Climate Dynamics (1998) 14:267–274

J. Oerlemans · B. Anderson · A. Hubbard · Ph. Huybrechts · T. Jóhannesson · W.H. Knap · M. Schmeits · A.P. Stroeven · R.S.W. van de Wal · J. Wallinga · Z. Zuo

Modelling the response of glaciers to climate warming

Received: 30 June 1997/Accepted: 21 October 1997

Abstract Dynamic ice-flow models for 12 glaciers and ice caps have been forced with various climate change scenarios. The volume of this sample spans three orders of magnitude. Six climate scenarios were considered: from 1990 onwards linear warming rates of 0.01, 0.02 and 0.04 K a^{-1}, with and without concurrent changes in precipitation. The models, calibrated against the historic record of glacier length where possible, were integrated until 2100. The differences in individual glacier responses are very large. No straightforward relationship between glacier size and fractional change of ice volume emerges for any given climate scenario. The hypsometry of individual glaciers and ice caps plays an important role in their response, thus making it difficult to generalize results. For a warming rate of 0.04 K a^{-1}, without increase in precipitation, results indicate that few glaciers would survive until 2100. On the other hand, if the warming rate were to be limited to 0.01 K a^{-1} with an increase in precipitation of 10% per degree warming, we predict that overall loss would be restricted to 10 to 20% of the 1990 volume.

J. Oerlemans (✉) · W.H. Knap · M. Schmeits ·
R.S.W. van de Wal · J. Wallinga · Z. Zuo
Institute for Marine and Atmospheric Research, Utrecht University, Princetonplein 5, The Netherlands

B. Anderson · A.P. Stroeven
Department of Physical Geography, Stockholm University, Stockholm, Sweden

A. Hubbard
Department of Geography, University of Edinburgh,Drummond Street, Edinburgh EH8 9XP, Scotland

P. Huybrechts
Department of Geography, Free University of Brussels, Pleinlaan 5, Brussels, Belgium

T. Jóhannesson
Icelandic Meteorological Office, Bústaðavegi 9, IS-150 Reykjavik, Iceland

1 Introduction

Valley glaciers affect human activities in a significant way in many mountainous regions. People benefit from the presence of glaciers by using meltwater for hydropower reservoirs and irrigation systems. Glacier ice has been used extensively for cooling of cellars and caves. Exploitation of glaciers for tourism is also an important factor in some local economies.

On the other hand, even in the recent past ice avalanches and outbursts of glacier-dammed lakes have caused large catastrophes (e.g. Tufnell 1984). More regular fluctuations of glacier extents may also threaten roads, constructions and property.

Although valley glaciers and small ice caps contain only a small amount ($\sim$0.5 m of sea-level equivalent, Warrick et al. 1996) of all land ice on earth, it is believed that they contribute significantly to sea-level fluctuations on a century time scale (Meier 1984; Warrick et al. 1996). The reason lies in their relatively short response times associated with large mass turnover.

From the point of view of the climatologist, historic records of glacier length contain valuable information on past climate. There is no doubt that valley glaciers are sensitive to climate change. Records of glacier fluctuations contain many examples of changes in glacier length of several kilometers within the last century. Most likely, such changes are associated with secular temperature fluctuations within the range of 1 °C (Grove 1988; Oerlemans 1994). The reason for this large sensitivity is found in the nature of the melting process. Because the melting point is fixed, both the downward sensible heat flux and the longwave radiation balance, increase when air temperature rises. There are no compensating effects like increasing longwave emission (which would occur with the energy balance of a soil layer, for instance).

Glaciers of different geometry, located in different climatic regimes, will react in different ways to a climatic signal (see Kuhn 1985 for a well-documented example). This makes the interpretation of glacier records from a global selection of glaciers more difficult. Moreover, it is unclear how the behaviour of glaciers can be generalized to make inferences about all glaciers on earth. Despite this, when looking at the possible implications of global warming and concommitant glacier retreat on sea level, the issue needs to be considered.

In an attempt to address this problem, methods have been proposed to deal with all glaciers "in a single equation" (e.g. Oerlemans 1989; Wigley and Raper 1993). Essentially, these equations attempt to account for the fact that under any warming scenario there will be:

1. An immediate response of surface melt and runoff
2. A changing glacier geometry affecting the mass balance characteristics
3. An overall decrease of glacier area and volume

Apart from this, it has been noted that the climate sensitivity of glaciers varies over at least one order of magnitude, mainly depending on the annual mean precipitation (where the continental glaciers are less sensitive and the maritime glaciers more sensitive, Oerlemans and Fortuin 1992). This difference in climate sensitivity is related to different energy-balance characteristics as well as different hypsometry.

Of similar importance is the difference in response time. Subpolar ice caps will respond much more slowly to climate change than small valley glaciers. Glacier size, steepness of the bed, mass turnover and hypsometry are known to play a role.

Ideally, all these factors should be taken into account in any global model of glacier change. However, there is a lack of basic information with which to construct such a model. Data are required on the dynamic response of glaciers and small ice caps which have contrasting geometry and climatic settings. Also, glacier inventories, which exist in detail for some regions, are globally sparse (Haeberli et al. 1989). In particular, distribution of glaciers over classes of area size would be very helpful, but exists only for a few glaciated regions.

Handling all glaciers with a simple formulation may be possible, but so far little justification has been provided for such an approach. However now that several numerical models have been developed that treat individual glaciers, it appears worthwhile to carry out a set of identical numerical experiments with all these models. This will help validate and further identify the direction that such a generalized equation should take.

In this work we present results from 12 glaciers and small ice caps located in Europe, New Zealand and the Antarctic Peninsula. Models of these glaciers have been run to investigate the specific response to a set of climate change scenarios. These scenarios are fairly simple: we have imposed constant warming rates of 0.01, 0.02 and 0.04 $\mathrm{K a^{-1}}$ for the period 1990–2100 AD. The experiments have been repeated with an additional change in precipitation of 10% per degree warming. We refer to these six scenarios as 0.01, 0.01^+, 0.02, 0.02^+, 0.04 and 0.04^+.

2 Glaciers and models

Table I provides a listing of the glaciers used in this study. In terms of volume and area three orders of magnitude are covered. The amount and type of data available varies considerably for each glacier and thus some models have been calibrated more thoroughly than others. In Table I references are given where more details can be found. Most of the glaciers listed are valley glaciers. However, Blöndujökull and Illviðrajökull are sectors of an ice cap, whereas KGI ice cap applies to the entire ice cover on King George Island in Antarctica. All the glaciers and ice caps are considered to be essentially temperate with calving occurring only at KGI. In the model for KGI ice cap calving was

Table 1 Glaciers and ice caps studied. Blöndujökull and Illviðrajökull are part of Hofsjökull. KGI ice cap is King George Island ice cap, which we use as a name for the entire land ice cover on King George Island, Antarctica Peninsula. Labels denote: # historic length record available (see last column for first data point), $ mass balance observations available, ($) limited mass balance measurements available (a few years and/or covering a small part of the glacier only). Ice volumes given are modelled volumes and may deviate somewhat from real volumes

Glacier/ice cap	Reference	Area ($\mathrm{km^{2)}}$)	Ice volume ($\mathrm{km^{3)}}$)	Ela (m a.s.l.)	First info (year AD)
Franz Josef Glacier (New Zealand) #	Oerlemans (1997a)	34	4.89	1650	1750
Glacier d'Argentière (France) # ($)	Huybrechts et al. (1989)	15.6	1.91	2900	1590
Haut Glacier d'Arolla (Switzerland) # $	Hubbard (1996)	6.3	0.33	3200	1933
Hintereisferner (Austria) # $	Greuell (1992), Van de Wal (unpubl.)	7.4	0.44	2950	1770
Nigardsbreen (Norway) # $	Oerlemans (1986, 1997b)	48	3.8	1550	1700
Pasterze (Austria) # ($)	Zuo and Oerlemans (1997)	19.8	2.62	2880	1605
Rhonegletscher (Switzerland) # $	Wallinga and Van de Wal (unpubl.)	17.7	2.58	2930	1602
Storglaciären (Sweden) # $	Stroeven (1996)	3.1	0.30	1460	1897
Unt. Grindelwaldgl. (Switzerland) #	Schmeits and Oerlemans (1997)	21.7	1.83	2770	1534
Blondujökull (Iceland) ($)	Jóhannesson (1997)	226	46.9	1300	recent
Illvidrajökull (Iceland) ($)	Jóhannesson (1997)	116	25.8	1250	recent
KGI ice cap (Antarctica) ($)	Knap et al. (1996)	1402	155	100	recent

treated in a very simple way: ice thickness was simply set to zero at the coastline (Knap et al. 1996).

Although the models are not identical, the majority are based on the assumption that ice velocity is determined by the local driving stress (e.g. Paterson 1994). The prognostic equation is the vertically integrated continuity equation, describing how the change in ice thickness H is related to flux divergence and specific balance B (e.g. Oerlemans and Van der Veen 1984):

$$\frac{\partial \mathrm{H}}{\partial \mathrm{t}} = -\nabla \cdot (\mathrm{HU}) + \mathrm{B} \tag{1}$$

Here **U** is the vertical mean velocity vector parallel to the bed. In all models, velocity (sliding, deformation, or both lumped together) is determined by the "local driving stress F", which is given by

$$\mathbf{F} = -\rho \mathrm{gH} \nabla \mathrm{h} \tag{2}$$

In this equation ρ is ice density, g acceleration due to gravity, and h surface elevation. This means that ice always flows along the gradient of surface elevation. For KGI ice cap these equations are applied on a 2-dimensional grid. For all other glaciers an approach with a flow line is used, in which the 2-dimensional geometry is parameterized.

The mass balance is treated in different ways. For glaciers where good mass balance observations and climate data exist, calibrated mass balance models (based on degree-day or energy balance considerations) are employed. In other cases a more schematic approach had to be taken. For details the reader is referred to the papers listed in Table 1. For most glaciers it was possible to calibrate the flow models with historic records of glacier length (marked # in Table 1). Some of these records go far back in time, as indicated in the last column of the table. Calibration with the historic length record was done by seeking a mass balance history in such a way that an (almost) perfect match between observed and simulated length was achieved. As an example, Fig. 1 shows how this procedure is applied to Nigardsbreen. In this case a reference mass balance profile has been perturbed by a value $\delta B(t)$ independent of altitude. From 1962 onwards mass balance observations are available and they have been used and extended back in time by a δB history that leads to a good simulation of the historic record. In this way the mass-balance fluctuations preserved in the memory of the glacier can be used to influence the future behaviour. More details about this procedure (called "dynamic calibration") can be found in Oerlemans (1997b). For a basic discussion on inferring a mass balance history from glacier front variations the reader is referred to Nye (1965).

Historic length variations of the glaciers for which dynamic calibration was made are shown in Fig. 2. Note that most glaciers were retreating in the period 1850–1980. Around 1980 some of the glaciers in the

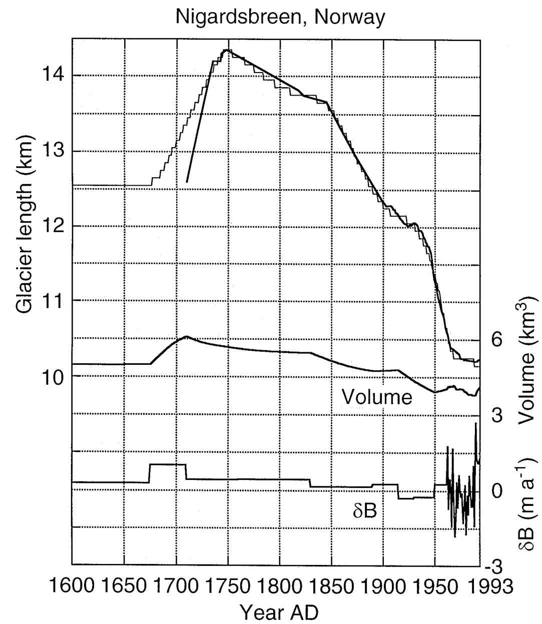


Fig. 1 Illustration of the dynamic calibration procedure for Nigardsbreen (from Oerlemans 1997b). A mass balance history δB is derived in such a way that simulated glacier length (*stepped curve*) matches the observed record (*solid line*). The *lower part* of the figure (*scale at right*) shows the mass balance curve. Mass balance observations are used from 1962 onwards

sample stopped retreating or started to advance (Franz Josef Glacier, Storglaciären, Nigardsbreen, Untere Grindelwaldgletscher). After calibration with the observed records the models have been integrated in time until 2100, assuming that climatic conditions (in terms of mass balance) remain the same as in the period 1961–1990. Note that this "climatological mass balance" is largely determined by the calibration (except for Nigardsbreen and Storglaciären where mass balance observations exist for 28 years and 30 years respectively of this 30-year period). In the next century, Franz Josef Glacier and Untere Grindelwaldgletscher would grow significantly. Most glaciers in the sample, however, would continue to retreat, indicating that over the period 1961–1990 these glaciers were larger than the equilibrium size corresponding to the prevailing climatic conditions.

This 30-y period has been taken as a reference for all glaciers. Neverthless, it should be realized that for some glaciers this may have been a relatively warm period in the long-term mean, for others a cold one.

For Blöndujökull, Illviðrajökull and KGI ice cap dynamic calibration in the sense described was not possible. All integrations with models for these ice caps start with an equilibrated geometry for the year 1990. For Blöndujökull and Illviðrajökull there is evidence that these ice cap sectors have been almost steady for several decades, but for KGI ice cap this information is not available.

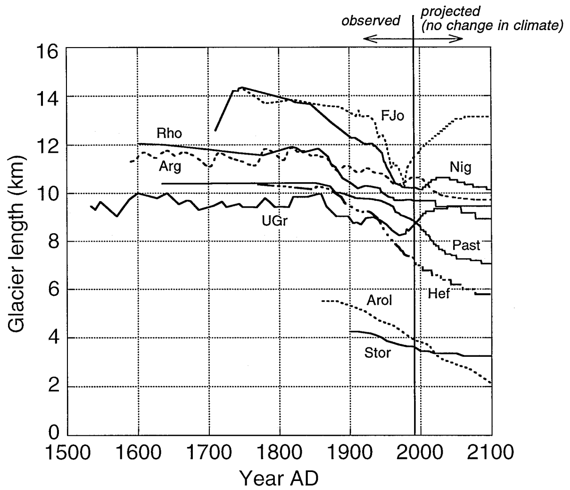


Fig. 2 Records of glacier length used for calibration. Note that all glaciers shown here retreated substantially during the last 100 y. The projections (1990 – >) in this graph have been made with the assumption that, for each individual glacier, climate conditions will be constant and equal to the average climate conditions over the period 1961–1990. Abbreviations: *FJo*, Franz Josef Glacier (New Zealand); *Rho*, Rhonegletscher (Switzerland); *Arg*, Glacier d'Argentière (France); *Nig*, Nigardsbreen (Norway); *Ugr*, Untere Grindelwaldgletscher (Switzerland); *Past*, Pasterzen Kees (Austria); *Hef*, Hintereisferner (Austria); *Arol*, Haut Glacier d'Arolla (Switzerland); *Stor*, Storglaciären (Sweden)

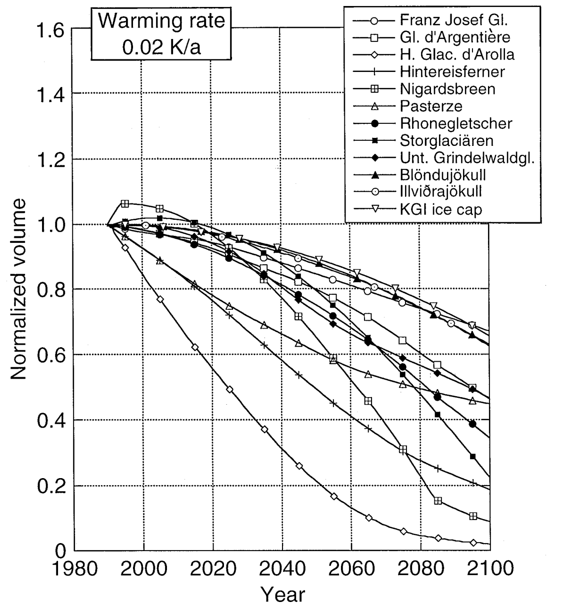


Fig. 3 Ice volume for the experiment with a warming rate of 0.02 K a^{-1} without a change in precipitation. Volume is normalized with the 1990 volume. Initial states are *not* steady states, except for Blöndujökull, Illviðrajökull and KGI ice cap

3 The 0.02 K/a climate change experiment

As explained in the introduction, six climate change scenarios have been used to investigate the possible response of the modelled glaciers to global warming. Here we present a selection of results.

Figure 3 shows glacier volume for the 0.02 K/a warming experiment. Volume has been scaled using the 1990 volume. As expected, the differences are large. Haut Glacier d'Arolla, a small glacier already in a strongly retreating mode, will virtually have disappeared by the year 2060. In general the smaller glaciers lose relatively more mass. By the year 2100, the large glaciers and ice caps have typically lost 40% of their volume. Nigardsbreen is an exception. In spite of its size and the fact that the glacier is currently advancing, it loses a large amount of mass. This is related to the characteristic hypsometry of Nigardsbreen: the bulk of the glacier is resting on a plateau. Consequently, a modest warming turns a large part of the accumulation zone into an ablation zone. This example illustrates the difficulty involved in generalizing an all-embracing glacier response to climate change.

It is striking that the ice cap (sectors) considered here, Blöndujökull, Illviðrajökull, and KGI ice cap all behave similarly: by 2100 they all lose about 35% of their volume. It should be noted that integrations for these ice caps start from an equilibrium state. Nevertheless, the fractional loss of ice follows a remarkably similar pattern.

Figure 4 shows the difference between the 2100-volume and the 1990-volume. Although they have the smallest fractional losses, the ice caps dominate in the overall picture of ice wastage. Nigardsbreen shows up as the valley glacier with the largest loss.

To analyze further the behaviour of the different glaciers, we define a static and a dynamic sensitivity of a glacier (its mass balance, rather). The static sensitivity to temperature S_T is defined as

$$S_T = \frac{\partial B_m}{\partial T} \approx \frac{B_m(+1K) - B_m(-1K)}{2} \text{ or}$$

$$\approx B_m(+1K) - B_m(0K) \qquad (3)$$

Here $B_m(x)$ is the mean specific mass balance corresponding to a temperature perturbation x. It is defined for the 1990 glacier geometry (i.e. not necessarily for a steady state). The static sensitivity can be used to make a first estimate of change in glacier volume for a given temperature scenario. This is achieved by multiplying S_T with glacier area and with the integral of the temperature perturbation over the time period considered. The procedure assumes that the effect of changing geometry on mass balance can be ignored (fixed-geometry approach). For longer time periods this does not hold and further insight into the significance of geometric effects can be obtained by a comparison of the static sensitivities S_T with the dynamic sensitivities

D_T (see also Jóhannesson 1997). This quantity is defined by comparing the volume of two states of a glacier:

$$D_T(t) = \frac{V(t) - V(t_0)}{A(t_0)\,(t - t_0)\,\bar{T}'} \tag{4}$$

Here the overbar denotes a mean value over the period considered $(t - t_0)$. V and A are glacier volume and area, respectively. It is evident that D_T has strong dependence on time. D_T can either be smaller or larger than the static sensitivity, depending on how far the reference state is out of balance. When $t - t_0$ is increased, a point will be reached where D_T becomes significantly smaller and the fixed-geometry approach becomes invalid. Values of D_T for the 0.02 scenario have been calculated for all glaciers for 2050 and 2100 AD. The results are shown in Fig. 5, together with values for the static sensitivity S_T. For glaciers that are currently retreating rapidly, i.e. for glaciers that are too large for the climatic state over the last 30 y, values of D_T for the year 2050 are much larger than values of S_T. On the basis of the definition of D_T, this is to be expected. The effect is seen best for Pasterze and Haut Glacier d'Arolla. Franz Josef Glacier displays opposite behaviour, as it is currently advancing.

Given the ideal condition of a glacier both being in equilibrium with climate and subsequent change in its geometry playing an insignificant role, then the resulting static and dynamic sensitivities would be identical. Thus, when considering KGI ice cap, Illvidrajökull and Blondujökull, which are all in equilibrium in 1990, given that the dynamic sensitivities do differ significantly, then even for these relatively large ice masses, geometric effects become important after a few decades. Typically, the dynamic sensitivity in 2100 is about 2/3 of the static sensitivity.

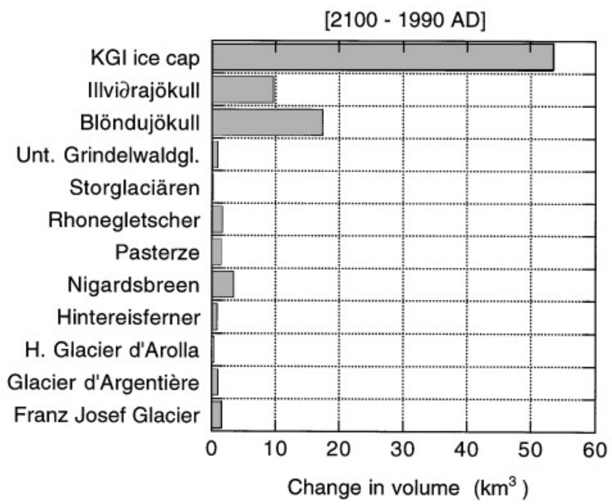


Fig. 4 Absolute change in volume (km^3) for the 0.02 $K\,a^{-1}$ warming experiment without a change in precipitation. The difference between the 2100 and 1990 volumes, is shown

4 Scaling

As noted in the Introduction, one of the long-term goals of this project is to find ways to deal with all glaciers in a simplified scheme. The current material is too limited to generalize reliably. Nevertheless, it is still instructive to look at the overall response characteristics of the entire sample of glaciers considered here.

Normalized volume response may be defined in several ways. One method is to simply average all normalized response curves as shown in Fig. 3. We refer to the resulting quantity as $\langle V_{sc} \rangle$ (sc = scaled with 1990 volume; $\langle\rangle$ is mean over sample). If the resulting curve is considered to represent all glaciers in the world, the implicit assumption would be that the total glacier area in each "size class" as represented by the glaciers in Table 1 is equal. On the other hand, all volume changes can be totalled and the resulting curve scaled with the

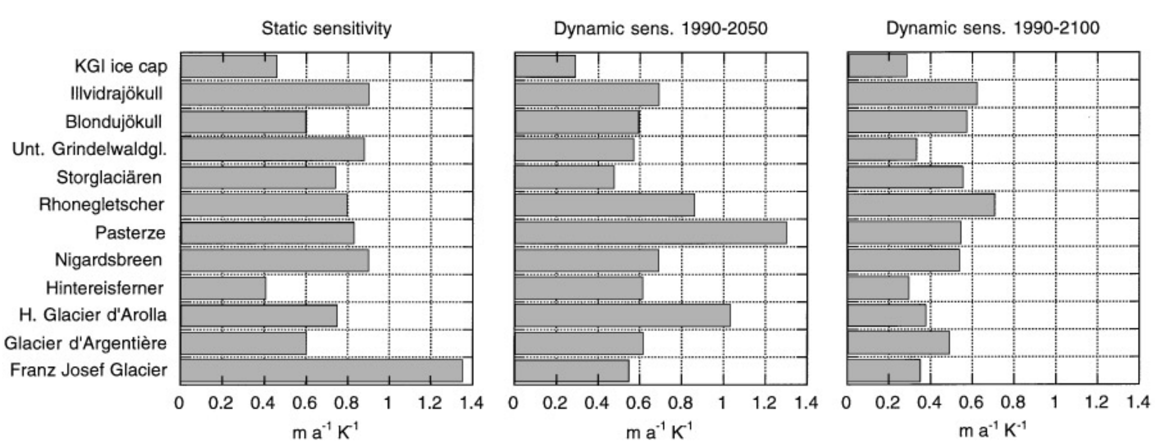


Fig. 5 An analysis of the sensitivity to temperature change. The static sensitivity is defined for the 1990 glacier geometries. See text for definitions of static and dynamic sensitivity

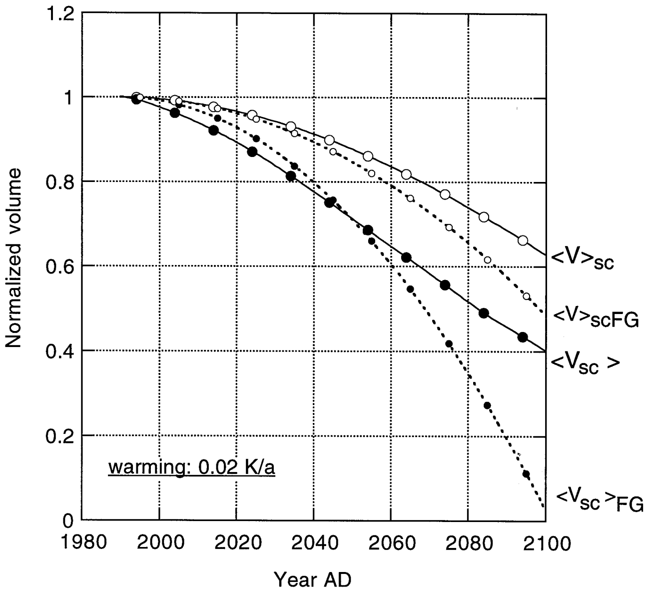


Fig. 6 Normalized volume for the 0.02 K a^{-1} warming experiment. The *curve* labelled $\langle V\rangle_{sc}$ shows total volume scaled with the 1990 total volume. This curve is dominated by the ice caps. The *curve* labelled $\langle V_{sc}\rangle$ shows the mean of the scaled volume of the individual glaciers. In this curve all glaciers in the sample have equal weight. Also shown are the associated results from a fixed-geometry approach, in which loss of ice volume is calculated from the static sensitivity (*FG*)

total 1990 volume. The method results in a value $\langle V\rangle_{sc}$ which is predominantly determined by the the KGI ice cap, Illvidrajökull and Blondujökull.

Figure 6 shows $\langle V_{sc}\rangle$ and $\langle V\rangle_{sc}$ for the period 1990–2100. As expected, the difference is large. Unfortunately, area class distributions are available for limited regions only (Haeberli et al., 1989), so it is impossible to validate either of the resulting curves. Also shown in Fig. 6 are the results obtained with the fixed-geometry approach, i.e. by ignoring the changing geometry of the glaciers as explained in the previous section. For an individual glacier, the ice volume then is defined by:

$$V(t) = V(t_0) - S_T A(t_0) \int_{t_0}^{t} T'\, \mathrm{d}t. \tag{5}$$

For an assumed temperature perturbation linear in time $[T' = \alpha(t - t_o)]$ the integral can be evaluated and the normalized ice volume becomes:

$$\frac{V(t)}{V(t_0)} = 1 - \frac{\alpha S_T A(t_0)(t - t_0)^2}{2\,V(t_0)}. \tag{6}$$

Averaging over the entire sample of glaciers (giving equal weights) yields:

$$\langle V_{sc}\rangle_{FG} = \frac{1}{12}\sum_{i=1}^{12} \frac{V_i(t)}{V_i(t_0)}$$

$$= \frac{1}{12}\sum_{i=1}^{12}\left\{1 - \frac{\alpha S_{Ti} A_i(t_0)(t - t_0)^2}{2V_i(t_0)}\right\}$$

$$= 1 - \frac{\alpha(t - t_0)^2}{24}\sum_{i=1}^{12}\frac{S_{Ti}A_i(t_0)}{V_i(t_0)} \tag{7}$$

The index *FG* refers to fixed geometry.

Similarly, to obtain a curve that can be compared to $\langle V\rangle_{sc}$, we define:

$$\langle V\rangle_{scFG} = \frac{\sum_{i=1}^{12} V_i(t)}{\sum_{i=1}^{12} V_i(t_0)}$$

$$= \frac{\sum_{i=1}^{12}\{V_i(t_0) - \frac{1}{2}\alpha S_{Ti}A_i(t_0)(t - t_0)^2\}}{\sum_{i=1}^{12} V_i(t_0)}$$

$$= 1 - \frac{\frac{1}{2}\alpha(t - t_0)^2 \sum_{i=1}^{12}\{S_{Ti}A_i(t_0)\}}{\sum_{i=1}^{12} V_i(t_0)}. \tag{8}$$

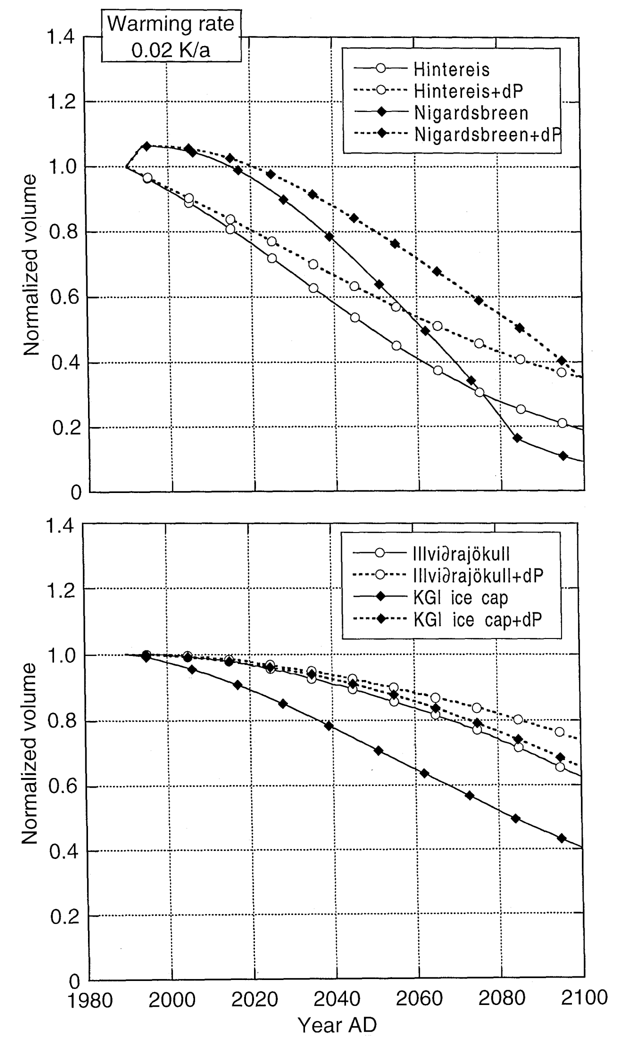


Fig. 7 The effect of an increase in precipitation (10% per degree warming) on the loss of ice for the 0.02^{+} K a^{-1} warming experiment. Results are shown for two glaciers (Hintereisferner, Nigardsbreen) and two ice caps (Illviðrajökull, KGI ice cap) only

For the 0.02 experiment the quantities defined in Eqs. (7) and (8) have been calculated and are also shown in Fig 6.

It becomes evident that after about 70 y the fixed-geometry approach starts to give significant errors. For the case of scaled total volume ($\langle V \rangle_{sc}$), the fixed-geometry approach always overestimates the loss of ice. Because this case is dominated by the voluminous ice caps, which are in equilibrium in 1990, this is understandable. However, for the case of mean scaled volume ($\langle V_{sc} \rangle$), the situation is different. Clearly, the fixed-geometry approach underestimates the loss of ice until about 2050. This is caused by the fact that, on average, the glaciers in the sample are retreating at the start of the integration. Thus the effect of the dynamic calibration procedure is clearly visible here. The curves in Fig. 6 show that it is possible to optimize the fixed-geometry approach by choosing lower values for the static sensitivity.

5 The effect of changes in precipitation

The effect of a changing climate on glaciers will not operate solely through temperature. Precipitation, humidity and cloudiness are other potentially important factors. In this study we also consider changes in precipitation. Spatial and temporal variability in precipitation is large, and one cannot assume that GCM output for greenhouse warming scenarios is very reliable on the regional scale. Most models indicate an intensification of the hydrological cycle under enhanced greenhouse warming, however (IPCC 1996). We incorporate this effect by assuming a uniform increase in precipitation of 10% per degree warming. Although regional changes could be larger, for the global scale this number can be considered as an upper limit.

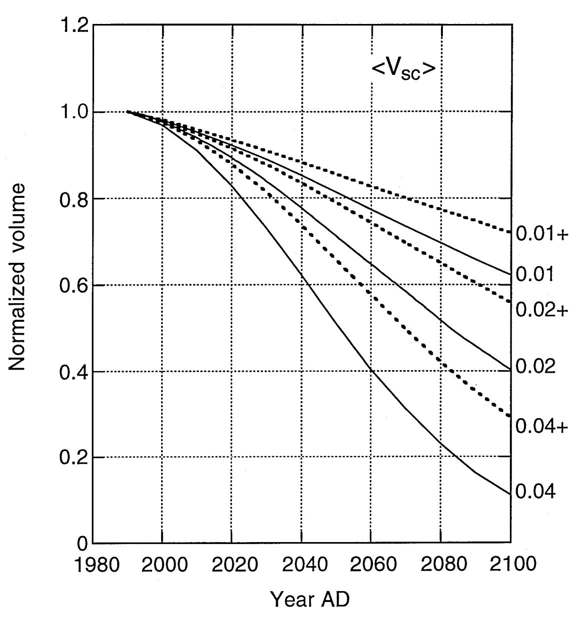
Some examples of the numerical experiments on this issue are shown in Fig. 7. The effect of a precipitation change of this magnitude is significant, but by no means enough to compensate for the enhanced melting due to the temperature rise. This would require much larger changes in precipitation (which is not impossible, as currently seen in western Norway where glaciers are advancing in response to strongly increased precipitation).

6 Other scenarios

In Fig. 8 results for all climate scenarios studied here are summarized. For a warming rate of 0.04 K/a, without a significant increase in precipitation, little ice would be left by the year 2100. On the other hand, if the warming rate would be limited to 0.01 K/a with a substantial increase in precipitation, we predict that the loss of ice would be restricted to 10 to 20% of the 1990 volume.

7 Epilogue

It is important to emphasize that the sample of glaciers studied here is not totally representative of all ice masses



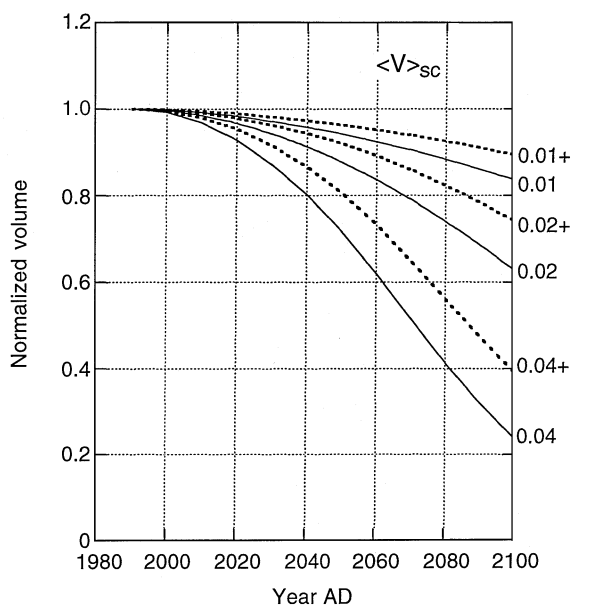


Fig. 8 Scaled ice volume for all climate change scenarios. *Labels* on the *curves* give the imposed warming rates (+ refers to including an increase in precipitation of 10% per degree warming). Again the $\langle V \rangle_{sc}$ curves are dominated by the ice caps

across the globe. In particular, the dry subpolar glaciers and ice caps, that have a low climate sensitivity but contribute a great deal to the total glaciated area, are not considered. Including these would probably lead to a smaller fractional loss of ice. Another product of this study is the mass-balance history derived for several glaciers from the calibration of the model against the observed record of glacier length. These mass balance reconstructions contain interesting climatic information. This point will be discussed elsewhere.

In this study we do not make a proposal for a simple mathematical formulation to represent (part of) the results. Our intention is to continue this work and, first of all, expand the resulting data set on numerical glacier experiments. We hope that this data set also proves useful for testing simpler models that can more easily be applied to large sets of glaciers (e.g. Haeberli and Hoelzle 1995). The work reported here is only possible because glacier data have been collected and compiled in a systematic way (Haeberli and Hoelzle 1993; and earlier volumes). The usefulness of models is determined by the possibilities to validate and calibrate. A continuation and extension of observations on glacier length and mass balance is thus crucial. It seems timely to develop a global strategy so that data sets become more representative and complementary.

Apart from measurements on individual glaciers, extrapolation of model results to all glaciers and ice caps can only be done when information on distribution of glacier size exists in the various glaciated regions in the world. Data sets currently available at the World Glacier Monitoring Service should therefore be extended. Combining high-resolution satellite imagery with digital terrain information and automated analyses guided by expert interpretation seems most promising.

Finally, all individuals running or planning to run dynamic glacier models are encouraged to carry out the climate change experiments described here. We hope this will ultimately lead to a more representative data set on the possible response of glaciers and ice caps to climate change.

Acknowledgements This study was initiated at a workshop sponsored by EISMINT (European Ice Sheet Modelling Initiative, a programme of the European Science Foundation). This workshop took place in Landgasthof Morteratsch, Pontresina, Switzerland, 17–18 October 1996. We are grateful to EISMINT for their generous support. During this meeting, Wilfried Haeberli gave useful comments to "the modellers" on the quality and representativeness of glacier data.

References

Greuell JW (1992) Hintereisferner, Austria: mass-balance reconstruction and numerical modelling of the historical length variations. J Glaciol 38: 233–244

Grove JM (1988) The Little Ice Age. Methuen, London, 498 pp

Haeberli W, Hoelzle M (1993) Fluctuations of glaciers 1985–1990 (vol VI). IAHS(ICSI)-UNEP-UNESCO, 319 pp

Haeberli W, Bösch H, Scherler K, Østrem G, Wallén CC (1989) World Glacier Inventory. Status 1988. IAHS(ICSI)-UEP-UNESCO

Hubbard AL (1996): High resolution glacier modelling. PhD Thesis. University of Edinburgh, UK.

Huybrechts P, Nooze de P, Decleir H (1989) Numerical modelling of Glacier d'Argentière and its historic front variation. In: Oerlemans J (ed) Glacier fluctuations and climatic change. Reidel, Dordrecht pp 373–389

IPCC (1996) Climate Change 1995, Chapter 6: climate models - projections of future climate. Cambridge University Press Cambridge, UK, 572 pp

Jóhannesson T (1997) The response of two Icelandic glaciers to climate warming computed with a degree-day glacier mass balance model coupled to a dynamic glacier model. J Glaciol 43(144): 321–327

Kuhn M (1985) Fluctuations of climate and mass balance: different responses of two adjacent glaciers. Z Gletscherk Glazialgeol 21: 409–416

Knap WH, Oerlemans J, Cadée M (1996) Climate sensitivity of the ice cap of King George Island, South Shetland Islands, Antarctica. Ann Glaciol 23: 154–159

Meier MF (1984) Contribution of small glaciers to global sea level. Science 226: 1418–1421

Nye J F (1965) A numerical method of inferring the budget history of a glacier from its advance and retreat. J Glaciol 5 (41): 589–607

Oerlemans J (1986) An attempt to simulate historic front variations of Nigardsbreen, Norway. Theor Appl Climatol 37: 126–135

Oerlemans J (1989) A projection of future sea level. Clim Change 15: 151–174

Oerlemans J (1994) Quantifying global warming from the retreat of glaciers. Science 264: 243–245

Oerlemans J (1997a) Climate sensitivity of Franz Josef Glacier, New Zealand, as revealed by numerical modelling. Arctic Alpine Res 29: 233–239

Oerlemans J (1997b) A flow-line model for Nigardsbreen: projection of future glacier length based on dynamic calibration with the historic record. Ann Glaciol 24: 382–389

Oerlemans J, Fortuin JPF (1992) Sensitivity of glaciers and small ice caps to Greenhouse warming. Science 258: 115–117

Oerlemans J, Van der Veen CJ (1984) Ice sheets and climate. Reidel, Dordrecht, 217 pp

Paterson WSB (1994) The physics of glaciers. Pergamon, Oxford, 3rd edn, 480 pp

Stroeven AP (1996) The robustness of one-dimensional, time-dependent ice-flow models: a case study from Storglacären, northern Sweden. Geogr Ann 78A (2–3): 133–146

Schmeits MJ, J Oerlemans (1997) Simulation of the historical variations in length of the Unterer Grindelwaldgletscher. J Glaciol 43: 152–164

Tufnell L (1984) Glacier hazards. Longman London, 97 pp

Warrick RA, Le Provost C, Meier MF, Oerlemans J, Woodworth PL (1996) Changes in sea level. In: Houghton JT, Meira Filho LG, Callander BA, Harris N, Kattenberg A, Maskell K (eds) Climate change 1995. Cambridge University Press, pp 363–405

Wigley TML, Raper SCB (1993) Future changes in global mean temperature and sea level. In: Warrick RA, Barrow EM, Wigley TML (eds) Climate and sea level change: observations, projections and implications. Cambridge University Press, pp 111–133

Zuo Z, Oerlemans J (1997) Numerical modelling of the historic front variation and the future behaviour of the Pasterze glacier. Ann Glaciol 24: 234–241