ON THE RESPONSE OF VALLEY GLACIERS TO CLIMATIC CHANGE

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ABSTRACT

In many cases the response of a glacier to changing climatic conditions is complicated due to the large number of feedback loops that play a role. Examples are: ice thickness - mass balance feedback, nonlinearities arising from complicated geometry, dependence of ablation on glacier geometry, coupling between debris cover, ice flow and ablation etc.

In this paper an attempt is made to quantify such processes by carrying out numerical experiments with an ice-flow model. Some conclusions and suggestions are:

(i) The longitudinal bed profile is very important. Apart from the well-known fact that glaciers are more sensitive when the bed slope is small, a reversed slope (slight overdeepening) creates branching of the equilibrium states, i.e., for the same climatic conditions two glaciers of different geometry can both be in a stable steady state.

(ii) Due to the height-mass balance feedback, glaciers on a smaller slope react slower to climatic change.

(iii) The mass balance gradient as observed on long valley glaciers is to a substantial part determined by systematic changes (along-valley) in glacier width and surface albedo. The balance gradient is thus coupled to the dynamics, and this should be studied further.

1. INTRODUCTION

In historic times, glacier variations have drawn the attention of many people inhabiting mountaneous regions. This applies in particular to those events that have brought damage to farmlands and buildings, either by direct advance of a glacier snout (e.g. Ostrem et al., 1977) or by blocking of rivers with subsequent flooding when the resulting lakes break through (e.g. Hoinkes, 1969). Such events have been documented by reports of all kind. For a discussion of glacier hazards, with many examples and additional references, see Tufnell (1984).

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Painters have also contributed significantly to our knowledge of glacier fluctuations. From the 18th and 19th centuries, a wealth of drawings, etches and paintings exist, making it possible, in combination with other investigations, to reconstruct front variations of large valley glaciers like for instance the Grindelwald Gletschers, the Rhône Gletscher, the Vernagtferner, and the Glacier d' Argentière. In more recent times, the last 100 years say, more systematic measurements have been carried out on a more global scale (Kasser, 1967, 1973; Müller, 1977; Patzelt, 1970; Reynaud, 1983). Some long records of front positions are shown in Fig. 1. In these curves the earlier parts are more uncertain, but extreme positions are probably fairly reliable.

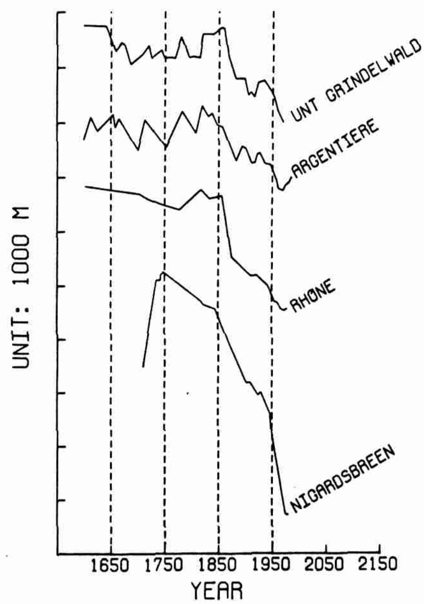

Figure 1. Variations in the ice-front positions of four valley glaciers. Data from Messerli et al. (1975), Vivian (1975), Aubert (1980), Ostrem et al. (1977).

The state of a glacier is of course not only determined by the position of the front, but also by its volume. Records of changes in volume are hardly available, however. Most systematic observations have been made in the last decades. The same applies to direct measurements of the mass balance. It is only since the use of hydroelectric power plants in glacierized areas that many glaciers are monitored on a regular basis. Series of mass balance measurements are at best about 30 years long, but most are shorter (Letréguilly, 1984). Nevertheless, it is the mass balance that relates glacier variations to climatic change. So a proper understanding of how the mass balance depends on climatic parameters like temperature, radiation and precipitation is crucial to the study of front variations.

It is generally assumed that glaciers have a relatively long time scale, typically 100 years. Although this is a realistic figure for the global mechanics of a large valley glacier, the front can react much quicker. A strong increase in melting rate on a glacier tongue, for instance, will lead to almost instantenous retreat of the snout. The record of the Brenva glacier (descending eastwards from the Mont Blanc) provides an interesting illustration. In 1926 a large rock slide covered part of the glacier and led to substantially lower melting rates. The ice front showed a quick reaction: in contrast to the fronts of other glaciers in the nearby regions, it advanced rapidly (Orombelli and Porter, 1982). Of course, the response time depends on the geometry; smaller glaciers react quicker.

The purpose of this paper is to draw attention to a number of feedback processes that make valley glaciers very sensitive to climatic change, and in particular to small changes in the global radiation balance. Changing glacier geometry will turn out to be an important factor when melting rates are considered. Such rates can be quite large, typically in the 5 to 10 m/yr range for large valley glaciers descending into regions where mean summer temperature is about 10 °C or so. Daytime surplus of the radiation balance of the ice surface and turbulent heat flux form the most important sources of energy for the melting process (e.g. Paterson, 1981). The relative importance depends on the particular geometric and climatological conditions. At lower elevation, where melting rates are higher, the contribution from the turbulent heat flux is comparable to that from radiation. In higher regions radiation becomes relatively more important.

To set the stage for some model experiments and the more detailed discussions on specific feedback loops in subsequent sections, we consider what might happen if, for some reason, the planetary radiation balance is perturbed. From models of the earth's climate we expect that a + 6 W/m^2 perturbation will lead to an increase of the mean surface temperature of about 2 K. However, the thermal intertia of the world oceans will damp the response; it may take a decade or longer before the new equilibrium temperature is actually approached. However, this 'oceanic damping' will be less significant when the energy balance of a specific site depends less on horizontal advection of heat. Surface temperature in the continental interiors should thus be expected to react more quickly to a change of the global radiation budget, and this probably also applies to some extent to deep valleys in mountanous regions.

If we now turn to valley glaciers, the first thing to note is that the increase in air temperature is the secondary effect (but not necessarily less important on the long run). The mass balance will react directly and <u>immediately</u> to the change in the radiation balance. Since, during the ablation season, the temperature of the glacier surface is at the melting point, the extra radiation will be used entirely for additional melting. As pointed out recently by the author (Oerlemans, 1986), there is another important aspect associated with the fact that the glacier temperature cannot increase. Any increase in the radiation balance will lead to a larger temperature difference between glacier and direct surroundings, causing a larger advective flux of sensible heat towards the glacier tongue ('oasis effect'). This mechanism is particularly effective when the glacier tongue is narrow: the increase in melting rate may be doubled.

So it appears that a glacier can make use of surplus radiative energy in a valley in a very efficient way. Important is the fact that the reaction of the mass balance is almost immediate. To this then comes the effect of the gradually increasing global air temperature, retarded by the thermal inertia of the climate system.

There are other factors, of a more general nature (i.e. not directly related to changes in the radiation balance), that make valley glaciers extremely sensitive to climatic change. For instance, the width of a glacier generally decreases when the ice thickness decreases. However, it is reasonable to assume that the contribution of the advective heat flux to melting is roughly inversely proportional to the glacier width. Thus, once retreat has been initiated and the ice thickness becomes smaller, additional melting will be caused by the decreasing glacier width.

In the following sections, an attempt is made to assess the potential importance of the feedback loops mentioned above. In many cases this cannot be done in a rigorous way, because observations on the mass and energy budgets of glaciers do not give a complete picture of the energy fluxes in a valley-glacier system. Order-of-magnitude estimates can be made, however. To link changes in the mass balance to front variations, a dynamic glacier model is needed. Such a model is briefly described in section 2, and it will be used as a tool throughout the rest of this paper. In section 3 the effect of geometry and topography of the glacier bed on the response to climatic change will be discussed. Section 4 deals with the feedbacks involving ablation at the glacier tongue. Then, in section 5, an attempt is made to tie the results together and to perform a simple 'carbon-dioxide experiment'. Here, perturbations of the global radiation balance and the mean air temperature are imposed as time-dependent functions to the glacier model.

2. A SIMPLE DYNAMIC GLACIER MODEL

To simulate the transient behaviour of glaciers for any valley profile and mass balance, a numerical model is required. We use a model that treats the vertically-integrated ice flow as a direct response to the driving stress. Such models have been used in glacier studies by various workers (e.g. Budd and Jenssen, 1975; Kruss, 1984; Oerlemans, 1986). The model used here is a flow-line model, in which the cross section is trapezoidal and may depend on distance x along the flow line, and on ice thickness H. The geometry is shown in Fig. 2. Here only a brief description is presented, for a general discussion on numerical modelling of ice flow the reader is referred to Oerlemans and Van der Veen (1984).

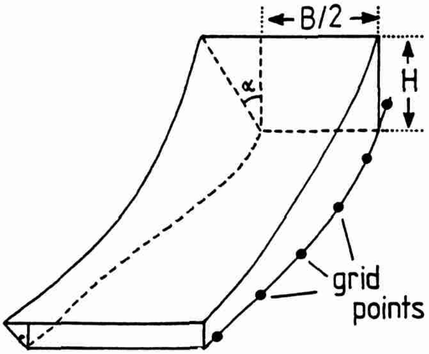

Figure 2. Geometry of the model glacier. The x-axis is aligned along the central flowline. At each grid point, width of the valley floor (B) and mean slope of the side walls has to be prescribed. The latter determines how glacier width is related to ice thickness.

The appropriate continuity equation reads:

$$\frac{\partial S}{\partial t} = - \frac{\partial (US)}{\partial x} + MB_s \quad (1)$$

S is the cross-sectional area, U the vertical mean ice velocity parallel to the bedrock, M mass balance, B_s glacier width at the surface [B_s = B+2Htg(α) = B+2μH]. Expressing S in H and B yields, after insertion in (1):

$$\frac{\partial H}{\partial t} = \frac{-1}{B+2\mu H} \left[(B+2\mu H) \frac{\partial (UH)}{\partial x} + UH \frac{\partial}{\partial x}(B+\mu H) \right] + M \quad (2)$$

This equation can be used to calculate the transient behaviour of a glacier for any geometry, provided that the ice velocity is locally related to the driving stress τ, defined as:

$$\tau = -\rho g H \frac{\partial h}{\partial x} \quad (3)$$

Here h is surface elevation. The total velocity U is made up of a sliding part U_s and a deformational part U_d, related to the driving stress as follows:

$$U_d = F_1 \, H \, \tau^3 \qquad U_s = \frac{F_2 \tau^3}{N} \quad (4)$$

N is the normal load, F_1 and F_2 are flow parameters. Since the effect of basal water pressure is not taken into account in this study, the normal load is simply set equal to the overburden ice weight. The values of the flow parameters actually used are: $F_1 = 0.95 \times 10^{-22}\ m^6\ s^{-1}\ N^{-3}$, $F_1 = 0.9 \times 10^{-14}\ m^5\ s^{-1}\ N^{-2}$.

Substituting the expressions for the velocity components in the continuity equation leads a nonlinear diffusion equation for H. A forward time-differencing scheme is used for the integration, together with a staggered grid to evaluate the ice flux divergence. For a sufficiently small time step, depending on ice thickness and surface slope, this scheme is absolutely stable without additional smoothing. All experiments discussed in this paper were carried out on a grid of 50 points along the flow line, spaced at 300 m. No attempt was made to treat the glacier snout in a sophisticated way (interpolation between grid points), because this does not effect in any significant way the response of the model glacier to climatic change. The glacier length thus appears as a discrete variable.

3. GEOMETRIC EFFECTS

In this section we investigate how valley width and longitudinal profile influence the response of a glacier to changing environmental conditions. It has of course been recognised for a long time that the glacier front position will be particularly sensitive to a change in the equilibrium-line altitude when:

(i) the accumulation area is large and the glacier tongue narrow;

(ii) the longitudinal slope of the bed is small.

For a recent discussion, see for instance Furbish and Andrews (1984).

Before discussing some experiments with the numerical model, we first identify from an extremely simple analysis how the bed slope effects the sensitivity of a glacier. We consider a glacier of constant width resting on a bed with constant slope γ (see Fig. 3). The mass balance is assumed to increase linearly with height relative to the equilibrium-line altitude E, i.e. $M=\alpha(h-E)$ Here α is a positive constant. If the length of the glacier is denoted by L, equilibrium requires that:

$$\int_L M dx = \alpha \int_L (H+b_o-\gamma x-E)dx = 0 \tag{5}$$

Integrating and solving for L yields:

$$L = \frac{2(H^*+b_o-E)}{\gamma} \tag{6}$$

Here H^* is the mean ice thickness. We now assume that the base stress is more or less constant, implying that $H(dh/dx) = \Lambda$, where Λ is about 11 m for a base stress of 1 bar. It follows that $\gamma H^* = \Lambda$, so the solution for the equilibrium length of the glacier becomes:

$$L = \frac{2(\Lambda/\gamma+b_o-E)}{\gamma} \tag{7}$$

From this expression a few things can be noted. First of all, the height-mass balance feedback, reflected in the term Λ/γ, becomes more important when the bed slope is smaller. For a base stress of 1 bar and a slope of 0.05, for instance, the mean ice thickness would be 220 m, which, for many glaciers, would not be negligible as compared to $b_0 - E$. Secondly, the sensitivity of glacier length to changes in the equilibrium-line altitude is inversely proportional to the bed slope (i.e. $\partial L/\partial E = -2/\gamma$). So this simple calculation indeed illustrates that glaciers resting on a bed with a small slope should preferable be considered when studying climatic change.

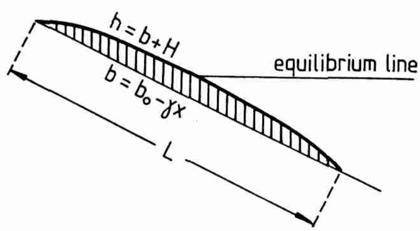

Figure 3. Geometry for the simple calculation of how glacier sensitivity depends on the slope of the bed. The x-axis is pointing in the direction of flow.

It is possible to do a similar analysis for a more complex geometry or a more detailed parameterization of the mass balance (upper limit for accumulation, for instance). However, this falls outside the scope of the present paper. Instead we now turn to some experiments carried out with the numerical glacier model.

In the following the mass balance is again parameterized in terms of elevation relative to the height of the equilibrium line, but now with an upper limit:

$$M = \min[1.5,\ 0.01(h-E)] \text{ m ice depth/yr} \qquad (8)$$

Here h and E are in m. With E between 2500 and 3500 m, this could represent midlatitude conditions in a climate that is not too dry.

The results of two integrations with different bed profile are shown in Fig. 4. The glacier width is a prescribed function of x, see plan view in the figure, and does not yet depend on ice thickness ! It is a valley glacier with a narrow tongue, and a wider accumulation basin. The integrations extend over 1000 yr of simulated time, and the equilibrium-line altitude is prescribed as follows:

$t < 300$ yr: E = 2700 m
$300 < t < 550$ yr: E = 3000 m
$550 < t < 800$ yr: E = 3300 m
$t > 800$ yr: E = 2700 m

The changes in snowline elevation are stepwise. In the first experiment the bed has a steep upper slope (up to 3500 m) with a flat plateau at its foot, and then a slope making a constant angle with the horizontal. The resulting equilibrium glacier profile (the one shown is for E=2700 m) is simple. Ice thickness in the lower part is small and decreases very smoothly towards the snout. In such a situation one expects a simple and quick respons to changes in the snowline elevation as is indeed shown by the plot of glacier length versus time. A 300 m increase of E leads to a retreat of about 1500 m within 25 yr. The 600 m lowering of the equilibrium line at the end of the simulation brings the glacier quickly back to its original shape.

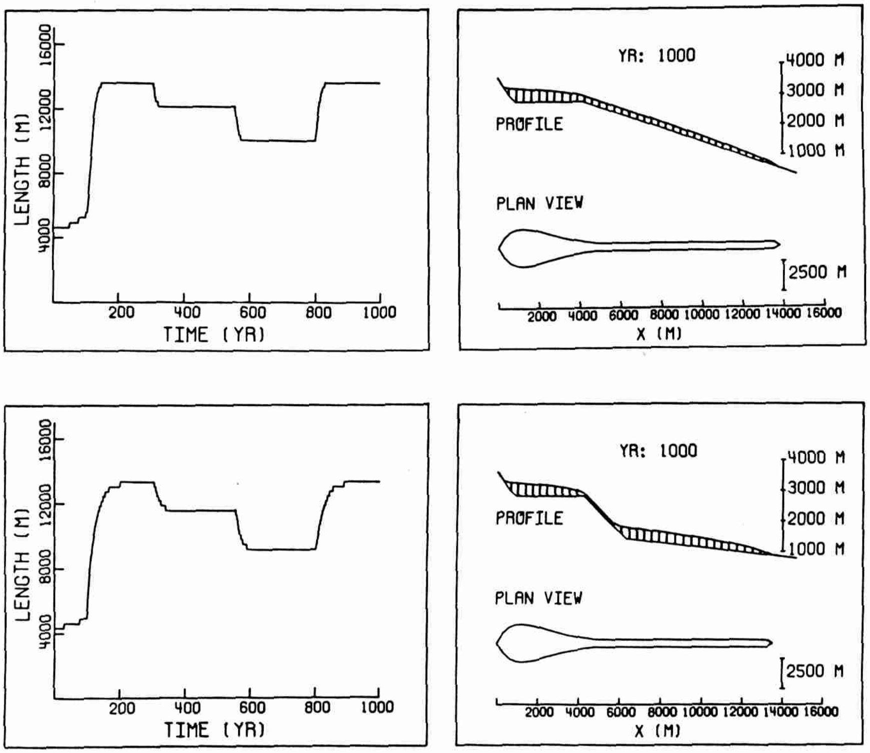

Figure 4. Glacier length as a function of time (left) and steady-state geometry (right) for a sensitivity experiment with different bed topography. A stepwise change in snowline elevation is imposed three times (see text). Note differences in response time due to differences in bed profile.

In the second experiment (lower part of the figure), the bed profile is somewhat more complicated. After 5 km there is an 'ice fall', changing abruptly into a rather flat valley bottom. This allows the glacier to grow thicker, leading to a situation in which the sensitivity to changes in E has increased (see analysis given above), and in which the response time is notably longer. Firstly, it takes about 50 years more to reach a steady state, but the reaction to the rising equilibrium line is also slower (compared to the first experiment, it roughly doubles). These effects are also a consequence of the height-mass balance feedback, in which the growing or shrinking glacier effects its own mean surface elevation such that it causes a significant change in the mass balance.

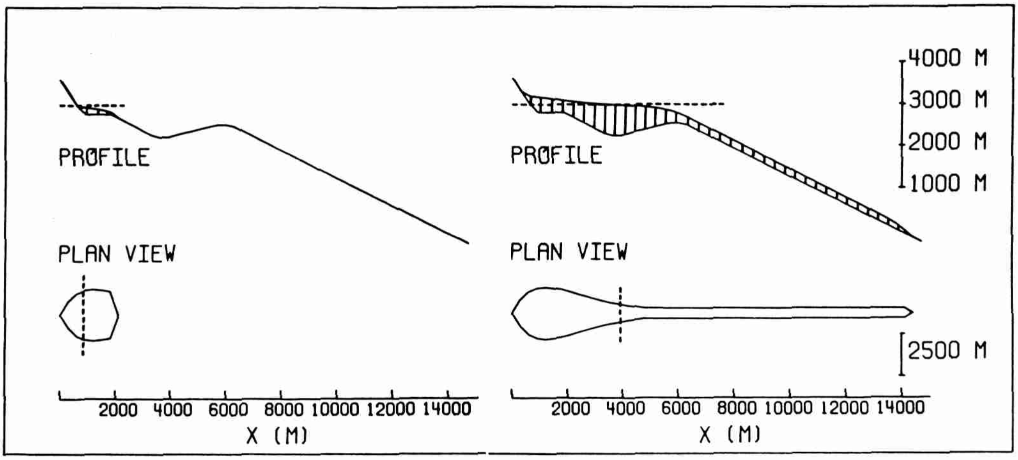

Figure 5. An illustration of the nonlinearity introduced by the height-mass balance feedback. These glaciers are both in equilibrium, under the same climatic conditions (E=3000 m). The equilibrium line is indicated by the dashed line.

In case of a bed where the slope changes sign, the height-mass balance feedback introduces essential nonlinear behaviour. An example is shown in Fig. 5. Both states shown represent stable equilibria for an equilibrium-line altitude of 3000 m. So although the glacier geometries are very different, they both have a zero net balance. The small glacier was obtained with a no-ice initial state , keeping E at 3000 m all the time. The large glacier, however, can only be simulated by starting with a lower equilibrium line and raising it after 100 yr or so to 3000 m. This experiment clearly shows that one should be careful with the interpretation of former glacier-front positions in terrain with complex topography (this point has recently been stressed also by Burbank and Fort, 1985). In such cases, using a relatively simple glacier model as the one presented here will certainly be helpful.

4. ABLATION

4.1 Introduction.

In recent years, ablation measurements have been carried out on many glaciers and a number of methods have been developed to parameterize ablation in terms of meteorological elements. For a general discussion see for instance Kuhn (1980).

The simplest method to estimate ablation from weather data is to 'integrate positive temperatures', i.e.:

$$A = \Sigma \int \max(0,T)dt \qquad (9)$$

Here A is total ablation, T air temperature of a nearby station and Σ the 'degree-day factor'. Methods based on this approach have been used for a long time (for a recent discussion, see for instance Braithwaite, 1984). For a particular location the method is succesful in explaining variations in ablation rate, but it is not very universal: the degree-day factor varies widely. There have also been attempts to derive statistical relations between ablation (and net balance) and temperature, radiation and precipitation (e.g. Martin, 1977; Letreguilly, 1984; Pollard, 1981). Although some general results have shown up [for instance: summer temperature and early summer precipitation (snowfall) are the most important factors concerning the net balance of glaciers in the Alps], the constants of proportionality again vary widely from place to place.

A more accurate calculation of ablation requires a consideration of the energy budget at the glacier surface. This has been done for instance by Ambach (1965), De la Casinière (1974) and Hogg et al. (1982). The basic assumption is that, in the ablation season, the ice is at the melting point, and that any positive energy balance will thus lead to melting immediately. A more detailed calculation carried out recently by Greuell and Oerlemans (1986), in which vertical heat fluxes in- and outside ice and/or snowpack are taken into account explicitly, has shown that this assumption is correct. Substantial differences only occur in regions where the ablation season is short.

So there is a complete hierarchy of models relating ablation to meteorological conditions. However, it is not so obvious that relations thus found are meaningful instruments in studying how glaciers react to climatic change. One should be aware of the fact that other factors, in particular those associated with glacier geometry, can become important once substantial changes occur in the shape of a glacier. One of these involves the horizontal exchange of energy between the atmospheric boundary layers above the glacier and above the ice-free surrounding grounds.

4.2 Glacier width and ablation rate.

Although the glacier-wind circulation and the thermal contrast between air just above a glacier and just above surrounding rock are well-known phenomena, very little attention has been paid to the implications for the energy budget of the glacier surface. To the knowledge of the author, systematic investigations of the contribution of the advective heat flux to melt energy at glacier tongues have not been carried out [A pilot study undertaken by Wendler (1974) suggests that it could be very important]. In estimating the contribution from the turbulent heat flux to the glacier surface, air temperature is normally prescribed (taken from observations). However, when the glacier geometry changes substantially, the advection of heat from the surroundings will change, and so will air temperature over the glacier (apart from any global climatic perturbation).

In a recent paper (Oerlemans, 1986), the author presented a simple model to calculate ablation rates, in which the advective heat flux was taken into account. Two important results emerged, namely, (1) the ablation rate depends in a significant way on the width of the glacier, and (2) the increase in ablation due to a change in the net radiation balance is further enhanced by the advective heat flux from ice-free grounds towards glacier. In this subsection we elaborate further on (1).

The importance of advective heat fluxes is directly reflected by the fact that on many glaciers ablation increases when going from the centre to the edge. It should thus be possible to use a transverse ablation gradient to examine the role of the advective heat flux, and to find a relation between glacier width and mean ablation. This could further support the theoretical result referred to above, which was obtained with a model in wich parameter values have to be chosen in a rather ambiguous way.

Suppose that the transverse ablation profile can be written as:

$$A = A_1 + A_2\, e^{-(W+y)/L} \tag{10}$$

The y-axis is perpendicular to the central flowline, and the region of interest is from y=-W (edge of glacier) to y=0 (middle of glacier). So the ablation rate at the edge equals $A_1 + A_2$. The length scale determines to what extent the ablation on the glacier is affected by the surrounding ice-free grounds. Integrating over the half-width of the glacier yields for the mean ablation A^* :

$$A^* = A_1 + A_2 \frac{L}{W} \lfloor 1-e^{-W/L} \rfloor \quad , \tag{11}$$

showing that the ablation goes to A_1 when the glacier width goes to infinity.

Although it is evident that the intensity of the thermal convection on the valley determines to a large extent the value of L, it is not so clear what this value actually should be. Data on cross-glacier ablation profiles are too scarce to derive a value for L; something of the order of 1000 m seems to be reasonable, however.

One of the few valley glaciers on which the ablation pattern has been measured in some detail is the Hintereisferner (Oetztaler Alpen, Austria). On its tongue, ablation increases substantially towards its sides. To arrive at an order-of-magnitude estimate, we apply ablation measurements from stakes at a section of the glacier where the surface elevation is about 2670 m for the period 1971-1973 (three ablation seasons). This elevation was choosen because here the stakes had favourable positions for the present purpose for a number of years (a few very close to the glacier margin, a few in the middle). Data were taken from Kuhn et al. (1979). In the period referred to, the mean annual ablation was 2.65 m water eq. in the middle, and 3.25 m water eq. at the sides. The difference is of the same order as that reported by Wendler (1974). At this location on Hintereisferner, the half width is about 340 m. Matching these data with the cross-glacier ablation profile proposed above yields (with L = 1000 m):

$$A^* = 1.18 + 2.07\,\frac{L}{W}\left\lfloor 1-e^{-W/L}\right\rfloor \text{ m water eq.} \qquad (12)$$

From this expression we find for example ablation values (m water eq.) of 3.21, 2.78 and 2.07 for half-width's of 50, 500 and 2000 m, respectively.

Whether the relation between ablation on glacier width effects the climatic sensitivity in a significant way depends on the geometry of the valley, of course. Here we consider just one numerical experiment in a situation where $\mu = \frac{1}{2}\,\partial B_s/\partial H$ is large. A simple linear bed profile is used (see Fig. 6) and μ is set to 16. The basic width B of the valley is 100 m. The mass balance is now written as the sum of ablation and accumulation according to:

$$M = Acc + Abl$$

$$Abl = \min\left\lfloor 0,\ A_1 + 2A_1\,\frac{L}{W}\,(1-e^{-W/L})\right\rfloor \qquad (13)$$

$$A_1 = \min\left\lfloor 0,\ (h-h_o)a\right\rfloor$$

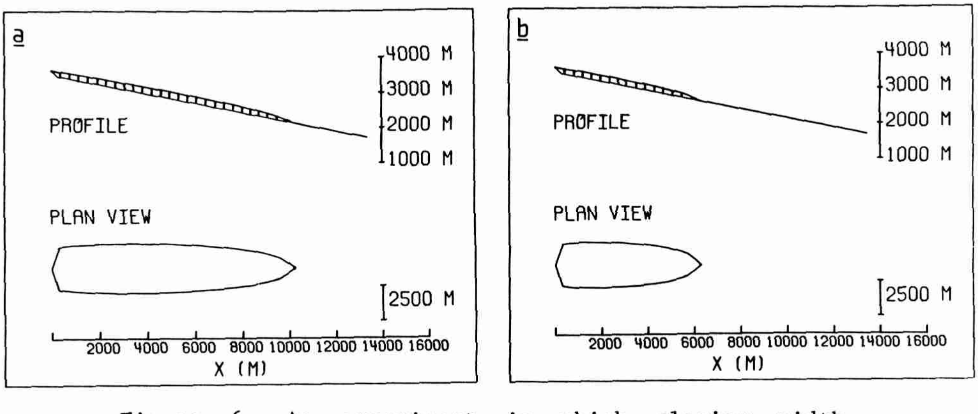

Figure 6. An experiment in which glacier width depends on ice thickness. In a a steady state is shown, in b the new steady state after a sudden 300 m increase of the 'melt line' (see text).

The accumulation is assumed to be constant (here, Acc = 1.5 m/yr), and the 'reference ablation' ($B \rightarrow \infty$) decreases linearly with surface elevation (here a = 0.007 m ice depth/m; h_0 = 3200 m; this implies an equilibrium-line altitude of 2915 m)

The glacier model was integrated in time until a steady state was reached. This state is shown in Fig. 6a. Then the value of h_0 was increased instantaneously by 300 m, and this of course leads to substantial retreat of the glacier (exp. 1). It takes about 100 yr before a new equilibrium has been established, but 75 % of the retreat occurs within 50 yr. The final state after the 'climatic warming' is shown in Fig. 6b.

For comparison, the experiment was repeated with a mass balance formulation independent of glacier width (exp. 2). This was done by replacing the ablation in equation (13) by: Abl=min [0, A_1], and through adjustment of A_1 in such a way that the initial equilibrium state is the same as in the former run. Retreat for some selected times after the stepwise change in h_0 are given in Table I. As expected, it appears that in the case of ablation depending on glacier width the sensitivity is larger. However, the difference decreases with time progressing, and the equilibrium states are rather similar. Anyway, this example shows that the ablation - glacier width coupling may be important, but it is difficult to make a general statement since so much depends on the actual geometry of the bed.

Table I. Retreat of the model glacier after a sudden climatic warming. The difference between exp. (1) and exp. (2) is described in the text.

time (yr)	exp. 1 (m)	exp. 2 (m)	difference (m)
10	290	180	90 (31%)
20	1110	780	330 (30%)
40	2690	2270	420 (16%)
80	3770	3480	390 (10%)
150	4210	3960	250 (6%)

4.3 Variations in albedo.

Systematic variations in surface albedo may well contribute to the longitudinal ablation gradient. The amount of exposed morainic material generally increases towards the snout, causing large differences in effective (area-averaged) short-wave albedo. For a large valley glacier a typical value for the albedo (during the ablation season), in the vicinity of the equilibrium line, may be 0.5. At the glacier tongue, however, this figure may go down to 0.25 (see Fig. 7 to illustrate the point). The resulting difference in ablation can easily be a few m ice depth. It depends on the surface slope of the glacier how large the actual contribution to the balance gradient is. For a 500 m altitude difference between equilibrium line and glacier front, the 'albedo effect' thus typically contributes 0.4 m ice depth per 100 m, i.e. about 40 % of a characteristic balance gradient!

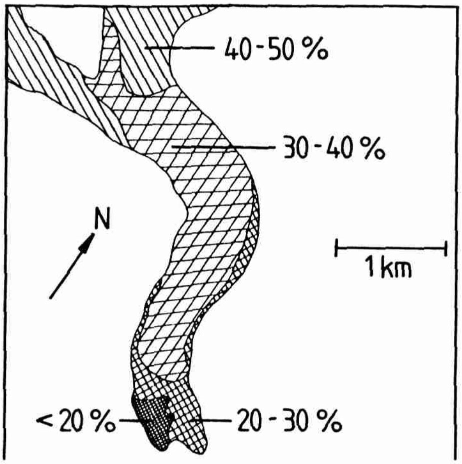

Figure 7. Variations in albedo over the lower ablation zone of a valley glacier (Nigardsbreen, 15 August 1972). From Tvede (1974).

The complicating factor now is that the amount of exposed morainic material (and atmospheric dust) interferes with the transient behaviour of the glacier. On a retreating glacier with an almost stagnant tongue (low ice velocities), the amount of debris on the surface increases in time, and so probably does the ablation rate, until the debris layer become so thick that the isolating effect starts to dominate. For an advancing glacier, on the other hand, morainic material is deposited more efficiently at the glacier snout (and at the sides) and ablation may decrease. It is outside the scope of this paper to discuss this mechamism in detail, but the order-of-magnitude estimate given above points to a certain importance.

4.4 Summary.

In virtually all studies in the literature, the mass balance of a glacier is represented in terms of surface elevation, and the balance gradient has generally been accepted as a parameter characterizing climatic conditions. For many purposes this may be acceptable, but when studying the effects of climatic change one has to be careful.

The objection raised above is that part of the (longitudinal) gradients as observed in the field stem from:

(i) differences in glacier width along the flow line (through the effect of advective heat transfer);

(ii) systematic increase of surface albedo with altitude on the glacier tongue (varying amount of exposed morainic material).

Depending on the particular geometry, these factors will increase the sensitivity of glacier length to climatic change slightly or strongly. The uncertainties involved call for further study of the energy budget of the entire valley-glacier system.

5. GREENHOUSE WARMING AND GLACIER RETREAT

As discussed in the Introduction, ablation on a glacier surface reacts to a change in the radiation balance as well as to a change in air temperature. There are a number of effects to be taken into account when the concentration of carbon dioxide (and other radiatively active gases) changes. For the present purpose, it seems reasonable to distinguish between:

(1) A change in the radiation balance at the earth's surface, all other things (distribution of moisture, atmospheric temperature profile, etc.) being equal. This effect is immediate.

(2) The feedback on the radiation balance. The increasing humidity associated with rising tropospheric temperatures is probably the most important effect (Ramanathan, 1981; Hansen et al., 1981) and leads to substantial magnification of the initial perturbation of the radiation balance. Over a land surface, the major part of the increasing downward longwave flux is cancelled by an increasing upward flux due to higher surface temperature. Over a melting ice surface this is not the case. However, since humidity is related to ocean surface temperature, a lag will occur.

(3) Changing air temperature associated with the perturbed radiation balance of the entire atmosphere, also with a lag.

For a 1 K increase in surface air temperature, the increase of the net longwave balance at the surface is estimated to be about 1 W/m^2 (Luther & Cess, 1985; Table B.1), whereas the downward component increases by about 6 W/m^2. The difference is large and suggests that melting ice bodies could be good indicators of changes in atmospheric longwave emissivity! Indications on former levels of carbon dioxide have recently been reviewed by Gammon et al. (1985), and the conclusion has been reached that atmospheric amounts of both methane and carbon dioxide were already increasing in the first half of the nineteenth century. The best evidence for this comes from carbon isotopes in tree rings (e.g. Stuiver et al., 1984), and substantial changes in land use are assumed to be the cause. However, the burning of fossil fuels soon took over.

It is not the intention to review here scenarios for the greenhouse warming as presented in the literature. The one shown in Fig. 8 is a kind of average picture accepted by many climatologists. It is fairly simple to apply this to a model of a schematic glacier, to arrive at an order-of-magnitude estimate of glacier retreat. The changes in air temperature and radiation balance are first translated into changes of h_o (the 'melting altitude'). Assuming accumulation to be constant, this is equivalent to shifting the equilibrium line. Kuhn (1980) analyzed glacier mass balance in slightly different climatic conditions in basically the same geographical region. His work suggests:

$$h'_o = 6.5\,Q' + 125\,T' \qquad (15)$$

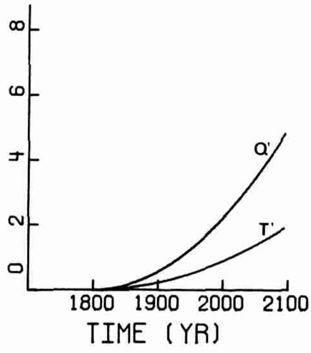

Figure 8. Forcing imposed on the glacier model, representing in a schematic way the effect of increasing atmospheric carbon dioxide and other greenhouse gases. Q' is the perturbation of the radiation balance (W m^{-2}), T' of air temperature (K).

The analysis did not contain the feedbacks discussed in the previous section, however.

Figure 9 shows the result of an integration for a 'typical' large valley glacier. The model was first run to a steady state and then a growing perturbation according to Fig. 8 and eq. (15) was imposed. As expected, the initial retreat is slow and small. Still, 'Greenhouse-warming retreat' predicted for the year 2000 is significant. It seems that for large valley glaciers a 1000 m retreat is typically what one expects to be the result of increasing carbon dioxide up til now. Or, to put it another way, a significant fraction of the observed world-wide retreat of valley glaciers could be due to the carbon dioxide warming.

6. FINAL REMARKS

Some authors, notably Meier (1984), have suggested that part of the observed sea-level rise over the last century can be attributed to melting of mountain glaciers. The sensitivity tests carried out here add that this melting probably is the result of the increasing concentration of carbon dioxide (and other radiatively active trace gases) in the atmosphere. So there seems to be increased evidence that the rise of world mean sea level is related to the carbon dioxide warming. This is an important point, because in many projections of future sea level the rate at which the sea rises presently (that is, over the last 100 years) is taken as basic trend, to which carbon dioxide effects are added.

It is quite obvious that many uncertainties still exist concerning the response of glaciers to a slightly changing climate. I hope to have identified some of them. Although many energy-balance and mass-balance studies have been carried out, it seems worthwhile to direct some effort to two specific points, namely:

(i) The role of the advective heat flux, related to a changing glacier geometry. This certainly involves the study of local wind systems.

(ii) The dependence of effective surface albedo on the dynamic history of the glacier ('debris dynamics').

Since snout variations are so well documented for a number of glaciers, and since there seem to be many reasons why glaciers should be very sensitive to a changing radiation balance, it seems worthwhile to employ them as basic climate indicators. However, a better understanding of the feedback loops discussed in this paper should first be achieved.

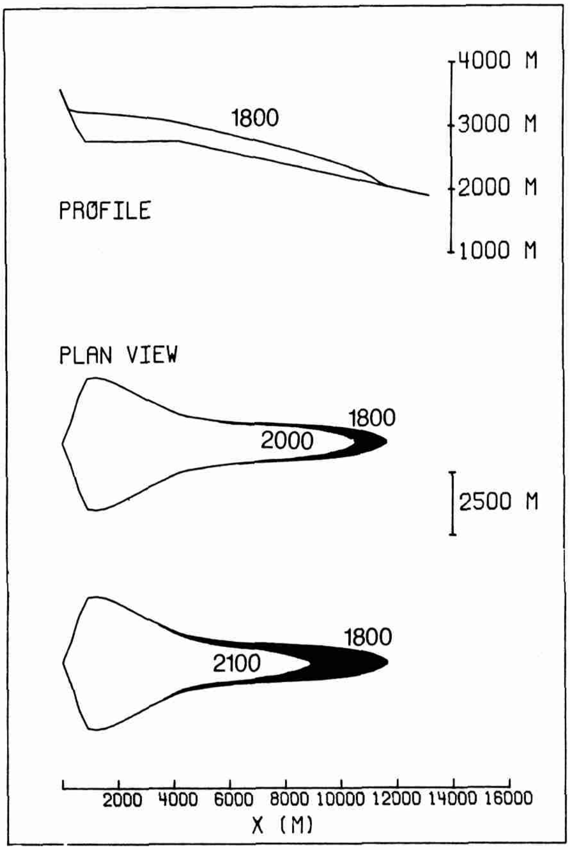

Figure 9. Greenhouse warming experiment for a schematic glacier. The black area shows the difference between the 1800 equilibrium profile and years as indicated.

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