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An Attempt to Simulate Historic Front Variations of Nigardsbreen, Norway

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With 8 Figures

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Summary

Nigardsbreen (Norway) is one of the very few high-latitude glaciers from which a long record of front positions is known (starting in the beginning of the 18th century). In this paper a dynamic glacier model is used to investigate the possible causes of the observed front variations. These variations are characterized by a rapid advance until 1748, and steady retreat, with some minor interruptions, after that time.

Two time series are used as forcing functions that drive the dynamic model: (i) central England summer temperature, and (ii) a tree ring record from northern Sweden. It appears that forcing (i) does not work at all: an almost continuously growing glacier is predicted by the model. Forcing (ii) performs much better, although the maximum glacier extent comes too early.

Interpretation of these results and some additional experiments suggest that the climatic warming after the little ice age has been much more pronounced in western Norway than in England, at least in summer. Winter precipitation in Norway probably decreased gradually during the last centuries.

Zusammenfassung

Ein Versuch, die historischen Gletscherbewegungen des Nigardsbreen (Norwegen) zu simulieren

Nigardsbreen (Norwegen) ist einer der wenigen Gletscher in hohen Breiten, von dem langjährige Aufzeichnungen über die Frontpositionen bekannt sind (seit Anfang des 18. Jahrhunderts). In der vorliegenden Arbeit wird zur Untersuchung der möglichen Ursachen der beobachteten

Gletscherfrontveränderungen ein dynamisches Gletschermodell verwendet. Diese Veränderungen sind bis 1748 durch ein rapides Vorrücken und nach dieser Zeit durch einen, mit einigen kleinen Unterbrechungen, stetigen Rückgang gekennzeichnet.

Als Antriebsfunktionen, die das dynamische Modell treiben, werden zwei Zeitreihen verwendet: 1. Die Sommertemperaturen von Mittelengland, und 2. Jahresringaufzeichnungen aus Nordschweden. Es zeigt sich, daß der erste Antrieb überhaupt nicht funktioniert, denn dieses Modell sagt einen fast kontinuierlich wachsenden Gletscher voraus. Der zweite Antrieb funktioniert viel besser, obwohl die maximale Gletscherausdehnung zu früh auftritt.

Die Interpretation dieser Resultate und einige zusätzliche Experimente weisen darauf hin, daß die Erwärmung nach der kleinen Eiszeit im westlichen Norwegen viel ausgeprägter war als in England, wenigstens im Sommer. Der Winterniederschlag hat sich vermutlich in Norwegen während der letzten Jahrhunderte allmählich verringert.

1. Introduction

Glaciers are generally thought to be very good climate indicators, at least when a time scale of decades to centuries is of interest. Since the larger alpine glaciers have wide accumulation basins and long, relatively narrow tongues, a slight change in the mass-balance conditions generally leads to significant variations in the

position of the snout. The equilibrium response is basically determined by the geometry, in particular the area-elevation distribution (e.g. Furbish and Andrews 1984).

A large number of studies has been carried out that relate glacier variations, and also mass-balance variations, to climatic conditions (e.g. Hoinkes 1968; Reynaud 1983; Letreguilly 1984; Young 1977). These studies have without doubt confirmed that a broad coupling exists between the degree of glacierization and climatic conditions in summer. The general tendency in a specific region, measured for instance by the percentage of advancing glaciers can thus be explained well.

The behaviour of individual glaciers is also of interest, however. A number of glaciers exist, of which front positions have been recorded, in one way or another, for quite a long time. Examples are the Argentière, Rhone and Untere Grindelwald glaciers in the Alps, having a fairly reliable record from 1600 onwards. When a good understanding of how these glaciers respond to "climatic forcing" could be achieved, these records would certainly help to get a more complete picture of historic climatic variations.

It has been recognized by many workers that both the change in mass balance due to

varying meteorological conditions, and the subsequent response of a glacier, are nonlinear, transient processes (e.g. Kuhn 1978). It is thus natural to employ a dynamical model when the behaviour of a specific glacier is investigated. A few attempts have been made in this direction, for instance by Kruss (1984), based on the numerical glacier models developed by Budd and coworkers (e.g. Budd and Jenssen 1975). In the present paper a similar approach is used to study Nigardsbreen, Norway.

This glacier is of special interest, because:

(i) It is one of the very few high-latitude glaciers having a long record of front positions (from about 1700).

(ii) It has a narrow tongue and a very wide accumulation basin, so the terminus position is extremely sensitive to changes in the mass budget.

(iii) It is part of the ice cap Jostedalsbreen, which covers an area of almost 500 km², and may have been a nucleus for growth of the Fennoscandian Ice Sheet in the initial stage of an ice age.

The geometry of the ice cap is obvious from Figs. 1 and 2. Fig. 2 also shows the flow line and grid points used in the ice-flow model employed in this paper.



Fig. 1. Air photograph (looking in northwesterly direction) of the small ice cap Jostedalsbreen, of which Nigardsbreen is a part (outlet glacier at the right-hand side). Widerøe's photo 223853

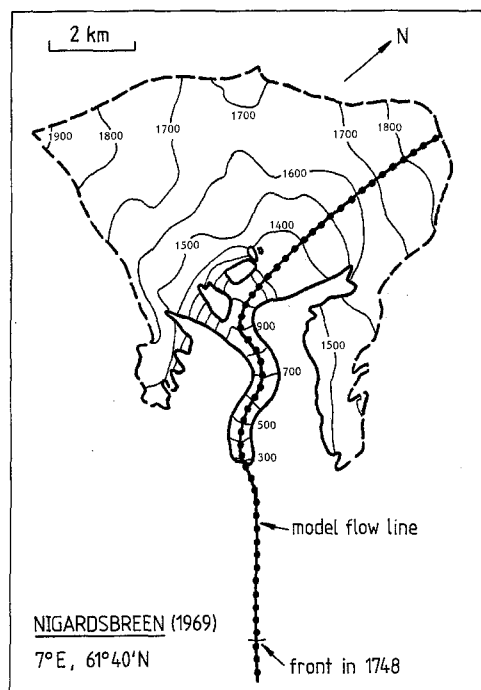


Fig. 2. Map of Nigardsbreen. The flow line with grid points along which calculations are made is shown (grid point spacing is 300 m). The maximum extent of the glacier (1748 AD) is indicated by a bar. For bed profile along the flow line, see Fig. 4

For a review of glaciological work carried out on Nigardsbreen, see Østrem et al. (1977). Here, only a few points are mentioned. In historical times, Nigardsbreen reached its maximum extent in 1748, and from the period 1700–1748 a number of written documents exist reporting on a very rapid advance of the glacier front. During this time, cultivated land and several houses were destroyed. Since 1748 the glacier has shown a steady retreat, with some interruptions around 1905, 1925 and 1936. It is noteworthy that the sequence of front positions deviates substantially from those known for glaciers in the Alps, in particular in the 18th century.

In this paper a computer model of Nigardsbreen is presented that allows a time-dependent simulation of terminus position and total volume, as a response to any prescribed forcing. With this model several experiments have been conducted. Apart from some examples concerning response time and sensitivity to changes in the mass balance, some runs will be discussed

in which an attempt is made to simulate the past record of front variations. Several forcing functions are used here, for instance the summer temperatures from the well-known central England series (Lamb 1977). It will turn out that the attempt is moderately successful, and possible reasons for the discrepancy between observed and simulated terminus variations are discussed.

2. Description of the Model

A one-dimensional ice flow model is used, with the independent variable x along the flow line, shown in Fig. 2. In the lower part of the glacier, the choice of a flow line is a straightforward matter, but in the accumulation basin ambiguity cannot be avoided. Most important, of course, is that the schematic geometry chosen for the model is such that the area-elevation distribution is not distorted too much. The glacier width should thus depend on x , but also on the ice thickness. The latter is necessary, because a lower glacier tongue with fixed width would lead to serious errors in the total ablation. For instance, in reality the thinning of a glacier tongue generally implies a decreasing width and, consequently, a smaller ablation area.

The dynamic behaviour of the glacier is described in terms of changes in ice thickness, which have to be calculated from a continuity equation. Ice-density variations are not considered, so a conservation equation for ice volume can be used

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

The geometry is shown in Fig. 3. In eq. (1), t is time, S the area of the cross section perpendicular to the flow line, and U the mean ice velocity in the cross section (parallel to the bed). So the first term on the right-hand side represents the divergence of the volume flux. The last term stands for net accumulation at the surface. M is the annual mass gain (or “volume gain”, rather), and B_s the width of the glacier at the surface.

To obtain an equation for changes in ice thickness, the glacier cross profile is assumed

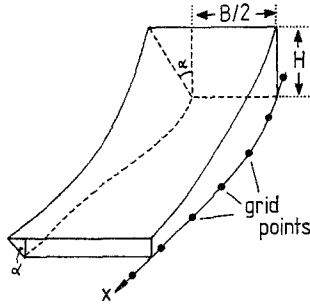


Fig. 3. Geometry for the ice-flow model. B and α are specified functions of x (see Appendix). Only half of the cross profile is shown, because symmetry with respect to the x -axis is assumed

to be reasonably well represented by a trapezoidal shape (one could also use a power fit, as done for instance by Kruss 1984). The width at the bed is denoted by B . Both B and the angle α (see Fig. 3) are prescribed functions of x . The area S of the cross section is now related to ice thickness H by

$$S = H [B + H \tan(\alpha)] = H(B + \mu H). \quad (2)$$

Substitution in eq. (1) yields a rate equation for H . It reads

$$\frac{\partial H}{\partial t} = \frac{-1}{B + 2\mu H} \left[(B + \mu H) \frac{\partial(UH)}{\partial x} + UH \frac{\partial}{\partial x} (B + \mu H) \right] + M. \quad (3)$$

So in this equation geometric effects are fully taken into account, notably (i) the effect of narrowing and widening of the valley through which the glacier flows on the changes in ice thickness, and (ii) the effect of increasing glacier width when the ice thickness increases, depending on the steepness of the sides of the valley. Note that for a glacier of constant width eq. (1) reduces to

$$\frac{\partial H}{\partial t} = - \frac{\partial(UH)}{\partial x} + M. \quad (4)$$

To complete the description of the glacier dynamics, an equation for the mean ice velocity U is needed. It is assumed that the glacier is temperate. The total ice velocity consists of a contribution from internal deformation (U_d) and sliding (U_s). Each of those is assumed to be determined by the driving stress τ , given by

$$\tau = -\rho g H \frac{dh}{dx}. \quad (5)$$

Here ρ is ice density, g acceleration of gravity and h surface elevation. The flow law used reads

$$U = U_d + U_s = f_1 H \tau^3 + f_2 \tau^3 / (N - P). \quad (6)$$

Here N is the overburden pressure of the ice, corrected for subglacial water pressure P when significant. In this study the contribution from P is neglected, however. For a discussion of flow laws and their application in glacier and ice-sheet modelling for climate studies, see for instance Paterson (1981), and Oerlemans and van der Veen (1984).

The flow parameters f_1 and f_2 are not accurately known. They depend on bed conditions, debris content and crystal structure of the basal ice layers. The values adopted here are

$$f_1 = 0.95 \times 10^{-22} \text{ m}^6 \text{ S}^{-1} \text{ N}^{-3}, \\ f_2 = 0.9 \times 10^{-14} \text{ m}^5 \text{ S}^{-1} \text{ N}^{-2}.$$

The only justification for using these values admittedly is that they have yielded good results in a number of studies and lie within the range of uncertainty found in the literature (see also Paterson 1981, for a discussion on this matter).

A full description of the numerical scheme will not be given here. It suffices to mention that a staggered grid is used for the computation of the flux divergence along the flow line. Time integration follows the simple forward scheme, which, for sufficiently small time steps, is stable for diffusion-type equations (e.g. Mesinger and Arakawa 1976).

3. Some Basic Experiments

Before any integrations can be carried out, the mass balance should be specified as a function of elevation h . The basic parameterization, which will be perturbed in the various experiments to be discussed later, is a piecewise linear fit to the observed mean mass balance for the period 1962–1975 (Østrem et al. 1977), in m ice depth/yr:

$$\begin{aligned} L < 900 \text{ m}: & \quad M = -5 + 0.012 (h - 900), \\ 900 < L < 1700 \text{ m}: & \quad M = 1.67 + 0.0083 (h - 1700), \\ L > 1700 \text{ m}: & \quad M = 1.67 + 0.0065 (h - 1700). \end{aligned} \quad (7)$$

In this 14-year period, the average height of the equilibrium line was close to 1500 m. Note that

the mass-balance gradient in the lower part of the glacier is rather large. In the highest parts of the firn basin, the mass balance reaches values of almost 3 m/yr. Østrem et al. (1977) have calculated that the specific annual balance for this period amounts to about 0.5 m ice depth. So relative to the observation period, the glacier length seems to be smaller than the equilibrium value.

Other input data, namely bed topography, glacier width B , and the coefficient μ , are given in the Appendix. The bed topography, in particular, forms a problem. It is not known for large parts of the glacier. In these regions the bed was adjusted until the model produced a glacier surface close to the presently observed one (although it is unknown how far this is from equilibrium!). Since the upper grid points represent wide areas, it is not clear how the proper bed elevation should be chosen for the model anyway, even if the real bed would be known accurately. It was found, however, that small changes in the upper bed profile have little effect on the response of the model glacier to a changing mass balance.

In a first experiment, the model was run to a steady state, starting from zero ice volume, with the mass balance M as given by eq. (7). Fig. 4 shows the result, by means of some profiles along the flow line. Apparently, an equilibrium is reached after about 300 years of simulated time. The model glacier is about 2 km longer than the presently observed one, which is in accordance with the positive specific balance mentioned above. In fact, the terminus position is very close to the 1937 position. The 1937-profile, as presented by Østrem et al. (1977) is indicated in the figure by the dashed line. Ice velocities in the model depend on x , of course, but are typically in the 50 to 120 m/yr range, with a maximum value of somewhat more than 200 m/yr in the ice fall (at $x = 5200$ m, see also Fig. 1). These velocities are somewhat higher than reported in Østrem et al. (1977), which should be expected in view of the fact that the equilibrium length is larger than the currently observed one.

Other experiments showed that the equilibrium position of the terminus is very sensitive to changes in the elevation of the equi-

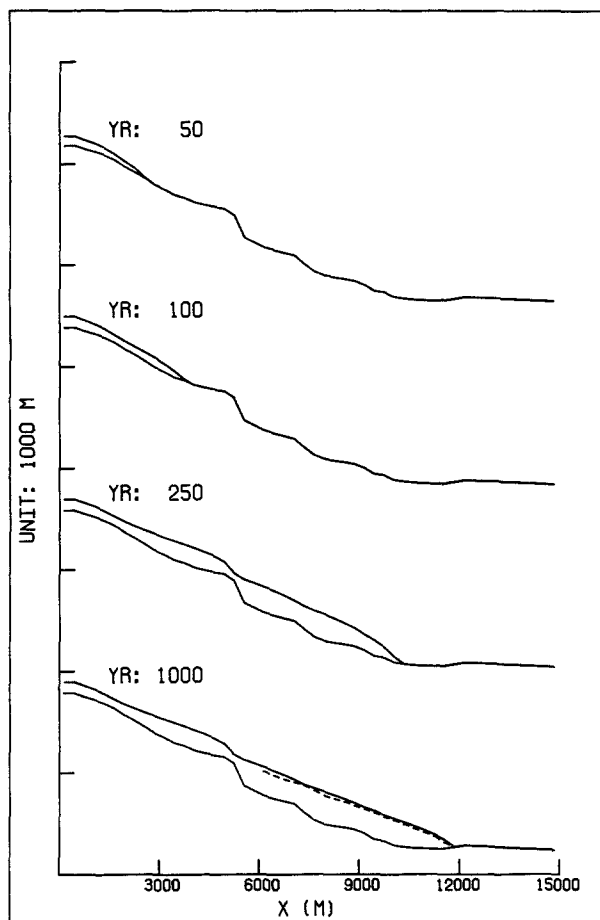


Fig. 4. Profiles for an integration with fixed mass balance conditions, as observed in the periods 1962–1975, with no ice as initial condition. The equilibrium profile produced by the model is very close to the profile observed in 1937, which is indicated by the dashed line. The position of the terminus is currently around $x = 9500$ m

librium line. As mentioned earlier, this is a purely geometric effect (large firn basin, narrow valley glacier), and the present model does not add much to this.

Of more interest are transient effects. For the sake of illustration, Fig. 5 shows a result from an experiment in which the mass-balance gradient was suddenly changed. This was done by adding a perturbation M' to the mass balance given above, according to

$$M' = 0.0025 (h - 1565) \text{ m/yr} . \quad (7a)$$

This hardly changes the steady-state terminus position, but causes a significant transient effect. In this experiment, the model is first integrated for 600 yr to obtain a perfect equi-

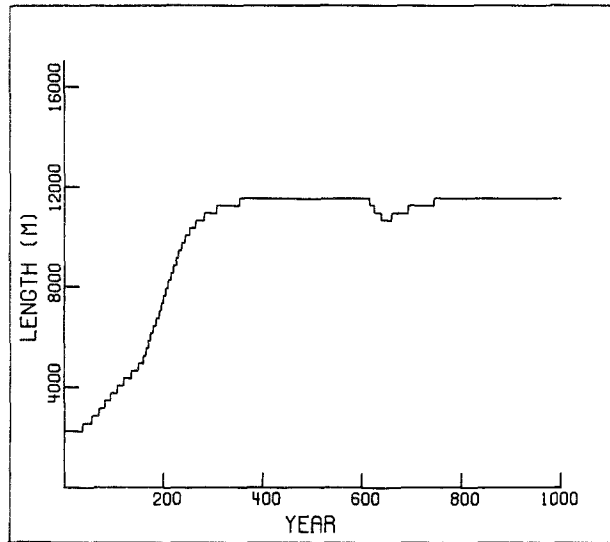


Fig. 5. Glacier length for the experiment with a sudden change in mass-balance gradient, imposed after 600 years of simulated time. To show the model's resolution in a proper way, no refined calculation of the terminus position was included

librium. Then the mass balance perturbation is imposed, and the glacier reacts by retreat of the snout over a distance of about 900 m. After 50 years, however, the accumulated mass in the firn basin starts to influence the terminus position, and after another 100 years the new steady state is reached. This example illustrates how complicated a glacier may react to changing environmental conditions.

Experiments with periodic forcing were also carried out, but results will not be discussed here because they are very similar to results from previous studies (e.g. Nye 1965; Budd and Jenssen 1975; Kruss 1984). We thus turn to the heart of the matter, the simulation of the historic record of front variations.

4. Simulation of Front Variations

As mentioned in the introduction, terminus positions of Nigardsbreen are known from the beginning of the 18th century. The glacier reached its greatest length in 1748, and has shown steady retreat since that time. Around 1950, the rate of retreat became sufficiently larger. Observed front positions are shown in

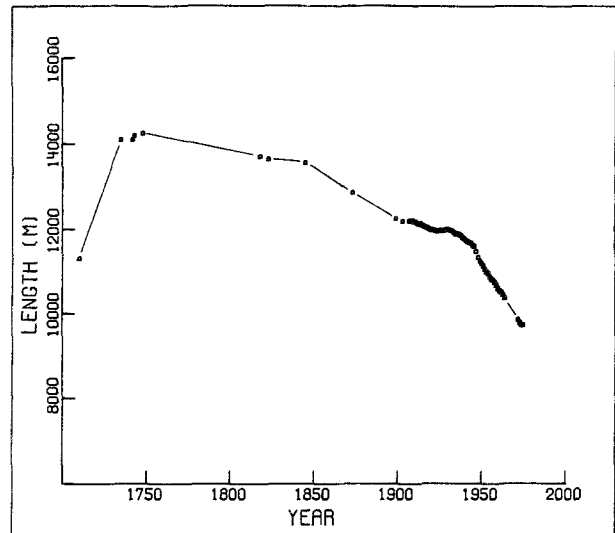


Fig. 6. Observed positions of the glacier front. In the period 1964–1972, accurate positions are not known because the glacier showed rapid irregular retreat in a lake

Fig. 6, from the data presented by Østrem (1977).

There are a number of problems when one tries to simulate such a record. First of all, a proper forcing function is needed. Direct mass-balance observations are not available, and over the time span of interest even reliable meteorological measurements are scarce (being a reason, in fact, why glacier front variations are so interesting). A second problem is associated with the response time of the glacier. To arrive at a proper simulation, the integration should start a few centuries before the record of observed front variations begins. However, this gives additional problems concerning the forcing function.

In this study, it is assumed as a starting point that the mass-balance perturbation connected with changing climatic conditions is independent of elevation. For most glaciers, this assumption is not valid, but it seems to work reasonably well for Nigardsbreen (Kuhn 1984). The first forcing function to be used is based on the central England summer temperatures, combined from various sources into a single series by Lamb (1977). This series is shown in Fig. 7.

The mass-balance perturbation is now formulated as

$$M' = C_1 (T_{ce} - C_2), \quad (8)$$

where T_{ce} is central England summer temperature, and C_1 and C_2 are constants. The former is estimated from the characteristic mass-balance gradient and the atmospheric temperature lapse rate [(i.e. $dM/dT = (\partial M/\partial h) \cdot (dh/dT)$]. This yields a value of -0.63 m ice depth/K. The constant C_2 is chosen such that the maximum glacier length produced by the model is close to the observed maximum length.

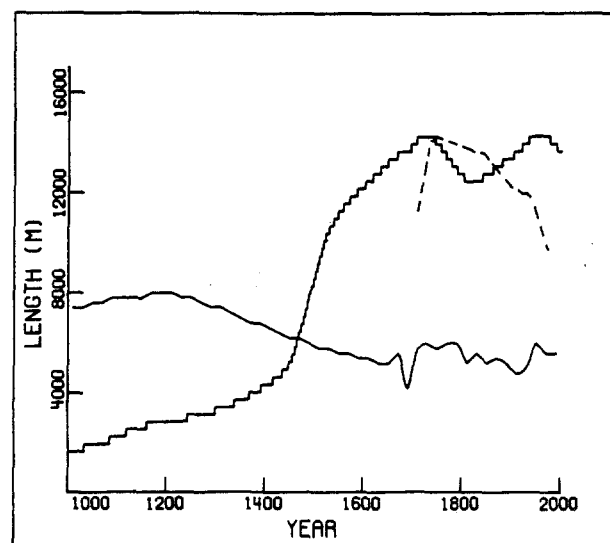


Fig. 7. The experiment with central England summer temperature as forcing function (20 yr mean values). The forcing is shown by the lower curve (arbitrary scale, maximum summer temperature: 16.5°C around 1200 AD; minimum: 14.6°C around 1690 AD). The other curve shows computed glacier length; the observations are given by the dashed line

The forcing function displayed in Fig. 7 consists of two parts. From about 1650 the data are based on real temperature measurements, but before that time only general trends are known from proxy data (Lamb 1977). The most striking feature is the general cooling starting in the late middle ages, culminating in the well-known little ice age.

As shown by the stepwise curve, imposing this forcing function to the model yields disappointing results. The integration starts at 1000 AD. In the first 300 years of simulated time the glacier is short, which is not only due

to the fact that it takes time for the glacier to grow, but also because the large sensitivity of front variations to changing climatic conditions occurs for somewhat lower temperatures. This is clear from the very rapid advance in the 15th century, when central England summer temperature drops slowly below 16°C . The large sensitivity of the terminus position around this temperature is essentially a geometric effect.

Apparently, there is no similarity between observed and simulated front positions. The model does not predict any retreat. Some sensitivity tests showed that this result cannot be due to errors in model formulation or geometric input. The conclusion should thus be that there is no correlation between mass balance conditions at Nigardsbreen and summer temperatures in England.

In a second experiment, the forcing function was derived from data on tree ring width. In Scandinavia, a global correlation between tree ring width and glacier advance/retreat seems to be present (e.g. Karlén 1984). Although the interpretation of variation in tree ring width is not a straightforward matter, there are reasons to believe that in a maritime high-latitude climate summer temperature plays a very important role. The forcing function used here was taken from Karlén (1984), and is displayed in Fig. 8. It is a composed curve for northern Sweden, in fact. The mass-balance perturbation was formulated in a way similar to the first experiment: $M' = D_1 (W - D_2)$. (9)

Here D_1 and D_2 are constants to be optimized, and W is tree ring width. The simulation shown in Fig. 8a was carried out with $D_1 = -0.5$ m ice depth per 0.1 mm tree ring width and $D_2 = 0.5$ mm.

A comparison of Fig. 8a with Fig. 7 makes clear that the simulation is much better in the second experiment. The maximum extent around 1750 AD and the subsequent retreat show up reasonably well, although the simulated curve is somewhat out of phase with the observed front positions. The model predicts a readvance starting around 1900 which did not occur in reality.

A possibly important factor that has not yet been included in the model is calving in situations where the glacier snout ends in a lake.

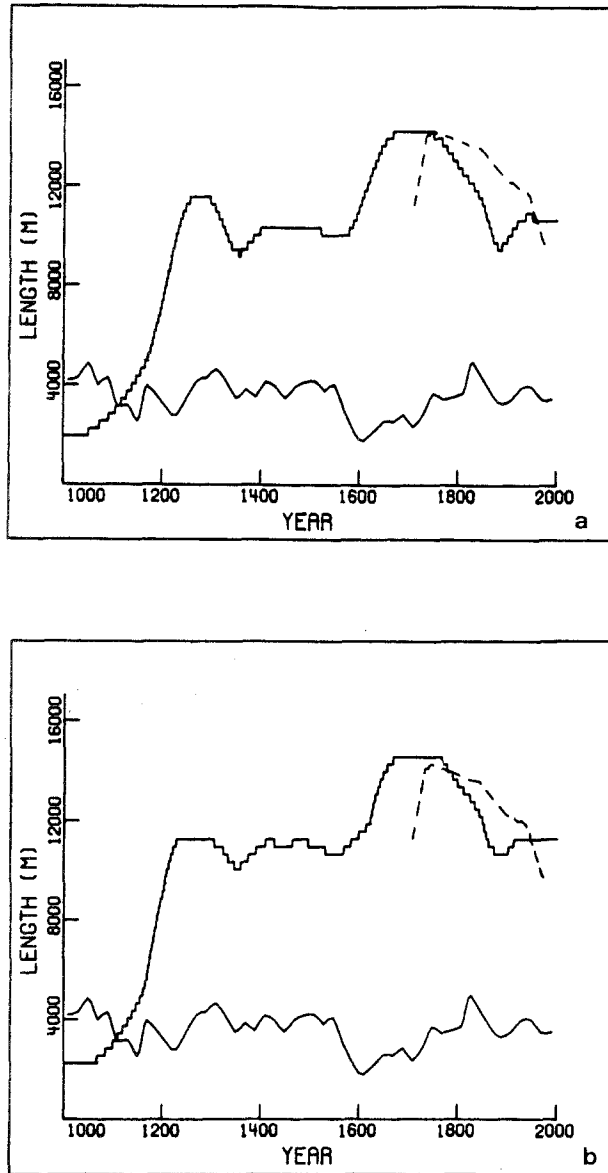


Fig. 8. As in Fig. 7, but now with tree ring width as forcing function (lower curve, maximum value 0.99 mm, minimum value 0.43 mm). Experiment b is with calving, experiment a without (with adjusted mass-balance parameters, see text)

Concerning the front position of Nigardsbreen, this definitely is an important process (Liestøl, personal communication). In order to see whether introduction of a schematic calving process in the model would lead to a better result, a third experiment was carried out. An additional perturbation of the mass balance was applied at the last grid point where ice is present, according to

$$M'' = \min [0, -QH_i(b_{i+1} - b_i)] \quad (10)$$

Here b is the bed elevation, and i the last grid point with ice. Q is the “calving parameter”. As can be seen in Fig. 4, the only location where a lake may form is around $x = 11000$ m. The presence of this lake caused the extremely rapid retreat of the front from 1940 until 1968, when the snout withdrew from the lake.

Using $Q = 0.1 \text{ (m yr)}^{-1}$ yields the result displayed in Fig. 8b. In this run, D_2 was adjusted (+ 0.05 mm) to arrive at the correct maximum glacier length again. The difference with Fig. 8a is not very large. Inspection of the model output reveals that during glacier advance/retreat the snout tends to stick at the upstream side of the lake, which is not unexpected. However, the similarity between observed and calculated front positions is not improved in a significant way. Of course, one should bear in mind that the present grid-point model is not very suitable for incorporating such a subtle process as calving. Other values of Q were tried, but this did not lead to any further improvement.

As demonstrated in Fig. 5, the response of the glacier front depends on changes in the mass balance gradient as well. One may anticipate that a mass balance perturbation depending on elevation would give a better result. A test run showed that this is indeed the case to some extent. Taking the tree ring series as forcing function, and assuming a negative mass balance perturbation below the equilibrium line and a positive perturbation above the equilibrium line (for a negative perturbation of tree ring width) yields, after tuning, a simulated curve that has a better phasing. However, there are no direct indications that such a type of formulation is justified.

5. Concluding Remarks

In this paper an attempt was made to study glacier front variations by means of a dynamic model. Since the objective was to cover a period of three centuries for which front positions were available, proxy data had to be used to construct the forcing function. So far, studies of

this type have only been carried out for shorter periods of time (at most one century, see references given in the introduction) for which relevant meteorological observations are available. A comparison with other studies is therefore difficult.

From sensitivity experiments carried out with the Nigardsbreen model presented here little evidence was found that it has basic shortcomings that would lead to fully wrong simulations. It is therefore justified to reverse the argument: the climatic variations in western Norway probably deviated in a significant way from those in England, and, to a lesser extent, from those in northern Sweden. The results suggest that the "post-little ice age warming" was much more pronounced in Scandinavia than in England.

A general problem with regard to interpretation of front variations of Nigardsbreen is the dependence of the mass balance on winter accumulation. In contrast to glaciers in the Alps, this is believed to be an important effect (Liestøl, personal communication). With this in mind, a few additional experiments were carried out with a forcing function describing variations in precipitation as suggested by Lamb (1977). Again, this did not lead to any significant reduction of the discrepancy between observed and calculated front variations. So, in summary, the results obtained here suggest that during the last two centuries climate in western Scandinavia was warmer in summer and drier in winter (or may be warmer and drier in general) as compared to central England.

Finally, it seems natural to study other long series of front positions in a similar way, in particular those of glaciers in the Alps. A comparison may then lead to a more coherent picture of the causes of glacier variations in Europe, and this will certainly add to our understanding of climatic change in historic times.

Acknowledgements

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Appendix

The following values were used for bed topography (in m), bed width B (in m), and the coefficient μ

Point	Top.	B	μ	Point	Top.	B	μ
1	1840	1800	0.0	26	540	500	0.6
2	1790	2700	0.0	27	500	450	0.8
3	1755	4650	0.0	28	485	450	0.8
4	1725	6500	0.0	29	470	400	0.9
5	1690	8650	0.0	30	447	400	0.8
6	1645	9600	0.0	31	410	400	0.7
7	1580	13000	0.0	32	350	350	0.8
8	1525	14500	0.0	33	335	400	1.05
9	1465	14000	0.0	34	290	450	1.1
10	1400	9600	0.0	35	275	450	1.1
11	1355	7200	0.0	36	265	450	1.2
12	1300	6400	0.0	37	260	450	1.3
13	1270	5600	0.0	38	252	450	1.3
14	1225	4900	0.0	39	260	400	1.3
15	1200	4400	0.0	40	275	400	1.2
16	1180	4000	0.0	41	290	400	1.2
17	1165	2200	0.1	42	285	400	1.2
18	1100	1400	0.3	43	280	450	1.2
19	880	1200	0.4	44	275	450	1.2
20	830	1025	0.5	45	270	500	1.2
21	780	1000	0.4	46	265	500	1.2
22	750	900	0.1	47	260	500	1.3
23	725	725	0.2	48	255	500	1.3
24	700	600	0.4	49	250	500	1.3
25	615	500	0.6	50	245	500	1.3

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