 20% increase in precipitation cannot compensate for the increased ablation due to a 1°C temperature rise. By contrast, decreased precipitation would substantially reinforce the effects of increased temperature.

Table 7. Calculated changes in mass balance of Griesgletscher due to temperature and/or precipitation changes. Units of balance change are m w.e. a^{-1}

Precipitation change	*Temperature change* −1°C	*Present*	*+1°C*
+20%	+0.91	+0.37	−0.30
Present	+0.55	0.00	−0.69
−20%	+0.18	−0.39	−1.13

A small experiment was performed by rerunning the model for Griesgletscher (model run 26) and varying the ratio of degree-day factors k_s/k_i from the standard value of 0.6 for this study to a value of 1.0. The result is generally to increase ablation by >0.5 m w.e. a^{-1} at the glacier snout, and to straighten the balance–altitude curve somewhat. Insofar as it would be more difficult to fit these straighter curves to the observed data, the use of differing degree-day factors for ice and snow in this and other studies (Huybrechts and others, 1991; Jóhannesson and others, 1993; Laumann and Reeh, 1993; Vincent and Vallon, 1997) seems justified. This means that the degree-day approach is more realistic than the power-law model of Krenke and Khodakov (1966), recently resurrected by Davidovich and Ananicheva (1996) and Pfeffer and others (1997), because it distinguishes between ice and snow. The latter model is obviously based upon a very extensive dataset from the former Soviet Union, and we urge our Russian colleagues to re-analyze their data in terms of the differing ablation properties of ice and snow.

The relation between temperature sensitivity and altitude is shown for all five glaciers in Figure 8, supporting the earlier finding for Griesgletscher. It is interesting that the five glaciers have roughly similar sensitivities for the same altitudes. The strong relation between sensitivity and altitude suggests that climate-change experiments should be performed for individual altitude bands rather than for whole glaciers. For example, as the tongue of a glacier, with the highest sensitivity, is melted away, the average sensitivity of the remaining part of the glacier must be reduced.

The sensitivities found for the five glaciers, i.e. −0.7 to −0.9 m w.e. a^{-1} °C^{-1} in round figures, probably represent an intermediate range in global terms, with lower values for subpolar and high-altitude glaciers and higher values for very maritime glaciers. From an energy-balance study of 12 glaciers from different regions, Oerlemans and Fortuin (1992) suggested a global sensitivity of −0.4 m w.e. a^{-1} °C^{-1}, which would give a world sea-level rise of 0.6 mm a^{-1} °C^{-1}. As the latter figure is only half that given in the 1990 Intergovernmental Panel on Climate Change report (Warrick and Oerlemans, 1990), it is important to confirm or refute this lower figure by further work.

https://doi.org/10.3189/172756500781833511 Published online by Cambridge University Press

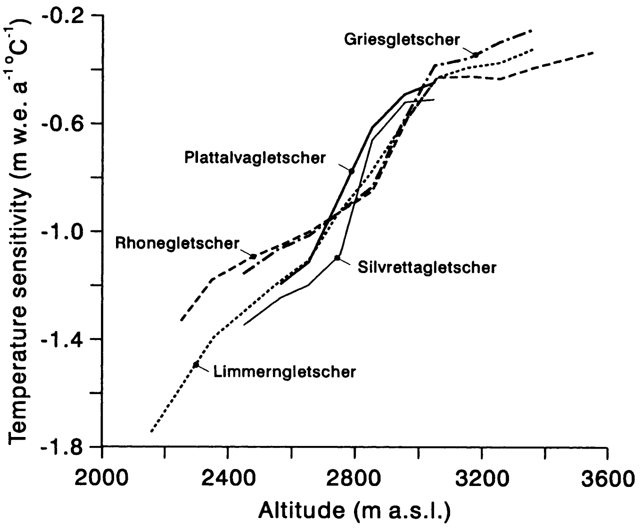

Fig. 8. Temperature sensitivity vs altitude for five Swiss glaciers.

CONCLUSIONS AND OUTLOOK

The degree-day model can be tuned to fit the mass balance of five Swiss glaciers by varying precipitation in the model and assuming a temperature lapse rate of 0.007°C m^{-1} and a degree-day factor for melting ice of 8.0 mm d^{-1} $°C^{-1}$. A 1°C temperature rise in the model gives an increased ablation of >1 m w.e. a^{-1} at the glacier snouts, with a progressively smaller increase with greater altitude. A 20% precipitation increase could partly offset the increased ablation caused by the 1°C temperature rise but could not compensate for it.

The area-averaged sensitivities for the five Swiss glaciers are –0.7 to –0.9 m w.e. a^{-1} $°C^{-1}$, probably representing an intermediate range in global terms, i.e. higher than sensitivities for subpolar glaciers and lower than for maritime glaciers.

ACKNOWLEDGEMENTS

The study is funded partly by the U.K. Natural Environment Research Council (grant No. GR9/01777 to the University of Manchester) and partly by the Climate and Environment Programme, European Union (grant No. ENV4-CT95-0124, co-ordinated by the Climate Research Unit, University of East Anglia, U.K.). M. Funk of ETH provided invaluable advice. R. Evans (Manchester Metropolitan University), R. Bassford (University of Manchester), J. Hancock (Manchester Grammar School) and the Nanzer family (Hotel Furka in Oberwald, Kanton Wallis) helped us with fieldwork on Griesgletscher in summers 1996 and 1997.

REFERENCES

Ambach, W. 1963. Untersuchungen zum Energieumsatz in der Ablationszone des grönländischen Inlandeises (Camp IV–EGIG, 69°40′05″ N, 49°37′58″ W). *Medd. Grønl.*, **174**(4).

Ambach, W. 1988. Heat balance characteristics and ice ablation, western EGIG-profile, Greenland. *In* Thomsen, T., H. Sögaard and R. J. Braithwaite, *eds. Applied hydrology in the development of northern basins.* Copenhagen, Danish Society for Arctic Technology, 59–70.

Braithwaite, R. J. 1977. Air temperature and glacier ablation — a parametric approach. (Ph.D. thesis, Interdisciplinary Studies in Glaciology, McGill University.)

Braithwaite, R. J. 1980. Regional modelling of ablation in West Greenland. *Grønlands Geol. Undersøgelse, Rapp.* 98.

Braithwaite, R. J. 1981. On glacier energy balance, ablation, and air temperature. *J. Glaciol.*, **27**(97), 381–391.

Braithwaite, R. J. 1985. Calculation of degree-days for glacier-climate research. *Z. Gletscherkd. Glazialgeol.*, **20**, 1984, 1–8.

Braithwaite, R. J. 1995a. Aerodynamic stability and turbulent sensible-heat flux over a melting ice surface, the Greenland ice sheet. *J. Glaciol.*, **41**(139), 562–571.

Braithwaite, R. J. 1995b. Positive degree-day factors for ablation on the Greenland ice sheet studied by energy-balance modelling. *J. Glaciol.*, **41**(137), 153–160.

Braithwaite, R. J. and O. B. Olesen. 1989. Calculation of glacier ablation from air temperature, West Greenland. *In* Oerlemans, J., *ed. Glacier fluctuations and climatic change.* Dordrecht, etc., Kluwer Academic Publishers, 219–233.

Braithwaite, R. J. and H. H. Thomsen. 1989. Simulation of run-off from the Greenland ice sheet for planning hydro-electric power, Ilulissat/Jakobshavn, West Greenland. *Ann. Glaciol.*, **13**, 12–15.

Braithwaite, R. J., T. Konzelmann, C. Marty and O. B. Olesen. 1998. Reconnaisance study of glacier energy balance in North Greenland, 1993–94. *J. Glaciol.*, **44**(147), 239–247.

Chen, J. and M. Funk. 1990. Mass balance of Rhonegletscher during 1882/83–1986/87. *J. Glaciol.*, **36**(123), 199–209.

Davidovich, N.V. and M. D. Ananicheva. 1996. Prediction of possible changes in glacio-hydrological characteristics under global warming: southeastern Alaska, U.S.A. *J. Glaciol.*, **42**(142), 407–412.

De Quervain, M. 1979. Schneedeckenablation und Gradtage im Versuchsfeld Weissfluhjoch. *Eidg. Tech. Hochschule, Zürich. Versuchsanst. Wasserbau, Hydrol. Glaziol. Mitt.* 41, 215–232.

Funk, M. 1985. Räumliche Verteilung der Massenbilanz auf dem Rhonegletscher und ihre Beziehung zu Klimaelementen. *Zürcher Geogr. Schr.* 24.

Funk, M., R. Morelli and W. Stahel. 1997. Mass balance of Griesgletscher 1961–1994: different methods of determination. *Z. Gletscherkd. Glazialgeol.*, **33**(1), 1996, 41–55.

Haeberli, W., *comp.* 1985. *Fluctuations of glaciers 1975–1980 (Vol. IV).* Paris, International Commission on Snow and Ice of the International Association of Hydrological Sciences/UNESCO.

Haeberli, W. and M. Hoelzle, *comps.* 1993. *Fluctuations of glaciers 1985–1990 (Vol. VI).* Wallingford, Oxon, IAHS Press; Nairobi, UNEP; Paris, UNESCO.

Haeberli, W. and P. Müller, *comps.* 1988. *Fluctuations of glaciers 1980–1985 (Vol. V).* Wallingford, Oxon, IAHS Press; Nairobi, UNEP; Paris, UNESCO.

Hock, R. 1999. A distributed temperature-index ice- and snowmelt model including potential direct solar radiation. *J. Glaciol.*, **45**(149), 101–111.

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

Jóhannesson, T., O. Sigurdsson, T. Laumann and M. Kennett. 1993. *Degree-day glacier mass balance modelling with applications to glaciers in Iceland and Norway.* Reykjavík, Orkustofnun. (Nordic Hydrological Programme. Rapport 33.)

Kasser, P. 1959. Der Einfluss von Gletscherrückgang und Gletschervorstoss auf den Wasserhaushalt. *Wasser- und Energiewirtschaft*, **51**(6), 155–168.

Kasser, P. 1967. *Fluctuations of glaciers 1959–1965 [Vol. I].* Paris, International Commission of Snow and Ice of the International Association of Scientific Hydrology/UNESCO.

Kasser, P., *comp.* 1973. *Fluctuations of glaciers 1965–1970 [Vol. II].* Paris, International Commission on Snow and Ice of the International Association of Hydrological Sciences/UNESCO.

Krenke, A. N. and V. G. Khodakov. 1966. O svyazi poverkhnostnogo tayaniya lednikov s temperaturoy vozdukha [The relationship between surface ice melting and air temperature]. *Mater. Glyatsiol. Issled.* 12, 153–164.

Kuhn, M. 1993. Possible future contributions to sea level change from small glaciers. *In* Warrick, R. A., E. M. Barrow and T. M. L. Wigley, *eds. Climate and sea level change: observations, projections and implications.* Cambridge, Cambridge University Press, 134–143.

Lang, H., B. Schädler and G. Davidson. 1977. Hydroglaciological investigations on the Ewigschneefeld — Gr. Aletschgletscher: ablation, meltwater infiltration, water table in firn, heat balance. *Z. Gletscherkd. Glazialgeol.*, **12**(2), 1976, 109–124.

Laumann, T. and N. Reeh. 1993. Sensitivity to climate change of the mass balance of glaciers in southern Norway. *J. Glaciol.*, **39**(133), 656–665.

Laumann, T. and A. M. Tvede. 1989. Simulation of the effects of climate changes on a glacier in western Norway. *In* Huttunen, L., *ed. Conference on Climate and Water. Helsinki, Finland, 11–15 September 1989. Proceedings. Vol. 1.* Helsinki, Suomen Akatemia / Academy of Finland, 339–352. (Suomen Akatemian Julkaisuja / Publications of the Academy of Finland 9/89.)

Liestøl, O. 1967. Storbreen glacier in Jotunheimen, Norway. *Nor. Polarinst. Skr.* 141.

Martin, S. 1975. Corrélation bilans de masse annuels — facteurs météorologiques dans les Grandes Rousses. *Z. Gletscherkd. Glazialgeol.*, **10**(1–2), 1974, 89–100.

Meier, M. F. 1984. Contribution of small glaciers to global sea level. *Science*, **226**(4681), 1418–1421.

Müller, F., *comp.* 1977. *Fluctuations of glaciers 1970–1975 (Vol. III).* Paris, International Commission on Snow and Ice of the International Association of Hydrological Sciences/UNESCO.

Müller-Lemans, H., M. Funk, M. Aellen and G. Kappenberger. 1995. Langjährige Massenbilanzreihen von Gletschern in der Schweiz. *Z. Gletscherkd. Glazialgeol.*, **30**, 1994, 141–160.

Oerlemans, J. and J. P. F. Fortuin. 1992. Sensitivity of glaciers and small ice caps to greenhouse warming. *Science*, **258**(5079), 115–117.

Oerlemans, J. and N. C. Hoogendoorn. 1989. Mass-balance gradients and climatic change. *J. Glaciol.*, **35**(121), 399–405.

Ohmura, A., P. Kasser and M. Funk. 1992. Climate at the equilibrium line of glaciers. *J. Glaciol.*, **38**(130), 397–411.

Orheim, O. 1970. Glaciological investigations of Store Supphellebre, west-Norway. *Nor. Polarinst. Skr.* 151.

Pfeffer, W. T. *and 7 others*. 1997. Numerical modeling of late glacial Laurentide advance of ice across Hudson Strait: insights into terrestrial and marine geology, mass balance, and calving flux. *Paleoceanography*, **12**(1), 97–110.

Schytt, V. 1964. Scientific results of the Swedish Glaciological Expedition to Nordaustlandet, Spitsbergen, 1957 and 1958. *Geogr. Ann.*, **46**(3), 243–281.

Takeuchi, Y., R. Naruse and P. Skvarca. 1996. Annual air-temperature measurement and ablation estimate at Moreno Glacier, Patagonia. *Bull. Glacier Res.* 14, 23–28.

Van de Wal, R. S. W. 1992. Ice and climate. (Ph.D. thesis, Utrecht University.)

Vincent, C. and M. Vallon. 1997. Meteorological controls on glacier mass balance: empirical relations suggested by measurements on glacier de Sarennes, France. *J. Glaciol.*, **43**(143), 131–137.

Warrick, R. A. and J. Oerlemans. 1990. Sea level rise. *In* Houghton, J. T., G. J. Jenkins and J. J. Ephraums, *eds. Climate change: the IPCC scientific assessment*. Cambridge, etc., Cambridge University Press, 257–281.

Woo, M.-k and B. B. Fitzharris. 1992. Reconstruction of mass balance variations for Franz Josef Glacier, New Zealand, 1913–1989. *Arct. Alp. Res.*, **24**(4), 281–290.

Zingg, T. 1951. Beziehung zwischen Temperatur und Schmelzwasser und ihre Bedeutung für Niederschlags- und Abflüssfragen. *International Association of Scientific Hydrology Publication* 32 (General Assembly of Brussels 195 — *Snow and Ice*), Vol. 1, 266–269.

MS received 28 January 1998 and accepted in revised form 29 June 1999