NUMERICAL MODELLING OF GLACIER D'ARGENTIERE AND ITS HISTORIC FRONT VARIATIONS

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ABSTRACT

A numerical glacier model has been developed for Glacier d'Argentière (France) in order to study its relation with climate and investigate possible causes for the observed variations in the terminus record since the beginning of the Little Ice Age.

At first results are presented from a basic sensitivity investigation, with plots of steady state glacier length versus perturbations in mass balance and glacier reaction with respect to sinusoidal net balance oscillations. An attempt is then made to simulate the historic front variations. The mass balance history of the glacier is constructed assuming a linear relationship with (i) summer temperature anomalies and (ii) mean annual temperature anomalies for Basel dating back to the beginning of the 16th century.

Although model run (ii) turns out to yield better agreement with the observations, both simulations have in common that the observed glacier retreat comes too late. Improved simulations can only be obtained assuming an additional negative mass balance perturbation of around 0.1 m/year over roughly the last 150 years.

These results indicate that the assumption of a linear relationship between summer temperature and the glacier's mass balance may not be valid anymore when extrapolated to past environments. This might be evidence of additional micrometeorological and glacier surface conditions prevailing in valleys at maximum glacier extent, that are not absorbed well in the climatic records.

1. INTRODUCTION

Glacier d'Argentière, a valley glacier in the French Mont Blanc area (45.9°N, 7.0°E), is one of the few Alpine glaciers that have a fairly reliable historic record of front variations dating back to about 1600, i.e. since the beginning of the Little Ice Age (Vivian, 1975). This

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record is characterized by several rapid advances, most noticeably between 1625-1640, 1700-1720 and 1765-1780, culminating in the Neoglacial maximum in 1825. Since 1860 the glacier has been generally receding, alternatively slow and discontinuously (1883-1945: 1.1 m/year) and fast and uniformly (1866-1883: 37.1 m/year; 1945-1967: 21.4 m/year) over about 1600 m, reaching a glacial minimum in 1968. Since that time and up to the mid-eighties, Glacier d'Argentière is reported to have recovered some 300 m (Reynaud, 1986).

Glacier d'Argentière is also of special interest because it is quite well documented and surveyed, both in terms of its mass balance and ice depth, with velocity measured at about 5 locations along its length, in particular since 1975 by the Laboratoire de Glaciologie et de Geophysique de l'Environnement, Saint Martin d'Hères, France (Hantz, 1981; Reynaud, 1986). This relative wealth of data and the glacier's fairly simple geometry - it is a 'classical' Alpine non-surging valley glacier with a wide accumulation basin and a long narrow tongue - make Glacier d'Argentière a prime candidate for a modelling study.

Glaciers respond to climatic change in a complex way. Fluctuations in climatic parameters cause changes in snow accumulation and ice melt on the glacier surface. Over a sufficiently long period of time then, the terminus reponse will reflect the integrated changes in climatic input via the overall mass balance. Each glacier behaves as a particular case, with a dynamic response determined in essence by its own geometry. Important here are glacier length, bed topography and area-elevation distribution. Since the historic record goes back further than the instrumental record, an investigation of the relationship between terminus position and climatic input should aid in the reconstruction of past climate.

This suggests the use of a dynamical glacier model relating changes in ice thickness and hence, terminus position, to changes in the mass-balance. There have been some model studies in this direction, based on the numerical models developed by Budd and Jenssen (1975). In an early paper an attempt was made to estimate climatic change from the retreat of small mountain glaciers in Irian Jaya, Indonesia by Allison and Kruss (1977). Smith and Budd (1981) made an estimate of the difference in conditions between the present and the peak of the Little Ice Age by comparing known terminus histories and the reaction of the modelled glaciers (Storglaciärens, Hintereisferner, Vernagtferner, Aletschgletscher) to a simple sinusoidal function. Similar studies have been conducted by Kruss and Smith (1982) and Kruss (1983, 1984) on Vernagtferner and Lewis Glacier (Kenya).

In the present paper, as in Oerlemans (1987) in a study on Nigardsbreen (Norway), climatic series in part based on proxy data will be used to drive the glacier's mass balance history. Studying Alpine front position series in this way may then lead to a better understanding of the causes of glacier variations in Europe. At first results are discussed from a basic sensitivity investigation, in which glacier length versus perturbations in the mass balance are studied and in which the model is submitted to sinusoidal net balance forcing of various periods and amplitudes. An attempt is then made to simulate the historic front variations of Glacier d'Argentière. We performed experiments in which climatic data for Basel dating back to the first half of the 16th century are used, as compiled by Pfister (1984) on the basis of weather descriptions, environmental data and available instrumental records.

It will turn out, however, that the attempt is not very successful. In spite of the fact that a model of the present type is expected to reproduce the low frequency response only, in particular the glacier retreat since about 1860 is not modelled well by any of the performed runs. In a final section, possible causes of the discrepancy between modelled and observed front variations are discussed.

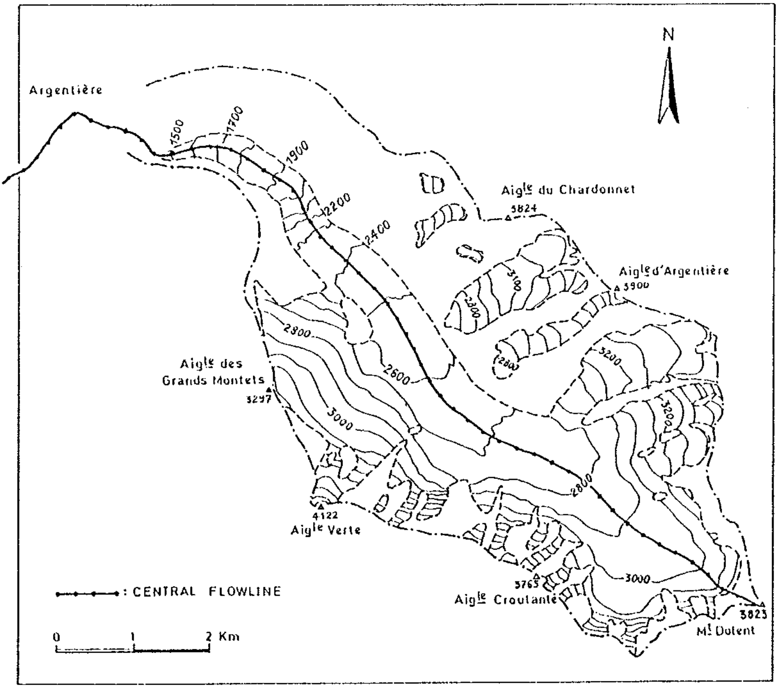

Figure 1. Map of Glacier d'Argentière showing present (1980) extent and the central flowline with gridpoint spacing of 250 m.

2. THE NUMERICAL MODEL

The dynamical glacier model employed in this study describes ice flow along a central flowline. Regarding the fairly simple geometry, this flowline is easily constructed and follows approximately the centre of the glacier, see Fig. 1. Every point x along this line, spaced 250 m apart, is assumed to represent mean lateral conditions. The most important parameter concerning three-dimensional geometry, namely the varying width distribution, is then accounted for in the continuity equation. Since ice density is assumed to be constant, we have:

$$\frac{\partial H}{\partial t} = -\frac{1}{b}\frac{\partial (HUb)}{\partial x} + M = -\frac{\partial (HU)}{\partial x} - \frac{HU}{b}\frac{\partial b}{\partial x} + M \qquad (1)$$

where H is local ice thickness, U is the mean velocity parallel to the bedrock, b the glacier width and M the annual mass balance, expressed in m ice depth per year.

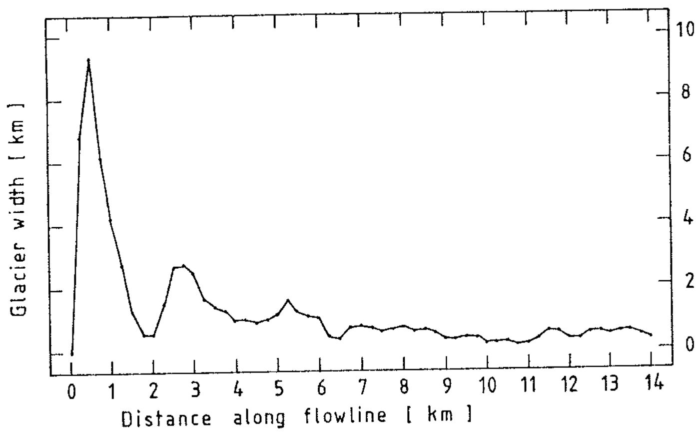

Figure 2. Present glacier width distribution (b_{ref}) as used in the ice flow model.

The gridded width values b_{ref}, displayed in Fig. 2, have been constructed respecting the present area-elevation distribution. This approach may at first glance present some peculiarities. For instance, it explains the remarkably large width gradient and apparent flow convergence found between x = 1000 m and x = 2000 m, since the surface slope along the centre line (between the 3000 and 2900 m contour lines) is here not representative for the entire flowband. Obviously, conditions at grid points high up in the accumulation area are less well defined.

The width distribution should also depend on ice thickness. Especially in the lower tongue area, this may become important, since a thinning glacier generally implies a decreasing width and consequently, a smaller ablation area. We assumed a simple parabolic cross section leading to:

$$b = b_{ref} \left[\frac{H}{H_{ref}}\right]^{\frac{1}{2}} \qquad (2)$$

where H_{ref} is taken from the present (1980) modelled glacier and b_{ref} read from a map (IGN topographical map on scale 1/25,000, 1980). Values of b_{ref} downstream the 1980 front position were generated with a constant ice thickness H_{ref} of 100 m.

Other input data, namely bed topography is known at 5 locations along the glacier length (Hantz, 1981). Since surface topography is read from a topographical map, remaining bed values, or, alternatively ice thickness, can in first approximation be estimated from plastic ice flow theory, relating ice thickness to the local surface gradient, with the factor of inverse proportionality given by a specific choice of the yield stress τ_0 (e.g. Paterson, 1981). Calculated driving stresses (with a surface gradient taken over 1500 m in order to smooth local surface irregularities) at these 5 locations were then inter/extrapolated over the whole glacier to obtain the unknown ice thicknesses. Except for the glacier part below the ice fall (Sêracs de Lognan at x = 8250 m), these driving stresses (as defined in eq. (4)) appear to be in the range 150-200 kPa. As calculated 'shape factors' (assuming a parabolic cross section) for these same 5 locations all are between 0.5 and 0.6 (the difference with 1 gives the part of the driving stress opposed by side drag), this corresponds with basal stresses of the order of 80-120 kPa, i.e. within the expected value range (Paterson, 1981). In what follows, however, distinction between side and basal drag will not be made, in effect putting forward a constant 'shape factor', whose value is then most easily incorporated in the flow parameter.

To arrive at an equation for the mean horizontal velocity, the flow law is written as (e.g. Oerlemans and Van der Veen, 1984):

$$U = A\,\tau_b^3\,H \qquad (3)$$

with A the flow parameter and where the driving stress is given by the basal shear stress τ_b:

$$\tau_b = -\rho g H \frac{\partial(H+h)}{\partial x} \qquad (4)$$

Here ρ is ice density (taken as 870 kg m^{-3}), g acceleration of gravity and h bedrock elevation. Hence, the velocity is a locally defined quantity, depending on ice thickness and surface slope. Strictly speaking, equation (3) describes the velocity contribution resulting from internal deformation only. However, since there is with respect to (vertical) mean velocity not so much difference between deformation (concentrated near the base) and basal sliding, any basal sliding may be assumed to be reflected in the value of the flow parameter. As in the Smith and Budd (1981) model, preliminary experiments have been conducted in which the geometry of the glacier cross-section was taken into account with shape and velocity factors (relating the centreline velocity to the cross sectional 'mean'). It was found, however, that contrasts in these geometric parameters within realistic bounds were by no means crucial to the general behaviour of the model. For simplicity then, these effects are also assumed to be represented by the value of the flow parameter, that now essentially serves a tuning purpose. Good results are obtained while setting A = 0.8 10^{-16} $year^{-1}$ Pa^{-3}.

Substitution of (3) and (4) in the continuity equation (1) then leads to an expression that can be regarded as a diffusion equation for ice thickness H. The resulting equation is most easily solved with a straightforward explicit finite difference scheme. To ensure computational stability a staggered grid in space is used, in effect calculating volume fluxes in between grid points with a mean diffusivity. Details of this scheme can be found in Oerlemans and Van der Veen (1984). With a spatial resolution of 250 m, this scheme allows time steps up to 0.1 years.

Finally, a simple procedure is adopted to follow the glacier terminus position in between grid points, because the maximum length contrast in the period considered amounts to roughly 1600 m, i.e. about 6-7 grid points only. This problem has been adress by Kruss (1984) in a more sophisticated way by calculating the volume of ice past the last grid point in use, and then computing the length necessary to contain this volume within a specified longitudinal snout shape. The procedure employed here is given in by the observation that the model glacier will jump to the next grid point whenever the absolute value of the ratio of the flux divergence and mass balance in that point equals unity. An approximately 'linear' transition during glacier evolution from one grid point to another is then obtained by raising this ratio to some power m:

$$L = (N_g - 1) \cdot \Delta x + \left[\frac{\left[1/b\ \partial(HUb)/\partial x\right]_{Ng+1}}{M_{Ng+1}} \right]^m \cdot \Delta x \qquad (5)$$

where L is total glacier length, N_g the index number of the last grid point in use and Δx the grid point distance. We used m = 0.2.

3. BASIC SENSITIVITY EXPERIMENTS

The basic mass balance parameterization versus surface elevation, to be perturbed in the experiments, is taken as a linear fit to the observed mean mass balance for the period 1976-1983 (Reynaud et al., 1986), in m ice depth/year:

$$M = 0.0071\ (H + h - 2910) \qquad \text{if } (H + h) > 1750 \text{ m} \qquad (6a)$$

$$M = 0.02\ (H + h - 2910) + 15 \qquad \text{if } (H + h) \leq 1750 \text{ m} \qquad (6b)$$

$$M = 3.60 \qquad \text{if } M > 3.60 \qquad (6c)$$

During this period the mean equilibrium line altitude (ELA) was about 2910 m a.s.l. Here it is assumed that the mean balance gradient of 0.71 m ice depth/100 m, as observed in the 1800-2800 m altitude interval, also applies higher up in the accumulation zone. The maximum value of 3.60 m/year is the observed 1954-1971 mean for Vallée Blanche (Mont Blanc area) at an altitude of aroung 3600 m. In the lower tongue area below around 1750 m, Hantz (1981) mentions a single year balance measurement in 1957, with a significant larger balance gradient (about 2 m/100 m). We feel this is a real feature, possibly related to radiative side wall effects as the glacier enters a more narrow valley, and, hence, should no be left out.

One can essentially distinguish two ways to express variations in the net balance curve following changes in the climatic state. Either one assumes a balance shift or an elevation shift, respectively shifting the net balance curve parallel to the balance or elevation axis. In the first approach the balance gradient at a specified location is considered to be constant, so that the imbalance is independent of altitude. It appears that this is the case for Alpine glaciers (Kuhn, 1984), who postulates that most likely increased ablation gradients in the long ablation period of negative years are balanced by increased gradients of solid precipitation and of albedo due to summer snow falls in positive years. Also, with respect to the limited mass balances measurements available, the linear balance model of Lliboutry (1974) seems to work well for Glacier d'Argentière (Hantz, 1981; Reynaud et al., 1986).

Since it is not known how far the present glacier is out of equilibrium, the flow law parameter A has been chosen such that the basic mass balance distribution produces a steady state length close to the 1980 position, that will be considered as a reference state. This glacier with total length = 10.49 km is shown in Fig. 3. Except for grid points in the upper accumulation area, where the central flowline represents less well mean lateral conditions anyway, the model is very well capable of reproducing the right thickness and surface elevation. For the present glacier, mean modelled velocities then generally range between 40 and 80 m/year, with a maximum of 150 m/year in the strong convergence zone at x = 2000 m and 140 m/year at the ice fall at x = 8250 m. Taking into account that these velocities should be interpreted as lateral and vertical mean values, they appear to be slightly higher (although by a factor 1.5 at most) than the observations (Hantz, 1981).

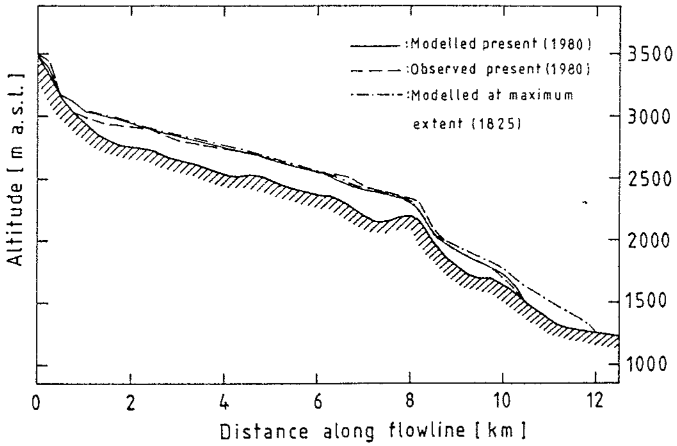

Figure 3. Longitudinal glacier profiles corresponding with the present glacier (10.49 km) and a glacier at maximum extent (11.98 km). Both profiles differ in the steady state with a balance shift of 0.37 m/year or an ELA shift of 52 m. The bed profile along the flowline is shown hatched

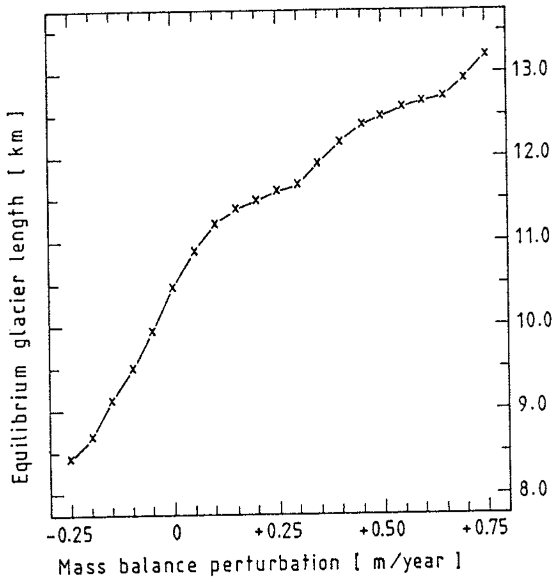

Figure 4.Steady state glacier length versus perturbations in the mass-balance with respect to the 1980 reference state. The total abscissa range of 1 m/year corresponds with a total ELA shift of 140 m.

Basic model sensitivity versus perturbations in the mass balance is reflected in Fig. 4. A noteworthy feature here is the differential behaviour with respect to positive or negative shifts in mass balance. The decreasing sensitivity encountered for positive balance perturbations appears to be, apart from an increased ablation gradient below 1750 m, essentially a geometric effect: the longer the glacier gets, the less important becomes the relative weight of the wide accumulation surface area in the total glacier area distribution. The asymmetric width distribution (large firn zone area and narrow glacier tongue) also explains the large sensitivity of the glacier terminus position with respect to the climatic state. A Little Ice Age glacier of maximum length 11.98 km and the 1968 minimum state of 10.30 km differ in the steady state by a mean change in net balance of 0.40 m/year only, or accordingly, a change in ELA or around 55 m. Of course, the concept of steady state is a theoretical one, and climatic inferences of this ELA-shifts are highly uncertain. Following theory developed by Kuhn (1980), a change in ELA of 100 m would correspond to about a 0.8 K temperature difference, a value derived for the eastern Alps. Typical thickness changes between these two glaciers are then about 10 to 20 m above the ice fall, and increasing to almost 150 m further below, as shown in Fig. 3.

Another basic parameter of Glacier d'Argentière concerns its response time, i.e. the order of time it takes to adjust to a change in its mass balance. This is investigated in Fig. 5, showing the response of the glacier following a stepwise change in mass balance. Defining this time as the time it takes for the glacier terminus to reach its final position within a fraction e^{-1}, the response time appears to be in the range 27-33 years for a positive perturbation and, the response being somewhat slower, 40-45 years for a negative mass balance change.

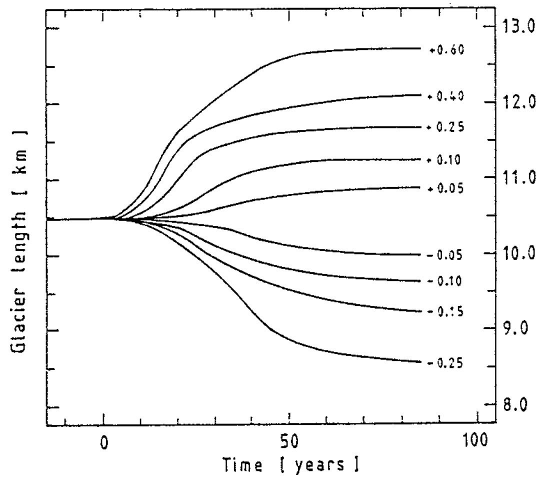

Figure 5. Reaction of the glacier terminus position to a stepwise change in mass-balance of given magnitude at time = 0.

Sensitivity of Glacier d'Argentière with respect to sinusoidal forcing around the 1980 reference state, is summarized in Figs. 6 and 7. Here again, the time lage between applied net balance forcing and terminus response (measured from minimum/maximum net balance extreme to the corresponding terminus extreme, Fig. 6) depends on the sign of the perturbation, but appears to be virtually independent of applied amplitude (with a maximum difference of 10% between amplitudes B = 0.25 m/year and B = 0.50 m/year). The interesting parameter here is the time lag at a 1000 year period, as it has been assessed for other glacier (Kruss, 1984). In the present case it is about 50 years (mean of max. and min. values), compared to 110 years for Hintereisferner (Austria) and 30 years for Lewis Glacier (Kenya). As in similar experiments discussed by Kruss (1984), the terminus response amplitude (Fig. 7) reaches a maximum for the longer periods, and depends linearly on applied amplitude. However, a doubling of the applied net balance amplitude does not result in a two-fold increase in terminus response, but rather by a mean factor of 1.75. So, the experiments discussed above clearly demonstrate basic sensitivity of Glacier d'Argentière with respect to changing climatic conditions. They also provide little evidence of basic model failures that could lead to fully wrong results.

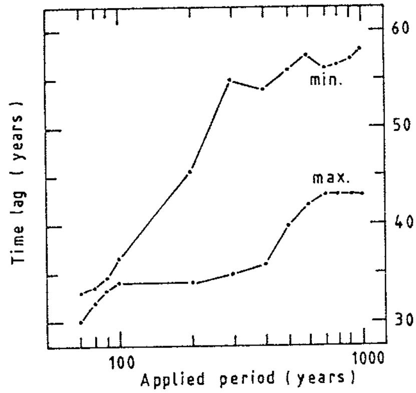

Figure 6. Time lage between sinusoidal net balance forcing and terminus position, respectively following a net balance maximum (max.) and minumum (min.). The net balance amplitude is 0.25 m/year here.

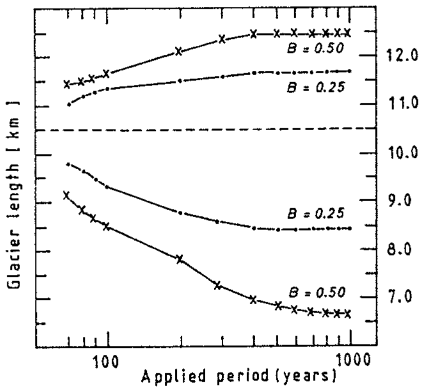

Figure 7. Amplitude of the Glacier d'Argentière's terminus response. Shown are the maximum and minimum glacier positions against applied period of net balance sinusoidal oscillations of amplitude B = 0.25 and 0.50 m/year respectively.

4. SIMULATION OF HISTORIC FRONT VARIATIONS

In order to get more insight in possible causes that might explain the variation in the terminus record and, in particular, the general retreat of Glacier d'Argentière since the middle of the last century, and attempt is then made to simulate the historic record. As mentioned in the introduction, a comparison between modelled and observed terminus histories is then expected to tell more about the type of climatic variables decisively affecting the glacier's mass balance. Naturally, mass balance measurements are not available over the time span covering the Little Ice Age. The same applies to instrumental climatic records for the first part of this period. Hence, a proper forcing function, in part based on proxy data, must be carefully chosen and relied upon.

In this study climatic data dating back to 1530 for the north face of the Alps are used (Pfister, 1984, personal communication). These data (seasonally and yearly precipitation and temperature) are given as 10-year running means and expressed as anomalies with respect to the 1901-1960 mean for Basel. It may be assumed that long-term climatic trends in the Mont Blanc Area are well described by this series, as the climate in the two locations is governed predominantly by the same large scale oceanic circulation types (a point also taken up in section 5).

Due to the fairly long characteristic response time scale of the glacier (of the order of 50 years) and the long time the glacier takes to reach equilibrium (approximately 300 years), calculations started at 1000 A.D. with zero ice thickness. The forcing function until 1530 is taken from the central England series (Lamb, 1977) and is made to match the Pfister series. It essentially shows a general cooling trend since about 1200. The last value of the Basel series (1970-1979 mean) is then kept constant until the year 2000, when calculations end.

Assuming that the linear balance model also applies to the past, the mass balance is perturbed in the experiments according to:

$$\Delta M = C_1 \ (\Delta T + C_2) \tag{7}$$

where ΔM is equal for all altitudes, ΔT is temperature anomaly (°C), and C_1 and C_2 are constants to be optimized. C_1 essentially controls the terminus response amplitude and the constant C_2 mainly influences the 'mean' glacier length. They are specified below.

It is generally accepted that mass balances of Alpine glaciers are to a large extent controlled by meteorological conditions in summer, in particular summer temperature (e.g. Reynaud, 1983). On the basis of a 27-year record, Martin (1977) found that 58% of the total variance in the mean specific net balance record for Glacier de Sarennes could be explained by summer temperatures (July and August) at the nearby Lyon station, with minor contributions in the total variance from winter precipitation (October to May) and early summer conditions (June precipitation).

Hence, in a first experiment summer temperature anomalies were used to force the glacier model. Results with $C_1 = -0.6$ m/°C and $C_2 = -0.3$ °C are shown in Fig. 8. Striking features in the forcing function (upper panel) include a marked cooling around 1600, a very cold interval around 1816 (year 'without summer') and to a lesser extent in 1886 and 1912, and a steep temperature rise between 1912 and 1947. The observed front record, based on continuous measurements over roughly the last 100 years and on dated moraines and sporadic written sources before that, has been described in the introduction (middle panel). The lower panel shows that the modelled response does not resemble the observed length variations very well, even when taking into account that only the long term climatic trend is expected to be reflected in the glacier response, as the glacier effectively acts as a low-pass filter. The model then shows a marked glacier increase towards 1600, approximately coinciding with the beginning of the Little Ice Age. Between 1600 and 1800, the glacier is, unlike the observed record, almost continuously receding, reaching a terminus position around 1810 rather close to the present state. Note that the forcing does not display remarkably cold conditions during this period either. Thereafter, the modelled glacier response exhibits an important growing phase, obviously related to the 1816 cold summer spell. The observed glacier retreat since 1850 then comes almost a century later.

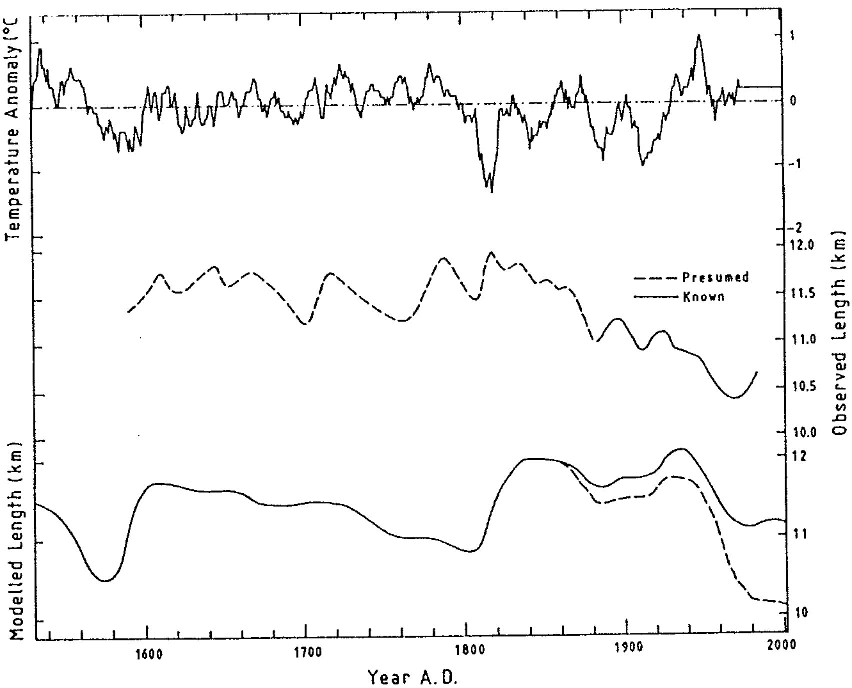

Figure 8. The experiment in which summer temperature anomalies (upper panel) were used to construct the mass balance history. The middle curve shows the observed length variations. The lower curve shows computed glacier length with full line: C_1 = -0.6m/°C, C_2 = -0.3°C and dashed line: C_2 = -0.1°C from 1850 onwards.

As displayed in Fig. 9, somewhat better agreement between modelled and observed histories can be obtained when relating mass balance perturbations to more general atmospheric conditions, in casu mean annual temperature. Nevertheless, both simulations appear to have in common that the Neoglacial maximum and concomitant retreat are out of phase by 50-100 years.

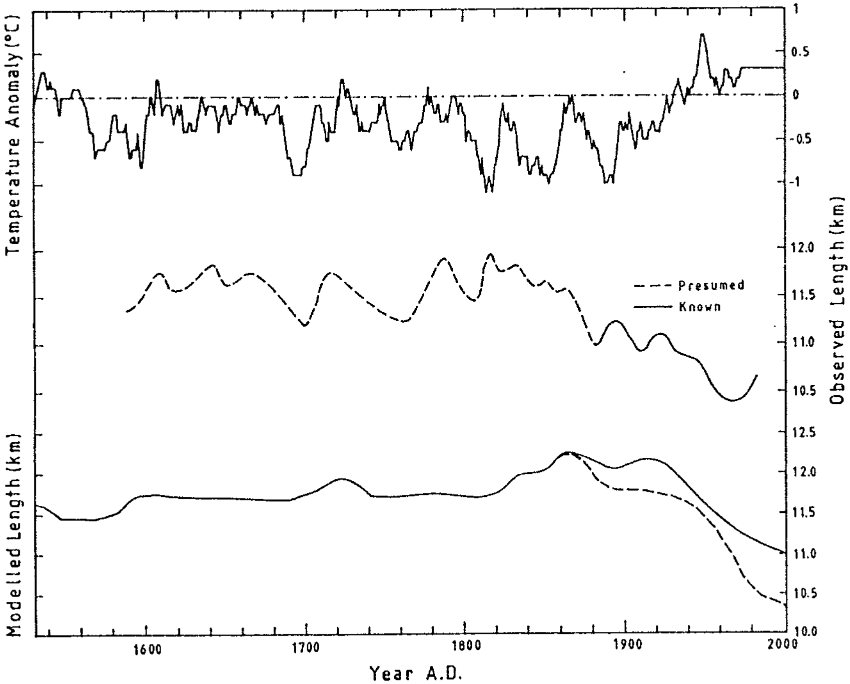

Figure 9. As in Fig. 5, but now with mean annual temperature anomalies as forcing function. C_1 = -0.45 m/°C, C_2 = -0.4°C (full line) and C_2 = -0.2°C from 1850 onwards (dashed line).

This point has been taken up in an experiment in which a (moderate) additional mass balance perturbation of around -0.1 m/year was imposed from 1850 onwards (dashed line, see Figs. 8 and 9), which turned out to significantly improve the simulation, in particular in the model run forced with mean annual temperature anomalies. In a recent paper, Oerlemans (1986) demonstrated that melting rates at the lower glacier parts may be extremely sensitive to changes in the radiation budget (through increased advective heat transport from surrounding rock areas), providing a possible link between glacier retreat and increased carbon dioxide levels in the atmosphere. This is certainly not in contradiction with the improved results obtained here, when crudely mimicing increased ablation since the 19th century. Similarly, assuming increased balance gradients during the same period also turned out to yield better agreement between modelled and observed glacier retreat (not shown here). Many more model runs were conducted, involving forcing functions based on other precipitation and temperature data. In a similar way as in Martin (1977), also composed series were derived from a multiple linear regression analysis of the Basel-data with available specific net mass balance records (Glacier de Sarennes, Aletschgletscher). However, as these model runs do no add much to the general picture, they are not discussed here.

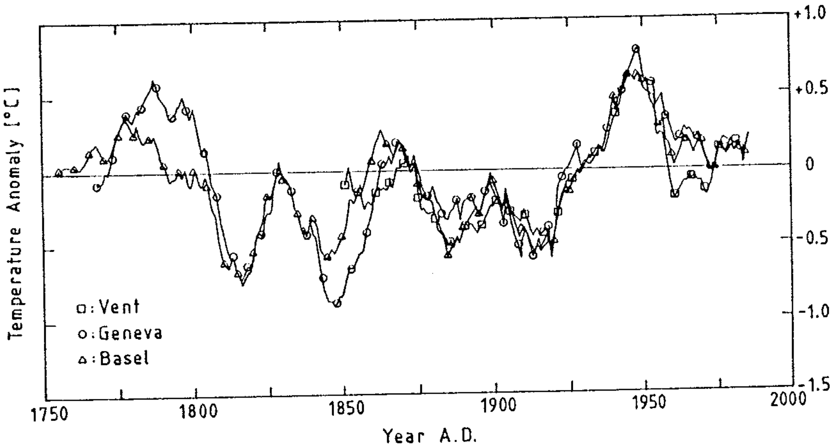

Figure 10. 15-year running mean summer temperature anomalies with respect to the 1901-1950 mean compared with the Basel series. The summer temperature is defined here as the mean over the three summer months (June, July and August) with contributions from May and September witha weighing coefficient of 0.5. Plot kindly provided by Wouter Greuell.

5. DISCUSSION

As is clear from the results in Figs. 8 and 9, none of the simulated glacier response curves compares really well with the recorded historic positions of the terminus. Numerical experiments, in which model parameters and data on the glacier were varied within realistic ranges (ice thickness, flow parameter, inclusion of variable shape and velocity factors) brought to light that the discrepancy between modelled and observed terminus histories cannot be due to errors in these parameters. Rather changes in these lead to re-adjustment of velocity and/or ice depth along the flowline, while the terminus response remains largely unaffected, in essence reflecting the integrated changes in overall mass balance only. For the same reason, also the inclusion of longitudinal stresses, providing a potential additional means of propagating disturbances along the glacier, is not expected to influence the glacier front response significantly. Moreover, one may anticipate local effects of extension and compression to cancel each other out, when changes in total glacier length are considered. This is corroborated by experiments conducted by W. Greuell on Hintereisferner (personal communication), showing the response time to be quasi unchanged by longitudinal stress effects.

This means that the unsuccessful simulation must be due to an inadequate description of the mass balance history. In Fig. 10 it is shown that there can be no doubt about the suitability of the Basel-series to describe climatic trends in the Mont Blanc area. In particular since 1875, there is a striking resemblance between summer temperature anomalies of Basel (north of the Alps), Geneva (west of the Alps) and Vent (Oetztal, central Alps). So, as the response curve shape turns out not to depend critically on the parameters C_1 and C_2, the obvious conclusion should then be that an oversimplified relation between climate and mass balance is put forward and that apparently, mass budget imbalances of Glacier d'Argentière are not linearly correlated with summer temperature anomalies on the time span considered.

6. CONCLUSION

In this study an attempt was made to investigate possible causes of glacier variations by simulating the historic record of Glacier d'Argentière with a dynamical model. Forcing the mass balance history linearly (spatially as well as in time) with summer and mean annual temperature anomalies for Basel, brought to light, that, in particular, the observed glacier retreat since about 1850 is not fully understood.

Apart from basic shortcomings in model formulation and glacier data, for which there is otherwise no direct evidence, it appears that the critical point in this study is connected with the assumption that the type of variation evident from a relatively short series of mass balance measurements can be extrapolated to past environments. Actually, the results of the simulations seem to suggest the contrary. Although that on the basis of available records, summer temperature seems to be highly correlated with mass balance variations that are in addition independent of altitude, this relationship does apparently not explain the dominant feature of the historic record, in casu the almost continuous retreat of the Argentière since the middle of the 19th century.

This result and the improved model simulations that could be obtained while assuming an additional negative mass balance perturbation during roughly the last 150 years, seems to point to additional features affecting the glacier's mass balance that are not captured well in the ambient climatic records. A thorough investigation of this point is far beyond the scope of the present study, but it may very well be that melting rates, and hence, balance gradients respond critically and non-linearly to changes in the radiation budget, valley geometry or glacier surface conditions much along the lines of what has been demonstrated in Oerlemans (1986).

ACKNOWLEDGEMENTS

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