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Why Do We Expect Glacier Melting to Increase Under Global Warming?

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1. Introduction

Media stories about global warming almost always mention "melting glaciers" and their effects upon global sealevel. The reader might therefore ask why the title of this chapter includes a question mark. It may seem blindingly obvious that global warming will cause a rise in global sealevel with a substantial contribution from melting glaciers, but the reasons are less obvious despite the copious literature. For example, the IPCC assessment reports 1991, 1996, 2001 and 2007 all include many references to published papers on glaciers (IPCC, 2011). These are admirable summaries of *who* has said *what,* or *who* has done *what,* but they do not explain *why*. My purpose for the present chapter is to provide a clear narrative on *why* we expect glacier melting to increase with any change in temperature, whether due to global warming or to natural fluctuations. By its very nature, the *why* of increased glacier melting must also answer the question of *how much extra melting?*

A simple and direct relation between glacier melt and air temperature has not always been as self-evident as it may appear today. For example, Hoinkes (1955) wrote: *In recent years many authors, on the basis of careful studies, have come to the conclusion that summer temperature is to be regarded as the most important factor influencing the behaviour of glaciers* and he quotes four references to support this statement, to which I could add many more. Hoinkes (1955) then goes on to say: *This result is not in contradiction to the results of the measurements which are given here (according to which radiation is the main source of energy for the ablation of the alpine glaciers) so long as it is not combined with the idea that the greater heat exchange from air to ice during a hot summer is sufficient to account for the greater ablation*. Braithwaite (1981) discusses the fallacy behind this statement from Hoinkes (1955) that seems to conflate large average values of radiative energy with large variations in melt energy. The present chapter demonstrates empirically that higher melt is associated with higher air temperature. I do this in three stages: (1) correlating daily melt with daily air temperature for some Arctic and/or Greenland locations, (2) linking the results to the wider literature on the degree-method, and (3) showing that recent changes in glacier mass balance in the Alps are consistent with higher air temperatures in and around the Alps. I refer the reader to Kuhn (1979), Braithwaite (1980 and 1981), Ambach (1988), Braithwaite and Olesen (1990a and 1990b) and Braithwaite (1995) for the theoretical interpretation of the melt-temperature relation in terms of the energy balance at the glacier surface.

2. Short-term variations in glacier melt

I start the narrative by considering studies of short-term variations in glacier melt that were made in Arctic Canada (1960-63), South Greenland (1979-83), West Greenland (1980-86) and North Greenland (1993-94). See Fig. 1 and Table 1 for locations and periods. The high arctic bias of the measurements should be obvious with four of the six sites being in the region of year-round sea ice cover leading to a relatively dry continental climate (Braithwaite, 2005).

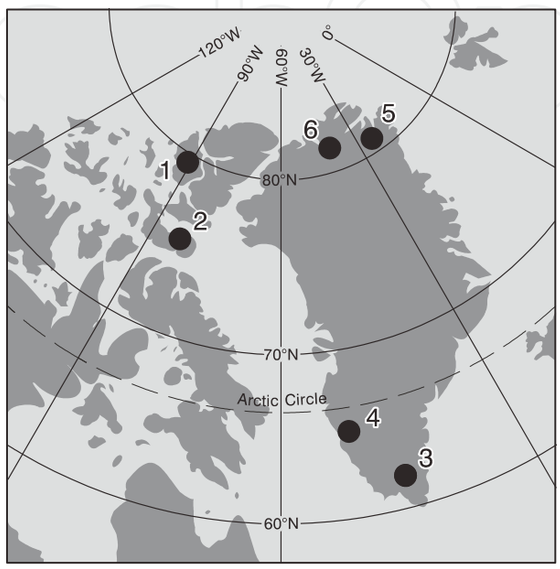

Fig. 1. Locations of the six sites in the present study. (1) is White Glacier, Axel Heiberg Island, (2) is Sverdrup Glacier, Devon Island, (3) is Nordbogletscher, South Greenland, (4) is Qamanârssûp sermia, West Greenland, (5) is Kronprins Christians Land, North Greenland, and (6) is Hans Tavsen Ice Cap, North Greenland.

The data consist of daily, or nearly daily, measurements made on stakes drilled into the ice near to a climate station, yielding a continuous record of air temperature measurements made 1.5-2 m above the surface, located on or very near the glacier. As a technical point, I should say that the stake measurements strictly refer to ice *ablation*, i.e. material lost by the glacier surface, rather than *melt* in the strict sense. However, the difference between ablation and melt is small in mass-balance terms under the conditions considered, i.e. with relatively small sublimation, condensation or re-freezing, and is negligible compared with stake measurement errors. You can therefore treat ablation and melt as almost synonymous for the present chapter but I will use the term ablation when referring to measurements made with stakes drilled into the ice.

Braithwaite (1981) analysed the data from Arctic Canada (see Section 2.1) and concluded that there was a useful relation between ice ablation and air temperature. When Braithwaite joined the staff of the Geological Survey of Greenland (GGU) in 1979, he persuaded colleagues Poul Clement and Ole B. Olesen to add daily ablation measurements to routine programmes at Nordbogletscher and Qamanârssûp sermia, see Sections 2.2 and 2.3 below.

In 1993 and 1994 Braithwaite was able to make relatively short visits to North Greenland (see Section 2.4), which is normally difficult to access, and gave high priority to daily measurements of ablation as part of energy balance studies (Braithwaite et al, 1998a and 1998b).

Site	Latitude (° N)	Longitude (° W)	Altitude (m a.s.l.)	Periods
1). White Glacier, Axel Heiberg Island	79	91	200	July-August 1960 June-August 1961 July 1962
2). Sverdrup Glacier, Devon Island	76	83	300	July-August 1963
3). Nordbogletscher, South Greenland	61	45	880	June-August 1979-1983
4). Qamanârssûp sermia, West Greenland	64	49	790	June-August 1980-1986
5). Kronprins Christians Land, North Greenland	80	24	380	July 1993
6). Hans Tavsen Ice Cap, North Greenland	83	36	540	July-August 1994

Table 1. Locations of the sites used for this study.

For this chapter, we represent the relation between daily ablation a_t and daily mean temperature T_t by the simple linear equation below:

$$a_t = a + \beta.T_t + e_t \tag{1}$$

The subscript t denotes the day of record and e_t represents the error in the equation. If we have a series of measurements for a_t and T_t covering N days, we can evaluate the intercept a and slope β parameters using the well-known least-square algorithm of linear regression, available in many computer data packages. According to this, a and β parameter values are chosen to minimise the variance of the error term e_t. The square root of the error variance, i.e. standard deviation of the error term, is often called the root mean square error, or RMSE. If the RMSE is relatively small compared with the fluctuating values of ablation we can say that air temperature is a good predictor of ablation. Alternatively, the correlation coefficient associated with the regression equation (1) should be relatively high.

We can regard a series of parallel ablation and temperature data for N days as a statistical sample. We cannot expect to find the same a and β values for different samples although we might hope that they will be similar to each other. For the study of the ablation-temperature relationship, the hypotheses are (1) the correlation coefficient should be relatively high (or the RMSE should be relatively low) for the sample in question, and (2) the a parameters from different samples should be similar if not identical, and same should hold for the β parameters. The proposed ablation-temperature relation would be useful if both (1) and (2) were true such that we could accurately calculate glacier ablation for situations where it is not measured.

2.1 Arctic Canada

I took the data from the work of Fritz Müller (1926-80) and his colleagues, working on White Glacier, (Axel Heiberg Island, NWT, Canada) and Sverdrup Glacier (Devon Island, NWT, Canada). Müller and Keeler (1969) discuss the accurate measurement of daily, or nearly daily, ablation in connection with studies of energy balance at the glacier surface. Their ablation measurements involved attempts to measure accurately both the lowering of the

glacier surface by mass loss and the loss of material from within the surface layer, i.e. changes in effective density of the surface material. The formation of a low-density "weathering crust" and its decay a few hours, or even a few days, later depends upon the weather conditions. For example, selective absorption of global radiation around grain boundaries or dirt particles, e.g. on relatively sunny days, leads to formation of the weathering crust such that more ice ablates than is indicated by the measurement of surface lowering. The weathering crust disappears when stormy/overcast weather follows sunny weather: measurement of surface lowering then overestimate the amount of ablation. Figs 1 to 3 in Müller and Keeler (1969) nicely illustrate these processes.

For the regression analyses underlying Table 2, Müller and Keeler (1969) give the daily ablation for White Glacier 1961 and 1962, and for Sverdup Glacier 1963, and I extracted daily ablation for White Glacier 1960 from Andrews (1964). I found the corresponding air temperature data in Andrews (1964), Müller and Roskin-Sharlin (1967), Haven et al (1965) and Keeler (1964). Fritz Müller and colleagues set an admirable standard in documenting their work by providing extensive tables of data for possible use by later researchers.

Location	Year	Month	α (mm d^{-1})	β (mm d^{-1} K^{-1})	Days	ρ	RMSE (mm d^{-1})
White Gl.	1960	Periods	4	5.67	16	0.67	±13
White Gl.	1961	June	7	3.74	14	0.37	±7
White Gl.	1961	July	6	4.64	31	0.75	±8
White Gl.	1961	Aug,	-13	8.51	18	0.77	±16
White Gl.	1962	Periods	-3	7.89	11	0.84	±11
Sverdrup Gl.	1963	July	10	2.18	23	0.20	±15
Sverdrup Gl.	1963	Aug.	-14	10.36	10	0.89	±9
		Combined	1	6.13	123	0.74	±13

Table 2. Regression equation linking daily ablation and temperature for two sites in Arctic Canada.

Correlation coefficients between daily ablation and daily mean temperature (Table 2) are generally reasonably high, i.e. greater than the 0.71 that corresponds to "explanation" of 50% of the ablation variance. However, depending on the way in which the samples are sub-divided, low correlations ρ can also occur, e.g. for 14 days in June 1961 for White Glacier and for 23 days in July 1963 at Sverdrup Glacier. In these cases, the low correlation coefficients coincide with relatively large values of intercept α and low values of slope β. This phenomenon is a property of regression lines that tend to the horizontal as correlation coefficients tend to zero. The low values of correlation coefficients for these periods could reflect excessive measurement errors in the ablation data or a real lack of temperature-dependence in the energy balance for these periods. I will explore this issue in future work.

I am reluctant to calculate confidence intervals for the α, β and ρ parameters for the small samples in Table 2 as the background theory assumes that the e_t should be purely random. This is probably not the case for various reasons, e.g. serial correlation of measurement errors and persistence of certain weather types over many days. The solution is to pool all the data into a single regression equation. This produces a larger sample of 123 days, which should be more reliable from the statistical point of view, but suppresses any real differences between the individual series. The bottom line in Table 2 shows the parameters

for this combined sample that represent a rough average of the individual samples. The underlying pattern is a small positive intercept (1 mm d^{-1}) and a slope of about 6 mm d^{-1} K^{-1}.

2.2 Nordbogletscher, South Greenland

The ablation data from Arctic Canada (Section 2.1) involve measurements of surface lowering together with attempts to measure density changes within the glacier surface. Müller and Keeler (1969) fully discuss the latter but the measurements seem very tedious to make and are probably not very reliable. We therefore decided to concentrate solely on measurements of surface lowering at Nordbogletscher and Qamanârssûp sermia, and to treat variations in surface density as an unavoidable error.

Year	Month	α (mm d^{-1})	β (mm d^{-1} K^{-1})	Days	ρ	RMSE (mm d^{-1})
1979	July	15	4.53	31	0.48	±21
	Aug.	10	4.64	30	0.60	±13
1980	June	4	7.58	30	0.81	±14
	July	13	4.74	31	0.57	±9
	Aug.	15	2.13	31	0.23	±14
1981	June	10	6.38	29	0.78	±14
	July	1	6.86	31	0.75	±16
	Aug.	1	5.02	31	0.83	±8
1982	June	14	5.54	30	0.82	±9
	July	6	6.38	31	0.70	±10
	Aug.	6	6.36	31	0.83	±10
1983	June	4	6.38	30	0.80	±12
	July	8	5.07	28	0.86	±9
	Aug.	8	5.22	31	0.85	±10
	Combined	7	5.68	425	0.74	±13

Table 3. Regression equation linking daily ablation and temperature at Nordbogletscher, South Greenland.

The field team made the ablation measurements at Nordbogletscher on one stake close to the edge of the ice and we take the temperature data from the nearby base camp. Air temperature above the melting point generally decreases as one proceeds onto a glacier (Braithwaite, 1980: Braithwaite et al., 2002) but the present measurement site is close enough to the ice edge not to show such "cooling effect". Ablation was measured every day in early evening, while temperature data refer to a 24-hour day in local time. This difference in timing introduces an extra small error into the melt-temperature correlation.

Once again, as in Table 2, correlation coefficients between daily ablation and daily mean temperature for Nordbogletscher (Table 3) are generally reasonably high, e.g. 0.70 to 0.86, but there are also periods with low correlations 0.23, 0.48, 0.57 and 0.60. The latter values are associated with relatively high intercept (10-15 mm d^{-1}) and low slope values (2.13 to 4.74 mm d^{-1} K^{-1}) as in Table 2. However, a high intercept (14 mm d^{-1}) also occur in June 1982 with a high correlation (0.82) when the slope is not especially low (5.54 mm d^{-1} K^{-1}).

The combined sample for all data, covering a total of 425 days, shows a slightly higher positive intercept (7 mm d^{-1}) and similar slope (about 6 mm d^{-1} K^{-1}) compared to the Arctic Canada results in Table 2. It is interesting that the respective combined samples for Arctic

Canada and for Nordbogletscher show identical root-mean square errors (±13 mm d^{-1}) despite not measuring variations in surface density at the latter site.

2.3 Qamanârssûp sermia, West Greenland

The ablation measurements at Qamanârssûp sermia were made on three stakes, within a few metres of each other and close to the edge of the ice while temperature data are taken from the nearby base camp. The results from Qamanârssûp sermia (Table 4) show generally similar patterns to the other sites (Tables 2 and 3). There are some high correlations (0.70 to 0.95) but also some low correlations (0.43 to 0.60). Both intercept α and slope β seem more variable than in the previous cases although the combined sample shows a small positive intercept (3 mm d^{-1}) with only a slightly higher slope of about 8 mm d^{-1} K^{-1}.

Year	Month	α (mm d^{-1})	β (mm d^{-1} K^{-1})	Days	ρ	RMSE (mm d^{-1})
1980	June	2	7.35	10	0.95	±9
	July	-6	7.70	28	0.49	±19
	Aug.	-20	10.34	23	0.79	±12
1981	June	4	7.24	26	0.85	±13
	July	-19	10.99	28	0.72	±23
	Aug.	4	6.09	31	0.76	±14
1982	June	6	6.13	17	0.74	±13
	July	0	8.88	27	0.81	±13
	Aug.	2	8.90	29	0.87	±16
1983	June	9	7.85	25	0.82	±18
	July	5	8.54	25	0.70	±24
	Aug.	7	6.87	29	0.70	±17
1984	June	9	4.76	21	0.58	±16
	July	-10	9.60	23	0.75	±20
	Aug.	-5	9.32	26	0.77	±17
1985	June	2	9.76	22	0.80	±20
	July	21	4.96	23	0.43	±16
	Aug.	9	6.61	25	0.76	±15
1986	June	10	4.74	27	0.60	±20
	July	-7	9.12	31	0.78	±18
	Aug.	-8	9.32	28	0.80	±19
	Combined	3	7.68	524	0.78	±18

Table 4. Regression equation linking daily ablation and temperature at Qamanârssûp sermia, West Greenland.

We had expected that measuring ablation at three stakes and averaging the results would give us slightly more accurate data than at Nordbogletscher where we used only one stake. It is, therefore, rather disappointing that the root-mean square error at Qamanârssûp sermia is actually slightly higher than at Nordbogletscher, i.e. ±18 compared with ±13 mm d^{-1}.

2.4 North Greenland

The ablation measurements in Kron Prins Christians Land (KPCL) and Hans Tavsen Ice Cap (HTIC) were made on 10 stakes, and air temperature data were taken from a station a few

metres from the stakes. The march of modern technology was marked by the fact that the temperature data were recorded on a digital data logger while earlier studies used data from thermographs checked by manual readings of mercury-in-glass thermometers.
The results from North Greenland show very low negative intercept and high slope in one case and small positive intercept with relatively low slope in the other case. When the two samples are pooled, the overall pattern is for a small positive intercept (3 mm d^{-1}) and a slope of about 7 mm d^{-1} K^{-1}.

Location	Year	Month	α (mm d^{-1})	β (mm d^{-1} K^{-1})	Days	ρ	RMSE (mm d^{-1})
KPCL	1993	July	-14	13.27	20	0.76	±6
HTIC	1994	July-Aug.	3	5.20	35	0.88	±6
		Combined	3	6.95	55	0.78	±10

Table 5. Regression equation linking daily ablation and mean temperature for two sites in North Greenland.

We had expected that measuring ablation at ten stakes and averaging the results would give us more accurate data than at Nordbogletscher and Qamanârssûp sermia, and the root-mean square error is indeed slightly lower than in the previous cases, i.e. only ±10 mm d^{-1} for the combined sample. More worrying is that the 10-stake measurements show that ablation varies by about ±10% of mean ablation within a few metres, probably due to micro-scale variations in albedo (Konzelmann and Braithwaite, 1995: Braithwaite et al., 1998b). This implies a ±10% error in any quantity calculated from stake measurements, including degree-day factors (see below).

2.5 Ablation-temperature correlation

The daily measurements of ablation in the four cases, covering a total of 1,127 days, show reasonably high correlations (0.74, 0.74, 0.78 and 0.78) such that air temperature variations explain slightly more than 50% of daily ablation variations. Errors in the measurements probably account for quite a sizeable percentage of the unexplained variance. The intercepts in the regression lines are slightly positive (1 to 7 mm d^{-1}) and the slopes are about 6 to 8 mm d^{-1} K^{-1} in round figures.
For readers who prefer graphs to numbers, Fig. 2 shows plots of ablation versus temperature for the four cases. Overall, the ablation-temperature relations are remarkable consistent despite differences in geographical setting from 61 to 83 °N. The occurrence of some negative ablation values in Figs 2b and 2c are clear signs of measurement errors as ablation should never be less than zero. The 95% confidence intervals around the regression lines give an impression of possible sampling errors in the regression lines. In the following section, I try to generalize this overall relation and compare it to results from other areas. However, I do intend in future to look at the energy balance variations between the different periods in Tables 2 to 5 to see if I can explain the apparent anomalies of low or high slope in the ablation-temperature regression equations.
The daily ablation measurements were very laborious to make as they involved manual measurements by human operators every day. This meant that fieldworkers had to live for many weeks on, or near, the glaciers. This probably explains why nobody has attempted to replicate the measurements elsewhere, e.g. in Svalbard. If we are realistic, we have to agree that glaciologists may never again spend such long continuous periods in the field although

new technology may allow unattended ablation measurements over periods of many months (Bøggild et al, 2004: Hulth, 2011). One purpose in writing this chapter is to motivate others to make similar ablation measurements in key areas.

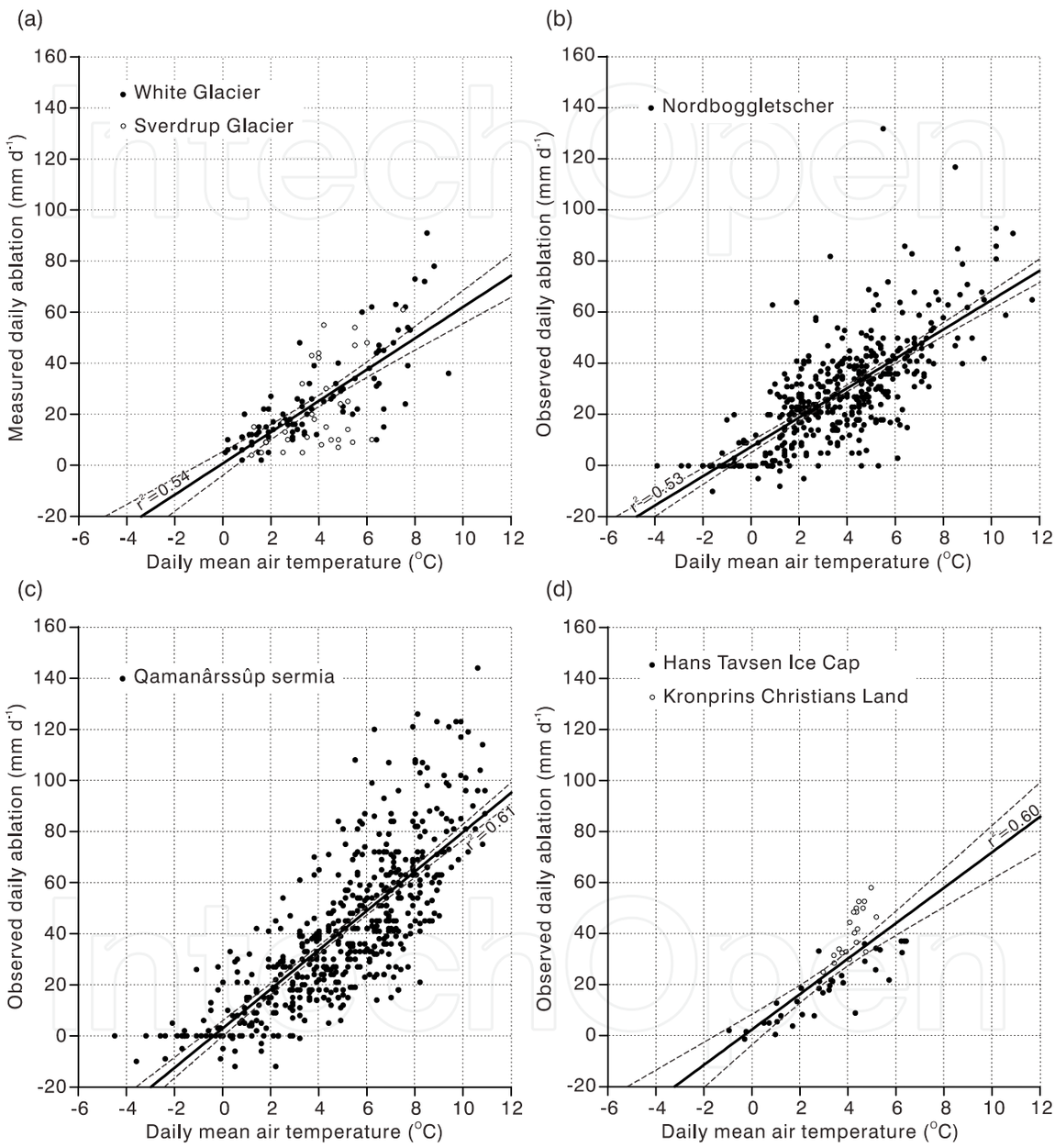

Fig. 2. Daily ablation versus daily mean temperature for: (a) Arctic Canada, (b) South Greenland, (c) West Greenland and (d) North Greenland.

3. Ablation and degree-day total

The intercept α in equation (1) represents the ablation that occurs when daily mean temperature equals 0 °C. Our statistical analysis shows that α is generally positive but quite

small, i.e. there is relatively little ablation when daily mean air temperature is at 0 °C or below. Melt may occur at low temperature if energy from global radiation is high enough to supply melt energy as well as maintaining a heat flux from the glacier surface to the overlying atmosphere (Ohmura, 1981: Kuhn, 1987). However, another explanation lies in the choice of daily mean temperature as our independent variable. For example, air temperature may be above the melting point for part of the day with substantial ablation even if the daily mean temperature is at or below the freezing point (Arnold and MacKay, 1964). We can overcome this problem by only considering air temperatures that are above the melting point. This leads to the degree-day approach, whereby melt during any period is assumed proportional to the sum of positive temperatures during the same period. The approach is well established in hydrology (De Walle and Rango, 2008).

3.1 Degree-day model

With modern data loggers, the sum of positive temperatures during any period can be achieved simply by adding successive hourly values and dividing the total by 24. Data at Nordbogletscher and Qamanârssûp sermia were collected by thermographs, which are now obsolete, supplemented by readings of maximum and minimum temperatures by mercury-in-glass thermometers. A reasonable estimate of the positive degree-day total for each day at these stations can be achieved as the sum of positive values of daily mean temperature (counted twice), daily maximum and daily minimum temperatures, and dividing the resulting daily sum by 4. Calculated in this way, the positive degree-day total is identical to daily mean temperature for high temperatures and zero for low temperatures. There is an intermediate region where daily degree-day total is already positive while daily mean temperature is negative. The extent of this intermediate region is determined by the daily temperature range.

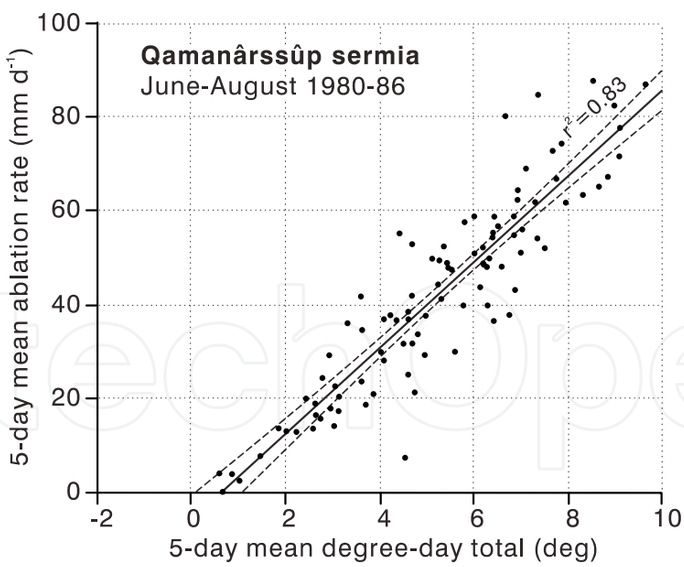

Fig. 3. 5-day averages of ablation and daily degree-day total at Qamanârssûp sermia, West Greenland.

There is a relatively good correlation between ablation and daily degree-day total at Qamanârssûp sermia, as calculated above, but the correlation is not much better than that between ablation and daily mean temperature (Fig. 2c). The apparently negative values of ablation in Fig. 2c are clear evidence of measurement errors as ablation cannot be less than

zero if accurately measured. There are similar values for Nordbogletscher (not shown). Braithwaite et al (1998b) avoided similar situations for North Greenland data by an objective method of screening gross errors, e.g. due to simple misreading of the measurement scale by several centimetres. However, for technical reasons, I cannot apply the same method retrospectively to the Nordbogletscher and Qamanârssûp sermia data.

As an alternative, I applied 5-day averaging to both daily ablation and daily degree-day total to see if this improves the correlation, which is obviously the case (Fig. 3). A particular type of measurement errors, involving a gross misreading on one visit to the stake, will affect ablation data for that day and the following day, and the averaging should eliminates this error if the two data points happen to fall within the same 5-day "window" for averaging. For this, or other reasons, there are no spurious negative values of ablation in Fig. 3. Krenke and Khodakov (1966) commented on a similar improvement in ablation measurements made over a few days. The results from Nordbogletscher are similar but I omit the graphs in the interests of conciseness.

3.2 Degree-day factor

Scatter plots like Fig. 3 demonstrate the validity of the degree-day approach for situations where daily ablation readings are available. This is not often the case. More commonly, workers visit the glacier at intervals of weeks or months and measure ablation totals for this time interval and then compare them with the degree-day total for the same period. The ratio of these longer-term totals is the degree-day factor. Estimates of degree-day factor are available for a number of locations (Table 6) and show widespread variations. We cannot regard the conditions underlying the listed values as being uniform and, no doubt, some of the variations in Table 6 will be due to methodological differences as well as real differences in glacier-climate conditions. However, the result that degree-day factors for snow are generally lower than for ice seems plausible, and Braithwaite (1995) explains this in energy balance terms. Ambach (1988) and Braithwaite (1995) also show that degree-day factors for melting ice may be quite large for low temperatures but not for high temperatures.

Many workers cite Ohmura (2001) for a physical explanation of the melt-temperature relation but Ohmura (2001) overlooks the temperature sensitivity of the different energy balance terms that is relevant rather than their absolute magnitudes. Ohmura (2001) is correct in saying that incoming longwave radiation is generally greater than the sensible heat flux to glacier surface, but the temperature sensitivity of sensible heat flux is generally greater than the temperature sensitivity of longwave radiation (Kuhn, 1979: Braithwaite, 1981: Ambach, 1988: Braithwaite, 1995) and hence accounts for a greater share of the degree-day factor.

I regard the results in Table 6 as "work in progress" and I hope that more data will become available representing a wider range of conditions. In particular, we can now recognize the effects of debris-cover and sublimation on ice ablation (Zhang et al. 2006) so future tables may not be so simply divided into results for "snow" and for "ice". Until we get more data, I regard the present results as a safe basis for three hypotheses that we can test in future studies:

1. We can reliably calculate snow and ice ablation from degree-day totals,
2. Degree-day factors for snow ablation are generally about 3 to 5 mm d^{-1}, and
3. Degree-day factors for ice ablation are generally about 6 to 8 mm d^{-1} K^{-1}.

In the following section, I discuss the modelling of glacier mass balance using the proposed relationship between ablation and positive degree-day total.

Location	Type	Mean	St. devn.	Sample size
De Quervain (1979)	Snow	4.2	±1.0	28 melt seasons
Braithwaite & Olesen (1988)	Snow	2.5	-	1 glacier
Laumann and Reeh (1993)	Snow	3.5 to 4.5	-	3 glaciers
Jóhannesson et al. (1993)	Snow	2.8 to 5.7	-	3 glaciers
Vincent and Vallon (1997)	Snow	3.8	-	1 glacier
Hock (1999)	Snow	4.4	-	1 glacier
Hock (2003)	Snow	5.1	±2.2	18 sites
Anderson et al (2006)	Snow	4.6	-	1 glacier
Radic (2008)	Snow	4.8	±1.5	44 glaciers
Braithwaite (2008)	Snow	4.1	±1.5	66 glaciers
Shea et al (2009)	Snow	3.0	±0.4	9 glaciers
Schytt (1964)	Ice	13.8	-	1 glacier
Orheim (1970)	Ice	6.1 & 6.5	-	2 Seasons at 1 site
Braithwaite (1977)	Ice	5.4	± 2.3	4 glaciers
Braithwaite (1981)	Ice	5.5 to 7.8	-	2 glaciers
Ambach (1988)	Ice	18.6		1 site
Braithwaite & Olesen (1988)	Ice	7.2	-	1 glacier
Laumann and Reeh (1993)	Ice	5.5 to 6.0	-	3 glaciers
Jóhannesson et al. (1993)	Ice	6.4 to 7.7	-	
Vincent and Vallon (1997)	Ice	6.2	-	1 glacier
Hock (1999)	Ice	6.3	-	1 glacier
Hock (2003)	Ice	8.9	±3.7	32 sites
Anderson et al (2006)	Ice	7.2	-	1 glacier
Zhang et al (2006)	Ice	6.5	±3.7	15 glaciers
Radic (2008)	Ice	7.3	±2.6	44 glaciers
Shea et al (2009)	Ice	4.6	±0.6	9 glaciers

Table 6. Mean and standard deviation of degree-day factors from different studies. Units are mm d^{-1} K^{-1}.

4. Modelling glacier mass balance

According to above results, we can calculate the snow or ice ablation for any period and location if we know the degree-day total, which we calculate as the sum of positive temperatures at the same location and period. If measured data are not available, we can estimate them by extrapolation of temperatures from some location where they are measured, e.g. from a weather station at lower altitude in the same region as the glacier. However, Braithwaite (1984) notes that it is tedious to find and store daily temperatures, or sub-daily temperatures, if one only needs to add them up to form a single total, and he suggests a simple method for estimating monthly degree-day totals from monthly mean and standard deviation of temperature. To do this, we assume that temperatures within the month are normally distributed about the mean temperature with a standard deviation s, and the degree-day total is given by a numerical integration of temperatures above 0 °C multiplied by their probabilities. The area under the Normal probability curve that lies above 0 °C is the duration of positive temperatures in days.

The paper by Braithwaite (1984), combined with the values of degree-day factor proposed by Braithwaite and Olesen (1989), was quite literally "seminal" in that many workers later

developed glacier mass balance models where ablation is calculated from the degree-day approach. See papers by Reeh (1991), Huybrechts et al. (1991), Lauman and Reeh (1993), Jóhannesson (1993), Van der Veen (1999, pages 355-363), Braithwaite and Zhang (1999a and 2000), Braithwaite et al. (2002), De Woul and Hock (2005), Anderson et al (2006 & 2010), and Radic and Hock (2007 & 2011) for examples. Tarasov and Peltier (1997) describe the degree-day approach as *the standard methodology for parameterization of ablation over both glaciers and the Greenland ice sheet*. This might be a slight exaggeration but Carlov and Greve (2005) regard the method as sufficiently useful to merit them developing a more efficient algorithm for the many repeated calculations needed for very long-term simulations of the Greenland ice sheet. In some of these models, ablation is calculated directly from monthly mean temperature while other workers follow a variant of the method where monthly mean temperature is estimated from annual mean and annual amplitude of temperature. The latter approach, developed by Reeh (1991), is especially interesting as it links the degree-day approach to several empirical studies where workers plot annual ablation, equal to annual accumulation at the ELA, as a nonlinear function of summer mean temperature as first suggested by Ahlmann (1924) and extended by Ohmura et al (1992) and Braithwaite (2008).

4.1 Calculating ablation and snow accumulation

Figs 4a and 4b illustrate the performance of the Braithwaite (1984) model. Fig. 4a shows monthly snow and ice ablation as functions of monthly mean temperature for suitable values of degree-day factor. Fig. 4b actually shows the calculated probability of temperatures under 0 °C in the month but we can interpret this as the ratio of snow accumulation to total precipitation if precipitation rate is constant throughout the month. The choice of standard deviation s affects the precise shape of the curves (Braithwaite, 1984). Recent results from Greenland (Fausto et al, 2009) suggest that standard deviation might generally be somewhat lower than assumed in Fig. 4. I intend to revisit this issue myself in the near future, and to extend the Braithwaite (1984) model to explicitly include variations in daily temperature range, which is relatively small at high latitudes but may be quite large at lower latitudes.

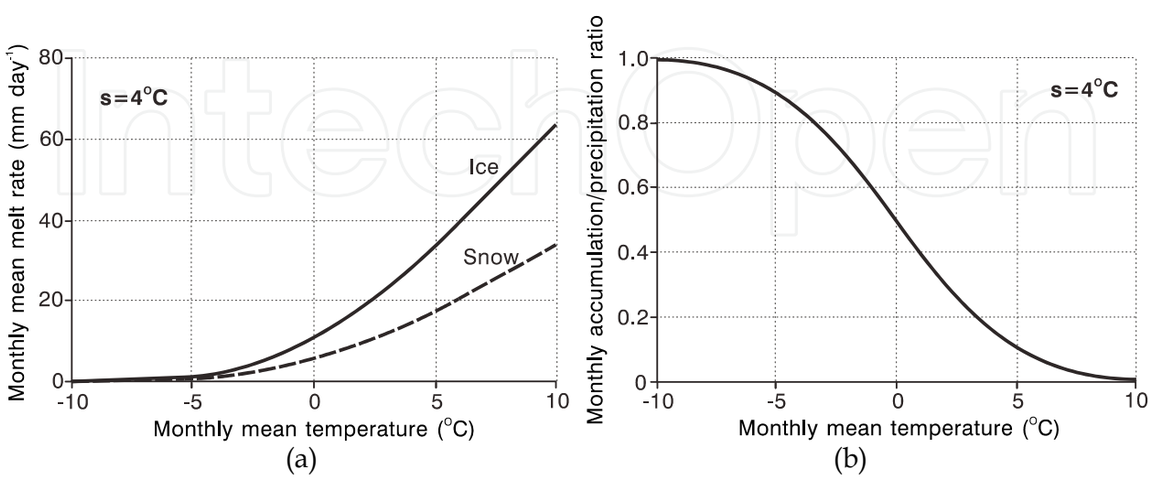

Fig. 4. Calculations of (a) monthly snow and ice ablation, and (b) monthly accumulation/precipitation ratio from the Braithwaite (1984) model.

Figs. 4a and 4b show that the range -10 to +10 °C in monthly mean temperature is the critical range for glacier-climate conditions. If monthly mean temperature is less than -10 °C in the warmest month, there will be no melting, which seems to be the situation over most of the Antarctic. As temperature rises there is a slightly exponential rise in melting as more and more days experience temperatures of over 0 °C and there is melting on every day in the month with monthly mean greater than 10 °C. Fig. 4b shows that there no days with below-freezing temperatures in months with mean temperature greater than 10 °C. This may explain the rough coincidence of the climatic tree line with the +10 °C July isotherm in the Northern Hemisphere, i.e. there is at least one frost-free month at the tree line.

4.2 Tuning the model

Fig. 5 sketches the operation of the glacier mass-balance model of Braithwaite and Zhang (1999a and 2000) and Braithwaite et al (2002). Monthly values of air temperature and precipitation are extrapolated to the glacier from a nearby weather station or gridded climatology (top left of the diagram). Degree-day factors for ice and snow are specified (very top right of figure). The model calculates a temporary value of annual balance (top right of figure) by summing monthly ablation and accumulation according to Figs 4a and 4b. The model then compares the computed mass balance with observed mass balance and adjusts the precipitation in successive small steps until the computed and observed mass balance agree closely. This represents the "tuning" of the model to fit observed mass balance (bottom right of figure) and we can now use the model for experiments to explore the sensitivity of glacier mass balance to changes in different factors (bottom left of figure).

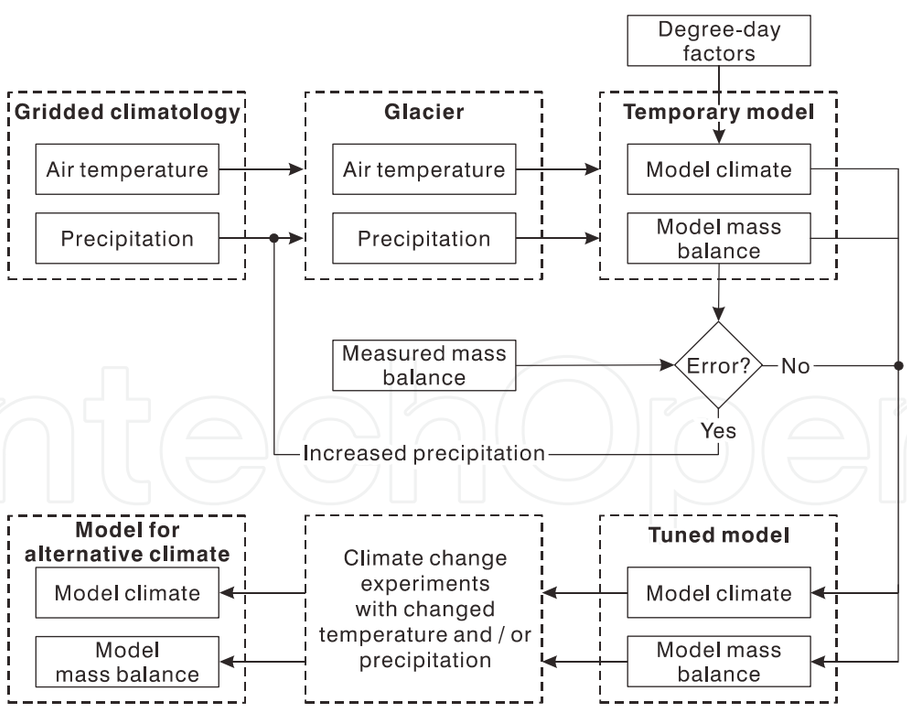

Fig. 5. Flow diagram illustrating the principles of a glacier mass-balance model based on the degree-day approach. From Braithwaite et al (2002).

We can tune the mass-balance model in Fig. 5 to fit an observed profile of mass-balance as a function of altitude. However, such data are probably only available for about 100 glaciers, i.e.

where workers have both made and published the necessary measurements. This limits our ability to apply the model but we can greatly extend it by noting that all we really need is a known value of mass-balance somewhere on the glacier. We do know that mass balance must be zero at the equilibrium line altitude (ELA) of the glacier and we can apply the model in Fig. 5 to the estimated ELA of the glacier. We can estimate this with an accuracy of about ±100 m for many thousands of glaciers using the "median" elevation parameter in the World Glacier Inventory (Braithwaite and Raper, 2009). This allowed Braithwaite and Raper (2007) to tune the mass-balance model for seven glacier regions with good coverage in the World Glacier Inventory, and then to extrapolate results to all mountain glaciers in the world to estimate 21st Century sea-level rise from melting glaciers (Raper and Braithwaite, 2006).

4.3 Temperature sensitivity of glacier mass balance

Common experiments on mass-balance models include changing air temperature by +1 °K, either throughout the whole year or just for the summer (June to August) , or changing annual precipitation by +10% while holding temperature constant. A number of sensitivity studies have shown that precipitation must increase by 20-40% to offset the effects of a +1 K temperature increase.

Fig. 6 shows the sensitivity of mass balance to a +1 K temperature change throughout the whole year for the seven glacier regions studied by Braithwaite and Raper (2007), who fitted the mass balance model (Fig. 5) to the estimated average ELA for each half-degree latitude/longitude grid cell in the region. The circles denote the average values for the N grid cells in each region, and the error bars denote standard deviations around these averages. As Braithwaite and Raper (2007) could not be completely certain of the correct values of degree-factors to use, they made calculations for low, medium and high estimates of the degree-factor for ice (6, 7 and 8 mm d^{-1} K^{-1}) in the hope that the true values will be somewhere within the range of results.

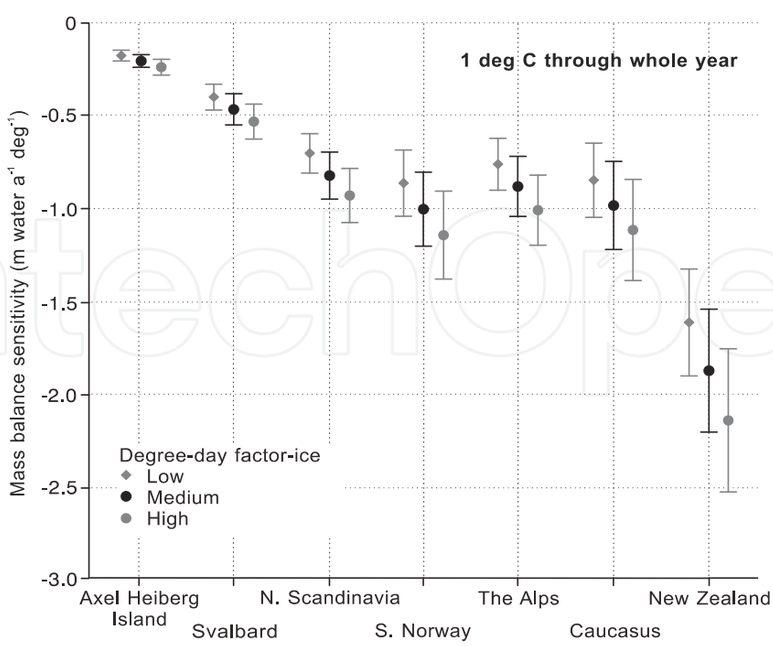

Fig. 6. Temperature sensitivity for glacier mass balance in seven regions calculated with a degree-day based mass-balance model. From Braithwaite and Raper (2007).

Braithwaite and Raper (2007) tuned the model with climate data from the gridded climatology of New et al (1997). This climatology is centred on averages for the 30-year period 1961-1990 so we can interpret Fig. 6 as a prediction of how mass balance will change in each region if the temperature increases by 1 K compared with the 1961-1990 averages. Despite uncertainties about degree-day factor, it is clear that mass-balance sensitivity is highly variable between regions, varying by almost an order of magnitude between Axel Heiberg Island, in the high arctic, and New Zealand. We can interpret this in terms of contrast between extreme continental and extreme maritime conditions as suggested by a number of workers. Svalbard is apparently somewhat less continental than Axel Heiberg Island while Northern Scandinavia, Southern Norway, the Alps and the Caucasus are relatively similar to each other on the continental/maritime scale.

For the purposes of the present Chapter, I take the results shown in Fig. 6 as a definite prediction that the mass balances of glaciers in the Alps will decrease by somewhat less than 1 m a^{-1} for each +1 K temperature change from the 1961-1990 average. I focus here on the Alps because it is the mountain region with best coverage of mass balance and climate data and I test this prediction with data from the Alps in the following sections.

You can regard the model in Fig. 5 as a method for "upscaling" daily measurements of ablation and air temperature, e.g. as described in Figs 2 to 5, to variations in annual balance. So, according to Fig. 6, degree-day factors of 6 to 8 mm d^{-1} K^{-1}, combined with a +1 K temperature increase, should be equivalent to a mass balance change of up to 1 m w.e. a^{-1} for Alpine glaciers.

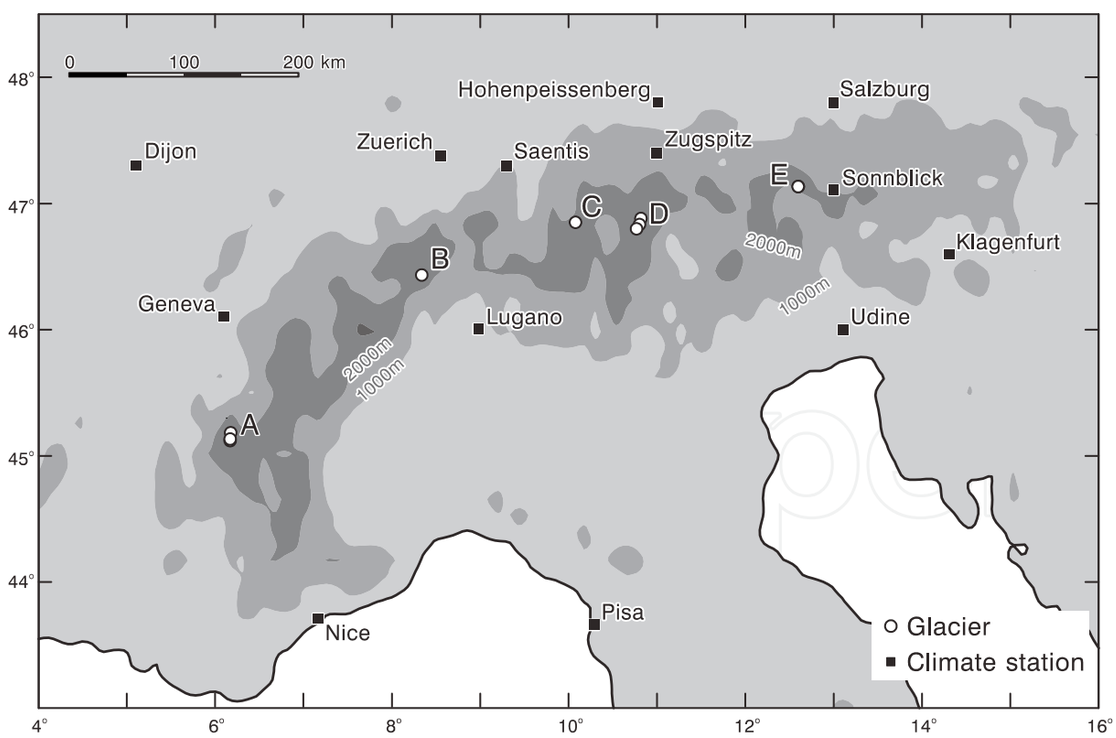

Fig. 7. Locations of glaciers and weather stations in the Alps with long records. (A) denotes Glacier de Sarennes and Glacier de Saint Sorlin; (B) denotes Griesgletscher; (C) denotes Silvrettagletscher; (D) denotes Hintereisferner, Kesselwandferner, Vernagferner; (E) denotes Sonnblikkees.

5. The Alps

The Alps occupy a key location in the middle of a continent with high levels of culture and education. It is no surprise therefore that the Alps, including its glaciers, have been a focus of scientific study for several centuries. However, actual measurement of glacier mass-balance by stakes and snow pits, i.e. the so-called "direct" glaciological method developed by H. W. Ahlmann (1889-1974), was relatively late coming to the Alps. For example, the longest continuous series is from Glacier de Sarennes that started in 1948 and continues today. Similar studies started at Limmern/Plattalvagletscher in 1948 but stopped in 1988. During the 1950's and 1960's mass-balance studies started on a number of glaciers and continue today.

The World Glacier Monitoring Service (WGMS, 2011) collects mass-balance data from the individual field workers and re-distributes them to potential users. For the present study, we need glaciers with full records throughout the whole 30-year period 1961-1990 and continuing up to nearly the present day, i.e. the balance year 2008/9 which is currently the latest year in the WGMS dataset. This is so we can compare recent mass-balance values with those in the climatic base period (1961-1990). I can identify mass-balance data for 37 Alpine glaciers although most of these have short records of only a few years. There are six glaciers with a full record for all years 1961-2009, and there are a further two glaciers that we can use because they have nearly full records, e.g. Griesgletscher for 1962-2009 and Vernagtferner for 1965-2009. Fig. 7 shows the location of these eight glaciers used in this study.

5.1 Mass-balance variations

Figure 8 shows year-to-year variations in mass balance for the eight glaciers. The circles denote the average balance for each year for the eight glaciers, or for six or seven glaciers for the very first years of record. The bars denote the corresponding standard deviations for each year, and give a measure of glacier-to-glacier variation.

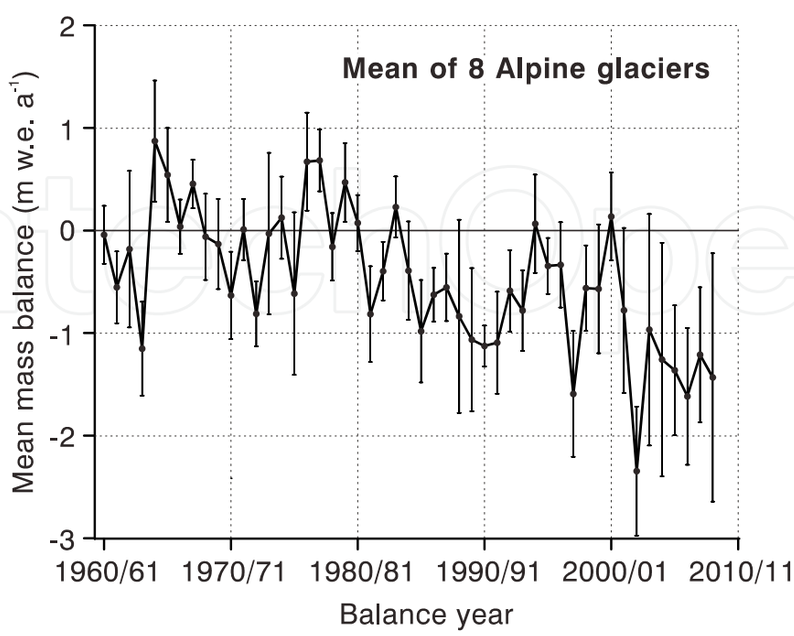

Fig. 8. Mass balance variations 1961-2009 for eight glaciers in the Alps with long records.

However, the year-to-year variations in balance are larger than these glacier-to-glacier variations, suggesting that the eight glaciers have very similar (but not identical) year-to-year variations. For example, principal component analysis (PCA) shows that a single principal component explains 80% of the total variance and a second component explains a further 10%. These two components are heavily loaded on mass balance for eastern and western Alpine glaciers respectively. The general agreement in year-to-year variations for the different glaciers agrees with Reynaud (1980) who suggested homogeneous mass-balance variations across the whole Alps.

Mass balance fluctuated around a constant base from 1961 to about 1985 and then becomes more negative in later years (Fig. 8). Positive balances occurred in just under half the years up to 1984 and then become very rare with positive balances only in 1995 and 2001. The highly negative balance in 2002/2003 is noteworthy as most of Europe suffered from a heat wave in the summer of 2003 (Schar et al, 2004). If we interpret the balance variations in Fig, 8 as mainly due to variations in ablation, it is obvious that ablation has increased markedly in recent years, especially after 2001. Does this increase in ablation agree quantitatively with an increase in air temperature according to our earlier hypothesis?

5.2 Temperature variations

The NASA/GISS (2011) website is a convenient source of temperature data from established weather stations. In and around the Alps, I could identify 13 weather stations (Fig. 7) covering long periods of record including the period of interest, 1961-2009. Temperature data are already available up to March 2011 (as of writing on 25 May 2011) but we only have glacier mass-balance data up to 2008/09 at present. Seven of the stations are noteworthy in that they have temperature records extending back well into the 19th Century. Three stations are located at 2500 m a.s.l. or above (Saentis, Zugspitz and Sonnblick) and may therefore reflect conditions near to the glacier ELA, i.e. roughly 2500-3000 m a.s.l. for the Alps. The 13 weather stations are situated at different locations and altitudes and obviously do not have the same temperatures. However, if we express temperatures at each station as deviations (or anomalies) from their 1961-1990 averages, the year-to-year variations in anomalies are remarkably similar. The three high-altitude stations also show very similar time variations to those at the ten lower-lying stations. These results show a pattern of very similar temperature variations over the whole region of the Alps in agreement with Bohm et al (2001).

Period	Summer	Mean	S.D.	Cases	%
1961-1990	June-August	0.00	±0.74	389	99.7 %
1991-2009	June-August	1.35	±1.12	239	96.8 %
Difference		1.35			
1961-1990	June-September	0.00	±0.75	388	99.5 %
1991-2009	June-September	1.02	±0.99	239	96.8 %
Difference		1.02			
1961-1990	May-September	0.00	±0.73	389	99.7 %
1991-2009	May-September	1.11	±0.89	239	96.8 %
Difference		1.11			

Table 7. Mean and standard deviation (S.D.) of summer mean temperatures (°C) at 13 weather stations for different lengths of summer.

Table 7 shows the mean and standard deviation of temperature anomalies for the 13 weather stations for the 30-year period 1961-1990 and for the 19 years 1991-2009. The mean temperature anomaly for 1961-1990 is obviously zero by definition of "anomaly". The column marked "%" denotes the percentage of expected data that are actually present. The 1961-1990 record is almost complete for the expected 30 × 13 = 390 records except for one missing month at one station. There are slightly more missing data for the later period. The standard climate definition of "summer" is June-August but Alpine glaciers generally have a longer melt season (Braithwaite et al, 2002) and I therefore calculated the anomalies for different lengths of summer to see if this is a critical issue. For any sensible choice of summer length, the period 1991-2009 is clearly on average 1.02 to 1.11 °C warmer than the base period 1961-1990. A three-month summer (June to August) is too short to cover the full melting period on a typical Alpine glacier which is more like 120-150 days, i.e. 4 to 5 months. I therefore adopt the average of temperature changes for 4- and 5-month summers, i.e. +1.07 K, for the analysis in the next section.

5.3 Temperature sensitivity of Alpine glacier mass balance

Table 8 shows the mean and standard deviation of mass balance for the eight glaciers and for the two periods 1961-1990 and 1991-2009. For all eight glaciers, the mean mass balance for the second period is much more negative than for the first period.

Glacier	1961-1990 Mean	1961-1990 S.D.	1961-1990 N	1991-2009 Mean	1991-2009 S.D.	1991-2009 N	Diff.	$\Delta b/\Delta T$
	m a^{-1}	m a^{-1}		m a^{-1}	m a^{-1}		m a^{-1}	m $a^{-1}K^{-1}$
Saint Sorlin	-0.20	±0.83	30	-1.31	±1.04	19	-1.11	-1.04
Sarennes	-0.57	±0.90	30	-1.64	±1.29	19	-1.08	-1.01
Griesgletscher	-0.35	±0.76	29	-1.13	±0.68	19	-0.78	-0.73
Silvrettagletscher	+0.03	±0.71	30	-0.60	±0.66	19	-0.63	-0.59
Hintereisferner	-0.33	±0.56	30	-0.98	±0.44	19	-0.65	-0.61
Kesselwandferner	+0.05	±0.40	30	-0.34	±0.49	19	-0.40	-0.37
Vernagtferner	-0.09	±0.41	26	-0.70	±0.47	19	-0.62	-0.58
Sonnblickkees	-0.08	±0.75	30	-0.77	±0.87	19	-0.69	-0.65
Mean	-0.19			-0.93			-0.74	-0.70
S.D.	±0.21			±0.42			±0.24	±0.23

Table 8. Mean and standard deviation (S.D.) of mass balance for eight glaciers for two different periods. Diff. is the average between means for the two periods and $\Delta b/\Delta T$ is the estimated mass-balance sensitivity assuming a temperature change of +1.07 K.

The standard deviation of mass balance in the period 1961-1990 varies greatly from relatively high values in the western Alps (Saint Sorlin and Sarennes), medium values in the central Alps, low values in the eastern Alps and a medium value in the far-eastern Alps. This pattern reflects the different maritime/continental character of the different glaciers as lower standard deviations of mass balance occur in more continental environments (Braithwaite and Zhang, 1999b)

We can divide the differences in mass balance for the two periods by the estimated increase in mean temperature between the two periods, estimated to be +1.07 K, to obtain estimates

of the mass balance sensitivity. These range from relatively large values, i.e. around -1.0 m a^{-1} K^{-1}, for the two western glaciers, to a relatively small value for Kesselwandferner, i.e. about -0.4 m a^{-1} K^{-1}. The overall average mass balance sensitivity for the eight glaciers is about -0.7 ± 0.23 m a^{-1} K^{-1}. This is in good agreement with the range of values predicted by the model for the Alps (Fig. 6) for low-medium degree-day factor (6-7 mm d^{-1} K^{-1}) but somewhat smaller than predicted for high degree-day factor (8 mm d^{-1} K^{-1}).

The model predictions in Fig. 6 are averages for all half-degree latitude/longitude grid cells in the Alps, i.e. covering the whole range of Alpine climates, while the eight glaciers are only a fairly small sample and may biased to more continental glaciers. However, there is also a methodological reason why the degree-day model should somewhat overestimate mass balance changes. This is because we calibrate the model with data for present-day glaciers and then change the temperature by +1 K while keeping the present-day area distribution of the glacier. In the modelling literature, we call the resulting mass-balance sensitivity the "static sensitivity". However, as a result of the increased melting, glacier areas in the second period are already smaller than in the first, thus reducing the negative mass balance somewhat. Given enough time, without any further temperature increase, the glaciers will arrive at a new equilibrium with zero mass balance. The time scale for this to occur for Alpine glaciers is of the order of 10^2 years (Raper et al, 2000: Raper and Braithwaite, 2009).

In the above discussion, I have attributed the whole difference in mass balance for the two periods to the effects of changing temperature. However, precipitation changes may be responsible for a part of the observed mass balance change, although sensitivity studies with the degree-day model (Braithwaite et al. 2002: Braithwaite and Raper, 2007) suggest this can only be a relatively small part. If we wanted to verify this empirically, we would have to use precipitation data at much higher spatial resolution than the temperature data used here as correlation distances for precipitation are much shorter than those for temperature.

6. Conclusion

The increasingly negative mass balance for Alpine glaciers and the recent rise in temperature in and around the Alps should be no surprise to the reader as both have been reported by other workers. It is, however, noteworthy that the temperature-sensitivity of the mass-balance change is in good agreement with the prediction in Fig. 6. This implies that the range of degree-day factors used in the model (6-7 mm d^{-1} K^{-1}) is also valid for the Alps. These parameters are inferred from daily ablation-temperature correlations in the high arctic and in Greenland (Tables 2 to 5) and from secondary data, covering a wider range of geographical conditions (Table 6). In other words, we can explain a recent trend to increasingly negative mass balance in the Alps in terms of a theory based on measurements made at other places, and even a few decades ago.

If we accept the validity of the degree-day model, which the results seem to demonstrate, we can be confident that high rates of melting on Alpine glaciers will continue in the future as long as temperatures continue to rise as predicted by the theory of global warming. This is not a conclusion that pleases me as it means that Alpine glaciers will largely disappear in the coming century. On the other hand, if temperatures do not rise any further, for any reason, the mass balances of Alpine glaciers will tend towards zero as the glaciers tend to a new equilibrium.

In strict logic, the above conclusion only applies to Alpine glaciers, and I should test the predictions for other areas, e.g. as shown in Fig. 6. Northern Scandinavia and southern Norway have relatively good coverage of mass balance records so I might be able to use the same approach as here. Other areas, with more restricted data, may need more flexibility. For example, instead of using a 30-year reference period (1961-1990) it might be possible to use a shorter reference period for which there may be more mass-balance records. Aside from glaciers with mass-balance measurements from stakes and snow pits, we could also apply the present approach to glaciers where longer-term mass changes are available from geodetic methods, including increasingly high precision survey by satellites.

7. Acknowledgment

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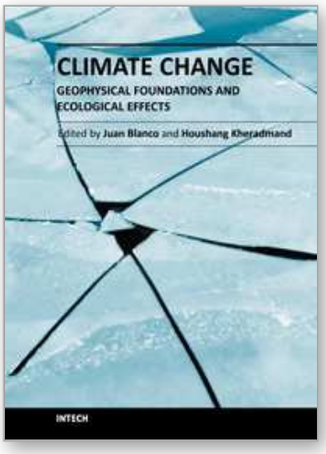

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This book offers an interdisciplinary view of the biophysical issues related to climate change. Climate change is a phenomenon by which the long-term averages of weather events (i.e. temperature, precipitation, wind speed, etc.) that define the climate of a region are not constant but change over time. There have been a series of past periods of climatic change, registered in historical or paleoecological records. In the first section of this book, a series of state-of-the-art research projects explore the biophysical causes for climate change and the techniques currently being used and developed for its detection in several regions of the world. The second section of the book explores the effects that have been reported already on the flora and fauna in different ecosystems around the globe. Among them, the ecosystems and landscapes in arctic and alpine regions are expected to be among the most affected by the change in climate, as they will suffer the more intense changes. The final section of this book explores in detail those issues.

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