Temperature and precipitation climate at the equilibrium-line altitude of glaciers expressed by the degree-day factor for melting snow

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ABSTRACT. Several authors relate accumulation (or precipitation) at the glacier equilibrium-line altitude (ELA) to summer mean temperature using exponential or power-law functions. I analyze the accumulation–temperature relation at the ELA with a degree-day model using data from the 1992 paper by A. Ohmura and others. The dataset includes estimates at the ELA of winter balance and of 'winter balance plus summer precipitation' which represent respectively low and high estimates of annual accumulation, which is seldom measured. The Ohmura dataset only lists summer mean temperature, but I recover monthly temperatures for the whole year for 66 of the glaciers by assuming sinusoidal temperature variation through the year and using annual temperature range from a gridded climatology. Monthly degree-day sums are then estimated from monthly mean temperature and summed to give annual totals so degree-day factors for melting snow at the ELA are obtained. The degree-day factors fall close to those reported in the literature for glacier snowmelt, with averages of 3.5 ± 1.4 and 4.6 ± 1.4 mm d^{-1} K^{-1} for low- and high-accumulation estimates on the 66 glaciers. The degree-day model gives a family of accumulation–temperature curves that depend upon the annual temperature range, representing the contrast between maritime and continental climates.

1. INTRODUCTION

According to modern definition (Armstrong and others, 1973), the equilibrium line on a glacier separates the glacier's ablation area from its accumulation area. However, 18th-century pioneers like P. Bouguer (1698–1758) and H. B. de Saussure (1740–99) would have understood this to be the 'snow line', as they only recognized accumulation in the form of snow. During the 19th century, scientists like Brückner, von Buch, Esmark, Hugi, von Humboldt, Kurowski, Partsch, Payer, Penck, Ratzel, Richter and von Sonklar studied the snow line and knew that it was lower on a glacier than in its immediate surroundings (Zeller, 1893). The new concept was denoted by 'firn line'. Snow/firn lines vary greatly in both space and time, and, before the days of remote sensing, their direct study on glaciers was very laborious, although I. Venetz (1788–1859) was able to notice that snow lines were generally lower in the Swiss Alps in 1815–17 compared with 1811 (Berchtold and Bumann, 1990). Various methods were proposed to estimate long-term annual snow-/firn-line altitudes on a regional basis (Richter, 1885; Brückner, 1887; Kurowski, 1891; Zeller, 1893). These included a method of averaging altitudes for neighbouring mountain tops with and without glaciers, which is associated with the concept 'glaciation limit' (Enquist, 1916) or 'glaciation level' (Østrem and others, 1981). Although Ahlmann (1923) and Schytt (1949) recognized the significance of meltwater refreezing for mass balance and clearly stated that the firn line does not divide the ablation area from the accumulation area, Baird (1952) was apparently the first to use the term 'equilibrium line' to divide ablation and accumulation areas. The equilibrium line is generally lower than the annual snow/firn line, and the intervening 'superimposed ice zone' (if present) is the lowest part of the accumulation area where annual accumulation is in the form of ice from refrozen meltwater (Paterson, 1994, fig. 2.1) rather than snow.

The equilibrium-line altitude (ELA) is often determined as part of a programme of mass-balance measurement, and we have direct observations of annual ELAs for several hundred glaciers (Braithwaite, 2002; Dyurgerov, 2002; Dyurgerov and Meier, 2005). Equilibrium line, firn line, snow line and glaciation level are closely related concepts with a shared history, and some current methods of indirectly assessing ELA (Braithwaite and Müller, 1980; Benn and Lemkuhl, 2000) started life as methods for snow-/firn-line or glaciation limit.

P. Bouguer suggested that the snow/firn line coincides with the 0°C isotherm of annual temperature (Zeller, 1893), which is probably correct for the tropical glaciers that he studied in South America. However, by the late 19th century the general location of the snow/firn line was known to depend on summer temperature and annual snowfall, with large variations due to exposition (Zeller, 1893). According to Zeller (1893), K. von Sonklar described a link between accumulation at the snow line and summer temperature in 1866, but Zeller (1893) gives no details and I have never seen von Sonklar's work. Ahlmann (1924, 1948) related accumulation at the glaciation level to summer mean temperature (defined as the average temperature for June–August, denoted by T_{6-8}) with an exponential curve. Loewe (1971) relocated Ahlmann's (1924) curve to the ELA, and Ohmura and others (1992) added further data points. Accumulation at the ELA of a glacier is a key glaciological characteristic because the accumulation at the ELA is approximately equal to the area average of accumulation over the whole glacier (Ahlmann, 1948; Hoinkes and Rudolph, 1962; Trabant and March, 1999).

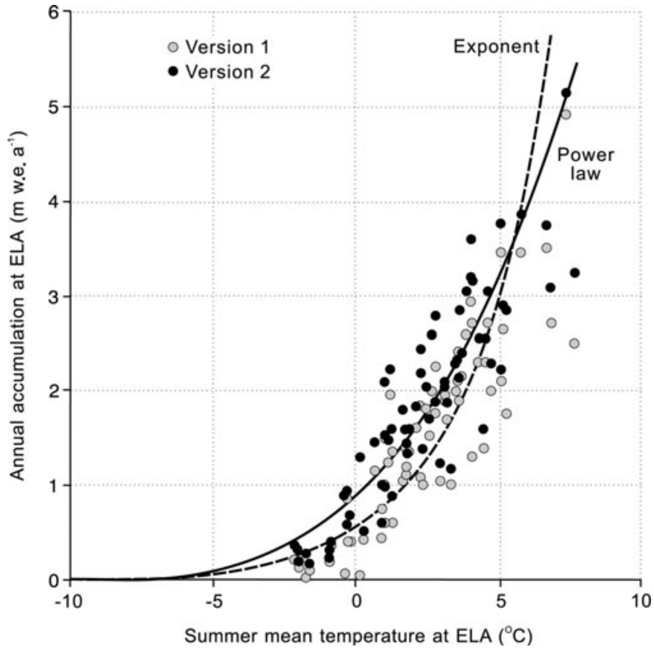

Fig. 1. Annual accumulation at the ELA vs summer (June–August) mean temperature (T_{6-8}) for 66 glaciers from Ohmura and others (1992). Version 1 refers to winter balance, and version 2 refers to 'winter balance plus summer precipitation'. Curves are for exponential and power laws.

There are several variants on the accumulation–temperature theme in the literature (see, e.g., Nesje and Dahl (2000, p. 67–71) for a concise summary and references). Some workers express the relationship in terms of 'winter precipitation' or 'winter accumulation'. The 'mean summer temperature' is variously expressed in terms of the mean of 1 May–30 September temperature (T_{5-9}) or the mean of 1 May–30 October temperature (T_{5-10}). Sutherland (1984), Ballantyne (1989) and Nesje and Dahl (2000) favour Ahlmann's exponential relation, but the exponent is variously expressed to base e or to base 10 (see Nesje and Dahl, 2000, figs 4.11 and 4.12).

Krenke and Khodakov (1966) use a third-power law to relate accumulation (or precipitation) to temperature. Leonard (1989) proposes an interesting extension of the power law of Krenke and Khodakov (1966) whereby data points from Loewe (1971) and Sutherland (1984) are replotted between two power-law curves. Leonard (1989) notes that two glaciers (Tsentralniy Tuyuksuiskiy in central Asia and Engabreen in Norway) fall outside the envelope formed by his two curves (see fig. 1 in his paper), but says 'The reason for these departures is not known'. I suggest an explanation in section 5 of this paper.

A relation between accumulation, or precipitation, and summer temperature is also implicit in maps of winter precipitation and summer degree months (in Østrem's terminology) at the glaciation level in southern Alaska (Østrem and others, 1981). In this example, the altitude of the glaciation level rises rapidly as one moves away from the coast towards the interior of North America. The winter precipitation and summer degree months also fall as one proceeds inland so that lower precipitation is associated with lower temperature, and the reverse.

A functional relation between accumulation/precipitation at the ELA and temperature is useful for a variety of purposes. For example, Kotlyakov and Krenke (1982) and Nesje and Dahl (2000, p. 70) assess precipitation in mountain areas by extrapolating temperature data to the ELA. Ballantyne (1989), Pfeffer and others (1997) and Hughes and others (2006) reconstruct former glaciers, while Davidovich and Ananicheva (1996) and Glazirin and others (2003) project the effects of future climate on glaciers.

It is not the point of the present paper to debate the relative merits of exponential and power-law relations between accumulation and temperature at the ELA, or to discuss the precise definition of summer mean temperature. The purpose of the paper is to apply a different kind of model, the so-called degree-day model, to the ELA of glaciers using the dataset of Ohmura and others (1992). Several workers have applied the degree-day model to calculate the full mass-balance profile over the whole altitude range of the glacier (Laumann and Reeh, 1993; Jóhannesson and others 1995; Braithwaite and Zhang, 1999, 2000; Braithwaite and others, 2003), but I now restrict it to the ELA (Braithwaite and others, 2006; Raper and Braithwaite, 2006; Braithwaite and Raper, 2007). The main results are (1) to estimate degree-day factors for melting snow on 66 glaciers, which represents a substantial extension of our present knowledge of degree-day factors, and (2) to show that the degree-day model predicts a family of curves linking accumulation at the ELA to summer temperature, similar to those reported in the literature.

2. CLIMATE AT THE ELA

Table 3 in Ohmura and others (1992) is an accessible and comprehensive listing of climate at the ELA for 70 glaciers, including several sites on the Greenland ice sheet. I drop four of the glaciers (Ward Hunt, Deception Island Glacier, Law Dome and Hodges Glacier) from further consideration, as they are not included in the topographic mask for the half-degree gridded climatology of New and others (1999), which is required below.

The data (Ohmura and others, 1992) refer to field measurements and include the ELA, winter mass balance, 'winter mass balance plus summer precipitation' and the average temperature T_{6-8} for the three summer months (June–August). The temperature data refer to 'the free atmosphere at the equivalent altitude as the ELA', and these data have 'an advantage over the screen-level air temperature, because the former [are] more easily accessible both in Nature and in models'. In the present study, I follow Braithwaite and others (2003) in correcting these air-temperature data to take account of the 'glacier cooling effect' such that screen-level temperatures over glaciers are supposedly lower than equivalent-altitude temperatures in the free atmosphere (Braithwaite, 1980). Kotlyakov and others (1997) assume a similar relation, which has the effect of reducing the variability of air temperature immediately over the glacier (e.g. at 2 m above the surface), compared with temperature variability in the wider region (Greuell and Böhm, 1998).

Ohmura and others (1992) are careful not to confuse winter balance with annual accumulation, although some workers regard accumulation, precipitation and winter balance as virtually interchangeable (Hughes and others, 2006). Accumulation is of great theoretical importance but is seldom, if ever, observed; see Anonymous (1969) for the distinction between winter balance and annual accumulation.

Ohmura and others (1992) are probably correct to regard their 'winter balance plus summer precipitation at the ELA' as being essentially the same as annual precipitation. For many glaciers, with a winter precipitation maximum, the difference between winter balance and annual accumulation may not be large. However, the dataset also includes glaciers with substantial precipitation in summer, where precipitation falls as rain as well as snow such that annual accumulation is not identical to annual precipitation. Tricart (1970) treats the ratio of accumulation to precipitation as a climatological characteristic of glaciers in different regions.

For the present paper, I assume that the (unknown) annual accumulation for each glacier is somewhere between the winter balance and 'winter balance plus summer precipitation' estimates of Ohmura and others (1992). The two variables are therefore treated as low and high estimates of the unknown annual accumulation (versions 1 and 2). This is safer than any wild assumption that accumulation always equals annual precipitation, and should also be an improvement over Braithwaite and others (2006) who calculate degree-day factors only for winter balance (for 180 glaciers). With this new interpretation, the data of Ohmura and others (1992) show a strong non-linear relation between accumulation at the ELA and summer mean temperature T_{6-8} (Fig. 1), where the curves represent exponential and third-power law relations respectively. These curves are recalculated to fit the plotted data rather than taken directly from the literature.

Throughout the study, accumulation versions 1 and 2 were correlated separately with various independent variables, but statistical results for the two versions (correlation coefficient and slope, but not intercept) are so similar for the same independent variable that I only give explicit results here for the combined sample. The correlations between accumulation (both versions) and exponential and third-power law functions of summer temperature are 0.78 and 0.85 respectively, and the corresponding root-mean-square (rms) errors are ± 0.66 and ± 0.56 m w.e. a^{-1}. Such curves 'explain' respectively 61% and 72% of the accumulation variance. Correlation coefficients for the two versions treated separately (sample sizes of 66) are already statistically significant at less than the 1% probability level.

In the following analyses, I use the data from Ohmura and others (1992) as they are, and I direct the reader to the original paper for details on sources and methods. I add two extra variables to the dataset from the gridded climatology of New and others (1999). These are the annual precipitation and annual temperature range for the 0.5° latitude/longitude square in which each glacier is located. These climate data refer to conditions at the average altitude of the topography within each square, which is usually hundreds, or even thousands, of metres below the ELA of glaciers within the square (Braithwaite and others, 2006). The annual temperature range probably does not change much with altitude, but the gridded precipitation is quite distinct from the precipitation at the glacier ELA (Braithwaite and Raper, 2002). Most meteorological stations in glacierized areas are located at low altitudes, i.e. in valleys, and the gridded precipitation therefore represents 'valley' precipitation as measured by standard meteorological stations, with all the problems of undercatch, exposure, etc. By contrast, the snow accumulation on the glacier is formed from the precipitation actually present (e.g. after any enhancement by increasing altitude, redistribution by snowdrift, topographical channelling and avalanche).

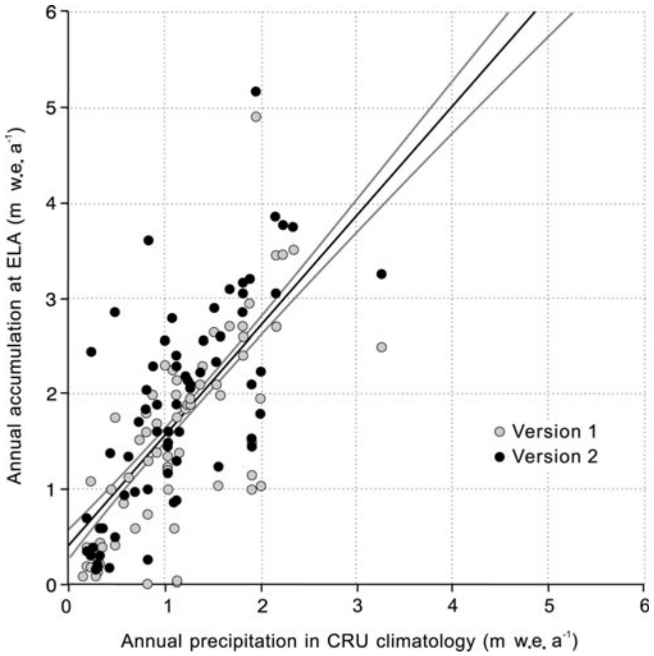

Fig. 2. Annual accumulation at the ELA plotted against annual precipitation for the grid squares in which the 66 glaciers are located. The regression line and its 95% confidence interval are shown. Version 1 refers to winter balance, and version 2 refers to 'winter balance plus summer precipitation' from Ohmura and others (1992). Precipitation data are from the gridded climatology of New and others (1999). CRU: Climatic Research Unit, University of East Anglia.

The accumulation on the remaining 66 glaciers from Ohmura and others (1992) is correlated with the (gridded) annual precipitation (Fig. 2). The correlation coefficient is 0.70, corresponding to an rms error of ± 0.75 m w.e. a^{-1}, which is somewhat lower than correlations between exponential and third-power law functions of temperature in Figure 1. For example, it only 'explains' 49% of the accumulation variance. Glacier accumulation is generally larger than gridded precipitation for the 66 glaciers. This probably involves a combination of effects. Firstly, precipitation usually increases with altitude such that glacier precipitation is higher than the gridded precipitation that is biased to measurements in valleys (Braithwaite and Raper, 2002). Secondly, the accumulation on the glaciers must be less than or at most equal to the (unknown) glacier precipitation (Braithwaite and others, 2003) because some precipitation may be in the form of rain that runs through snow at the ELA without contributing to mass balance (Tricart, 1970). The regression line in Figure 2 undoubtedly reflects these opposing effects but should not be regarded as 'universal' because precipitation lapse rates and glacier-valley elevation and exposure differences are implicit.

We might have expected the regression line (Fig. 2) to pass through the origin, as both accumulation and precipitation are ratio variables. However, even the 95% confidence interval in Figure 2 just fails to include the origin. Some points on the top lefthand side of Figure 2, with low precipitation and relatively high accumulation, may be skewing the regression line. These are Alaskan glaciers within a region of strong precipitation gradient (Østrem and

Table 1. Mean and standard deviation of degree-day factors for melting ice or snow, including results from the present study

Source	Type	Degree-day factor		
		Mean mm d^{-1} K^{-1}	Std dev. mm d^{-1} K^{-1}	Sample
Nordbogletscher, Greenland*	Glacier ice	6.9	±1.1	14 months
Qamanârssûp sermia, Greenland*	Glacier ice	7.8	±1.0	21 months
Hock (2003)	Glacier ice	8.9	±3.7	32 sites
Weissfluhjoch, Switzerland†	Seasonal snow	4.2	±1.0	28 melt seasons
Hock (2003)	Glacier snow	5.1	±2.2	18 sites
Present study				
(1) Winter balance (version 1)	Snow at ELA	3.5	±1.4	66 glaciers
(2) Winter balance plus summer precipitation (version 2)	Snow at ELA	4.6	±1.4	66 glaciers
(3) Combined (versions 1 and 2)	Snow at ELA	4.1	±1.5	66 glaciers

*Braithwaite and others (1995). †de Quervain (1979).

others, 1981), so possibly the average precipitation for a 0.5° latitude/longitude square greatly underestimates the precipitation on one or other side of the square.

3. DEGREE-DAY MODEL

As annual accumulation at the ELA is identical to annual ablation at the ELA, the supposed accumulation–temperature curve implies a relation between melting and temperature. Finsterwalder and Schunk (1887) were the first to assume that glacier melting depends on air temperature when it is above the melting point. This assumption is the main basis of the modern degree-day model (Braithwaite and Olesen, 1989; Huybrechts and others, 1991; Reeh, 1991; Laumann and Reeh, 1993; Jóhannesson and others 1995; Braithwaite and Zhang, 1999, 2000; Hock, 1999, 2003; Braithwaite and others, 2003, 2006; de Woul and Hock, 2005; Anderson and others, 2006; Lippert and others, 2006; Zhang and others, 2006a; Braithwaite and Raper, 2007). (The words 'degree-day' strictly refer to the way in which melting is calculated, but these models also include routines for snow accumulation and meltwater refreezing, so they calculate the full mass balance at the glacier surface.)

The basis of the degree-day approach is the assumed proportionality of melting to the positive temperature sum (degree-day sum) at the same point and for the same period (Braithwaite and Olesen, 1989). The temperature sums are defined relative to some chosen threshold, which is 0°C in the present case. The ratio between ablation and degree-day sum is called the degree-day factor. From field data, and from considerations of energy balance, it is generally believed that degree-day factors are lower for snowmelt than for ice melt under the same temperature conditions (Braithwaite, 1995; Hock, 1999, 2003; Braithwaite and Zhang, 2000; Braithwaite and others, 2003). Some workers have applied the degree-day concept to debris-covered glaciers (Kayastha and others, 2000; Mihalcea and others, 2006; Zhang and others, 2006b) and suggest that an increasing thickness of debris cover lowers the degree-day factor compared with the value for clean ice.

Because ablation depends on other variables than just temperature, the degree-day factor cannot be a universal constant. Braithwaite (1995) analyzed the degree-factor in terms of energy balance and demonstrated the obvious association between high albedo and low degree-day factor for snow. However, the energy-balance simulation of Braithwaite (1995, fig. 6) also suggests increased degree-day factors for ice, and reduced degree-day factors for snow, as summer temperatures are lowered to about +2 to –2°C. This may explain some high degree-day factors reported for ice, but the majority of the data points in Ohmura and others (1992) lie above this temperature range.

Some summary statistics for degree-day factors for ice and snow are listed in Table 1. Monthly estimates of degree-day factor for melting ice at the Greenland ice-sheet margin at Nordbogletscher and Qamanârssûp sermia (Braithwaite, 1995) are interesting, as they describe time variations by degree-day factor for ice. Similarly, time variations in degree-day factor for snow are illustrated by a 28 year series of snowmelt data from Weissfluhjoch, Switzerland, (2540 m a.s.l.) reported by de Quervain (1979). Degree-day factors listed by Hock (2003) for both ice (32 sites) and snow (18 sites) illustrate variations between different locations. The results clearly show a higher degree-day factor for ice than for snow. Even at the same locations (Nordbogletscher, Qamanârssûp sermia and Weissfluhjoch), there are substantial temporal variations, expressed by the standard deviation, presumably reflecting different weather conditions in different periods. The inter-site variations in Hock (2003) for both ice and snow are even larger than the temporal variations at the other sites.

From Table 1 it is clear that degree-day factors for ice and for snow are not precise single values, even at the same place. New estimates of degree-day factors for different conditions are therefore welcome.

4. DEGREE-DAY FACTORS AT THE ELA

Ohmura and others (1992) do not give the positive degree-day sums corresponding to the summer mean temperatures, but it is possible to estimate them by making some simple assumptions outlined here.

I follow Reeh (1991) and Hughes and others (2006) in assuming that monthly mean temperatures are sinusoidal around the annual mean temperature T_m, with an amplitude given by half the annual temperature range R. The summer

Table 2. Mean and standard deviation of degree-day factor at the ELA for 66 glaciers (Ohmura and others, 1992), depending upon different model assumptions

Sample	Cooling effect assumed	
	No	Yes
	mm d^{-1} K^{-1}	mm d^{-1} K^{-1}
(1) Winter balance (version 1)	2.7 ± 1.1	3.5 ± 1.4
(2) Winter balance plus summer precipitation (version 2)	3.6 ± 1.1	4.6 ± 1.4
(3) Combined (versions 1 and 2)	3.2 ± 1.2	4.1 ± 1.5

mean temperature T_{6-8} is given by Ohmura and others (1992), and I take the annual temperature range R from the gridded climatology of New and others (1999) for the grid square where the glacier is located. Mean temperatures for each month can then be calculated from these two quantities, and monthly degree-day sums and monthly probability of freezing can be estimated from the monthly temperature using the model of Braithwaite (1985), which assumes that temperatures within the month are normally distributed around the monthly mean. Annual degree-day sums and annual probability of freezing are then calculated by summing monthly values.

The accumulation at the ELA is just equal to the ablation, and, according to the degree-day model, snowmelt at the ELA is related to the annual degree-day sum by the degree-day factor for snowmelt. The possibility of meltwater refreezing at the ELA is neglected here so that all melting is in the form of snow, but this is re-examined below.

Different authors have defined 'summer temperature' (Northern Hemisphere) for different periods (e.g. using T_{6-8} for June–August, T_{5-9} for May–September or T_{5-10} for May–October). For a sinusoidal temperature distribution throughout the year, the temperature averages for different choices of summer period are highly correlated and it does not matter which one is used, although we should be consistent.

Calculated values of annual mean temperature T_m, annual degree-day sum and annual probability of freezing are assigned to each glacier in the dataset of Ohmura and others (1992). The accumulation at the ELA is strongly correlated ($r = 0.87$, significant at <1% level) with the annual degree-day sum in Figure 3, as we would expect. The corresponding rms error is ± 0.52 m w.e. a^{-1}.

The average degree-day factors at the ELA for the two versions of accumulation (Table 1) are much lower than the three averages for ice but are generally similar to the two averages for snow. If Ohmura and others (1992) have correctly estimated the summer precipitation at the ELA, the true annual accumulation is somewhere between versions 1 and 2, so the true degree-day factor for snow at the ELA lies somewhere between 3.5 ± 1.4 and 4.6 ± 1.4 mm d^{-1} K^{-1}. The 'combined' value of 4.1 ± 1.5 mm d^{-1} K^{-1} in Table 1 is midway between versions 1 and 2.

The above values of degree-day factor could depend upon the assumptions made in the model about glacier cooling effect (Braithwaite, 1980; Braithwaite and others, 2003). This possibility was tested by rerunning the degree-day model without the assumed cooling effect. The calculated degree-day factors for 'No' cooling effect are lower than before

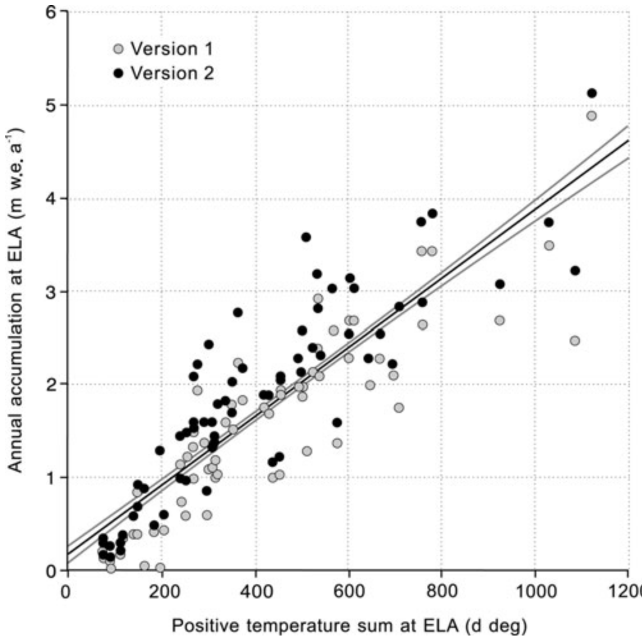

Fig. 3. Annual accumulation at the ELA plotted against annual positive degree-day total for 66 glaciers. The regression line and its 95% confidence interval are shown. Version 1 refers to winter balance, and version 2 refers to 'winter balance plus summer precipitation' from Ohmura and others (1992). Degree-day sums are estimated from summer (June–August) mean temperature (Ohmura and others, 1992) and annual temperature range (New and others, 1999), assuming a sinusoidal distribution of monthly temperatures throughout the year.

(Table 2), presumably because the model is now applying higher temperatures to the same accumulation. Some may argue that it does not really matter whether we assume cooling effect or not, as long as we use the appropriate lower or higher values of degree-day factor. However, the degree-day factors in the present study (Table 1) are already a little lower than the average from Hock (2003), so a further lowering due to neglect of cooling effect (Table 2) is going in the wrong direction. An even stronger cooling effect than assumed here would increase average degree-day factors in better agreement with the higher degree-day factors reported by Hock (2003).

Average degree-day factors for the different samples are plotted in Figure 4 together with 95% confidence intervals. There is a clear impression of statistically significant difference between degree-day factors for ice and snow, but differences between the different samples for snow are close to being insignificant. This is partly an artefact of the relatively large standard deviation in the Hock (2003) dataset compared with lower standard deviations for versions 1 and 2.

5. ACCUMULATION AT THE ELA

Accumulation at the ELA is re-plotted against summer mean temperature in Figure 5, together with curves calculated with a degree-day factor 4 mm w.e. d^{-1} K^{-1}. The curves represent respectively low (15 K), medium (25 K) and high (35 K) values for the annual temperature range. (Note that these are not 'fitted' curves but are simply smooth curves drawn through values calculated by the model.) The significance of annual temperature range for the relation between annual ablation and summer temperature was first pointed out by Reeh

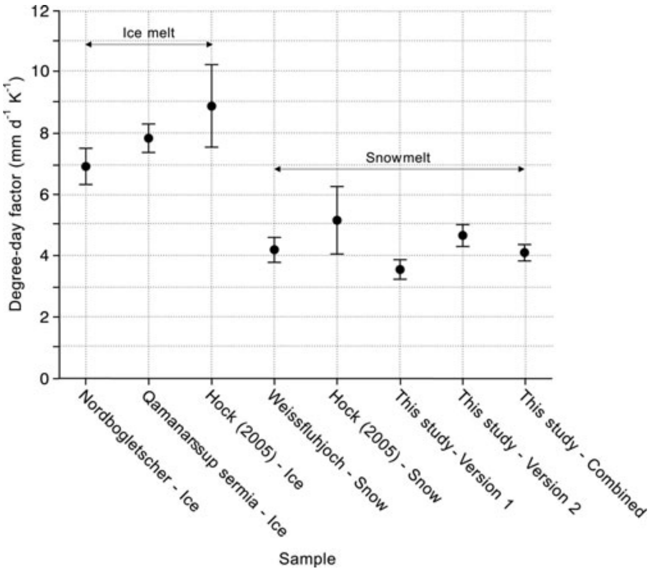

Fig. 4. Mean and 95% confidence interval of degree-day factors at the ELA compared with degree-day factors reported in the literature. Version 1 refers to winter balance, version 2 refers to 'winter balance plus summer precipitation', and 'Combined' is the average of the two.

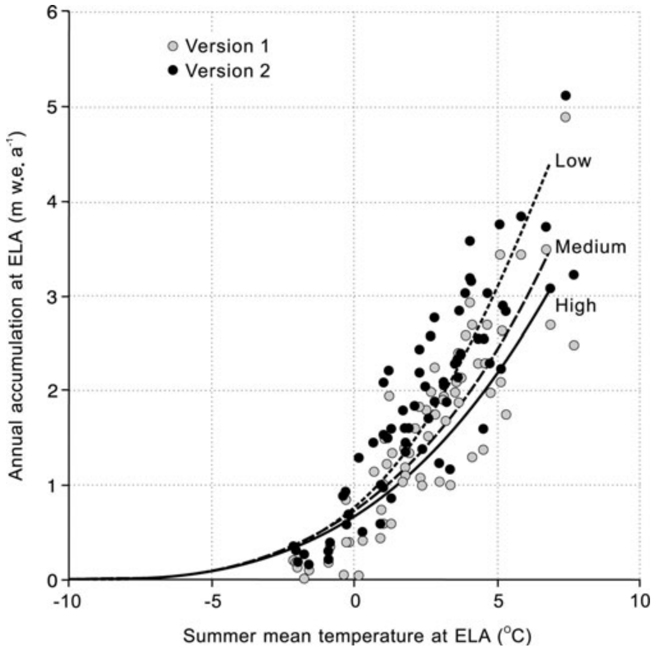

Fig. 5. Annual accumulation at the ELA plotted against summer (June–August) mean temperature (T_{6-8}) for 66 glaciers from Ohmura and others (1992). Version 1 refers to winter balance, and version 2 refers to 'winter balance plus summer precipitation'. Curves are for the degree-day model with 'low', 'medium' and 'high' values of annual temperature range, respectively 15, 25 and 35 K.

(1991). For a particular value of summer temperature T_{6-8}, a low temperature range implies relatively warm temperatures for other months outside the June–August period, especially May and September. This implies 'extra' ablation, in addition to that in June–August, and the annual accumulation must be correspondingly higher to offset this extra ablation. By the same token, a high temperature range implies lower temperatures outside the period June–August and correspondingly little extra melting at the ELA.

Nesje and Dahl (2000, p. 68) say the exponential relation between winter accumulation (or winter precipitation?) and summer temperature 'is of global application'. Referring to Ahlmann (1924), Ahlmann (1948) says, 'In the preliminary paper we concluded that the summer isotherms at the glaciation limit are relative isohyets. Our investigations since 1931 have shown however that the factors determining the ablation are so many, and cooperate in so complicated a manner, that this rule cannot be applicable generally, but only in limited areas. It is also probable that the correlation curve in Fig. 32 applies to Norway only, and would be different in other climatic districts.' On the basis of present results, I conclude that the truth lies between these two positions of 'global' (Nesje and Dahl, 2000) and 'local' (Ahlmann, 1948) applicability.

The accumulation at the ELA cannot be specified in terms of a single function of summer temperature. The degree-day model implies a family of curves corresponding to different values of the annual temperature range, as first pointed out by Reeh (1991) and confirmed by Figure 5. In section 1, I refer to Leonard (1989) who extended the approach of Krenke and Khodakov (1966) by plotting two power-law curves and found only two glaciers outside the envelope formed by the two curves. As these two glaciers are extreme examples of continental (Tsentralniy Tuyuksuiskiy, central Asia) and maritime (Engabreen, west Norway) conditions, Leonard's envelope reflects the extremes of annual temperature range, in agreement with results here.

The use of any model to calculate conditions under a different climate involves the difficulty of deciding what things change and what can be regarded as constant for the problem at hand. With the degree-day model (e.g. as used by Braithwaite and others, 2003), we have to assume that degree-day factors for snow and ice remain constant while we study the effects of changing temperature and precipitation on glaciers. Similarly, you have to assume that the favoured relation between accumulation and summer temperature remains the same (e.g. the coefficients in the exponential or power-law equations). However, quantities like annual temperature range, temperature lapse rate and precipitation distribution throughout the year can also change as climate changes. High-resolution climate models could give clues about how these quantities may change in glacierized areas.

6. CONCLUSIONS

The degree-day model describes the accumulation at the ELA across a range of climate regimes represented by different values of annual temperature range. Users of exponential or power-law relations between accumulation and summer temperature should be prepared to calculate different curves to use in different climatic regions.

No single degree-day factor is applicable to all glaciers, but 76% of accumulation variance at the ELA for 66 glaciers can be explained by temperature using a constant degree-day factor, i.e. 4.1 ± 1.5 mm d^{-1} K^{-1}. A constant degree-day factor is therefore a useful first assumption for an 'unknown glacier', although we should try to predict geographical variations in degree-day factor to achieve better predictions of accumulation at the ELA.

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