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MODELING GLACIER FLUCTUATIONS IN THE SØR RONDANE, DRONNING MAUD LAND, ANTARCTICA

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With 8 figures

ABSTRACT

Morainic deposits found above the present glacier surface in the central Sør Rondane Mountains give evidence of former glaciations. In this paper, an attempt is made to interpret these observed glacier fluctuations in terms of environmental change. To do this, a numerical flowline model, taking into account thermodynamics and a coupled ice shelf, has been developed and is applied to two outlet glaciers through the mountain range, Gunnestadbreen and Jenningsbreen. It is found that lower ice temperatures, reduced accumulation, and a drop in sea level corresponding to typical glacial conditions account for a 150—200 m rise in glacier level. From a comparison of these results with a palaeogeographical reconstruction for a "maximum" glacial stage for Jenningsbreen by Hirakawa et al. (1989) it is argued that their morainic deposits relate to an earlier Cenozoic glaciation involving a full grown ice sheet.

MODELLIERUNG VON GLETSCHERSCHWANKUNGEN IM SØR RONDANE GEBIRGE IN DRONNING MAUD LAND, ANTARKTIS

ZUSAMMENFASSUNG

Moränenablagerungen, die in den zentralen Sør Rondane Bergen über der heutigen Gletscheroberfläche gefunden wurden, sind Zeugen früherer Vergletscherungen. In der vorliegenden Arbeit wird versucht, diese Schwankungen als Folgen von Umweltveränderungen zu interpretieren. Dazu wurde ein numerisches Fließlinienmodell mit Temperaturverteilung entwickelt und auf zwei Abflußgletscher, Gunnestadbreen und Jenningsbreen, angewendet.

Die Modellierung zeigt, daß niedrigere Eistemperaturen, geringere Akkumulation und ein niedrigeres Meeresniveau zu einer eiszeitlichen Erhöhung der Gletscheroberfläche um 150—200 m führen. Nach einem Vergleich dieser Resultate mit einer paläogeographischen Rekonstruktion des Maximalstandes des Jenningsbreen durch Hirakawa und Mitarbeiter wird geschlossen, daß deren Moränen zu einer früheren, känozoischen Vereisung gehören müssen.

1. INTRODUCTION

The Sør Rondane is a 300 km long wedge shaped mountain range, 200 km south of the Princess Ragnhild Coast, Dronning Maud Land (20—30 E, 71—73 S), and belongs to a series of mountains surrounding the East Antarctic continent (fig. 1). It is one of the few areas around the Antarctic perimeter where nunataks and morainic deposits bear witness of former glacier stands. In the central part of the Sør Rondane there is

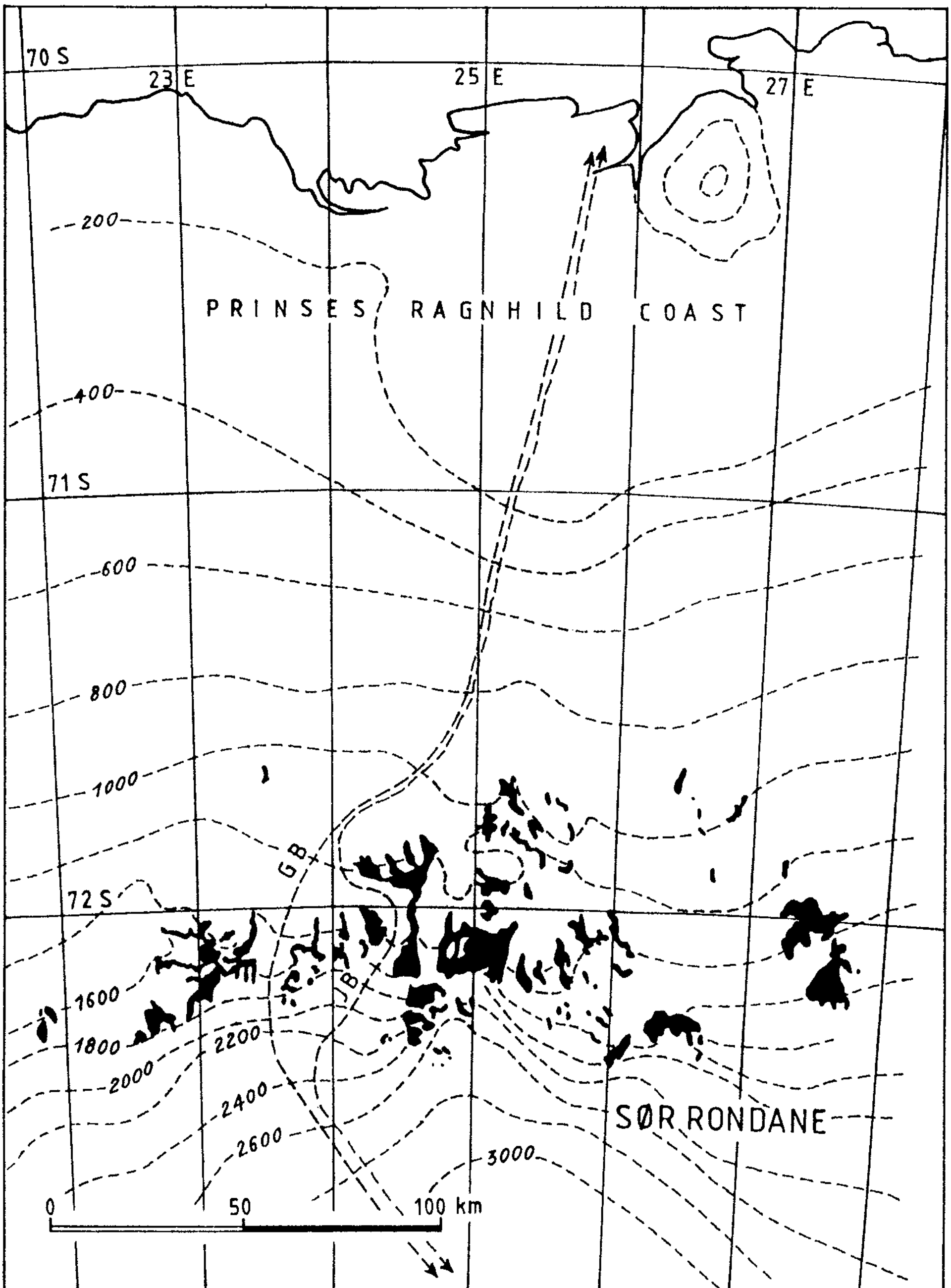

Fig. 1: Situation map of the Sør Rondane, showing the modeled flowlines (JB = Jenningsbreen, GB = Gunnestadbreen)

evidence that the ice once reached a level at least 300—350 m above the present ice sheet (Van Autenboer 1964; Hirakawa et al. 1988). It is the aim of the present work to attempt an interpretation of these geomorphological observations in terms of environmental change by means of numerical ice sheet modeling.

Several outlet glaciers cut through the mountain range, and reveal an ice covered fjord landscape. Changes in the surface level of these glaciers originate from a complex interaction with environmental conditions in a number of ways. A drop in the relative sea level stand, for instance, could increase the amount of grounded ice seaward of the mountain range, leading to a local thickening. A similar increase in length of the flowline could also result from a shift of the present ice divide further inland, so that ice from a larger catchment area has to be discharged. Changes in deformation characteristics of the ice due to temperature fluctuations in the basal shear layers may be another important factor to be considered, as colder ice deforms much less easily and consequently, builds up. On the other hand, reduced accumulation rates during colder periods, when less precipitable moisture is advected inland, may have resulted in lower glacier stands. In view of this qualitative description and in an attempt to disentangle the relative role of the many, in part counteracting processes, we developed a numerical ice sheet model for the glacier system in the Sør Rondane. This model basically simulates the glacier geometry in response to changing environmental conditions. Sensitivity experiments on the glacial-interglacial contrast may then provide physically-based arguments in order to better understand these observed glacier fluctuations.

In this paper, an overview will be presented of these sensitivity experiments. In particular, the response of two outlet glaciers of the Sør Rondane, namely Gunnestadbreen and Jenningsbreen to "glacial" shifts in mass balance, temperature, sea level stand and ice divide position has been investigated. These results are then compared with glacial-geological reconstructions of former glacier stands in the Sør Rondane (Hirakawa et al. 1988).

2. MODEL FORMULATION

The numerical gridpoint model used in this study computes vertically-integrated flow and involves basic modeling techniques similar to the ones described in Oerlemans and Van der Veen (1984). Given the basal topography, accumulation rate, surface temperature and sea-level, it calculates the time-dependent ice thickness distribution along a flowline. Changes at the seaward edge are then taken into account by coupling the ice sheet with an ice shelf and treating the grounding line (where the ice sheet starts to float) in a dynamic fashion. A fixed grid in space is employed, with 120 grid points along the x-axis (parallel to the geoid) and a grid spacing of 5 km. The z-axis points positive upwards (z = 0 at sea-level). The continuity equation, defining conservation of ice mass, then relates the local rate of change of ice thickness with the flow field, where ice density (p) is kept constant [917 kg m^{-3}]:

$$\frac{\partial H}{\partial t} = -\frac{\partial}{\partial x}(UH) - \frac{UH}{b}\frac{\partial b}{\partial x} + M \tag{1}$$

In this equation, H is ice thickness [m], t time [a], U the vertical mean horizontal velocity [m a^{-1}], M the mass balance [m a^{-1}] defined as the result of local accumulation and runoff (if present), and b(x) the width of the glacier along the flowline. In order to calculate the flow, we assumed a situation of plane strain in the x-z plane, so that all strain rate components involving y are zero. In the grounded ice sheet, where longitudinal deviatoric stresses (defined as the full stress minus the hydrostatic component) can be neglected, the basic stress equilibrium can be integrated along z to yield an expression for the basal stress τ_b:

$$\tau_b = F\tau_d = -\rho g F H \frac{\partial(H+h)}{\partial x} \tag{2}$$

Here, g is the acceleration of gravity [9.81 m s^{-2}], τ_d the driving stress and h the bedrock elevation [m a.s.l.]. In the mountain range, where the ice sheet transforms into a valley glacier, a shape factor F is introduced to account for the effect of side drag on the valley walls (Paterson 1981). Assuming a parabolic cross section, this shape factor is found to vary between 0.6 and 1.0 (the difference with 1 gives the part of the driving stress opposed by side drag). Substituting eq. (2) in a "Glen-type" flow law, with exponent $n=3$ and performing one more integration along the vertical gives the depth-averaged velocity U:

$$U = AH\tau_b^3 = A(\rho g F)^3 H^4 \left|\frac{\partial(H+h)}{\partial x}\right|^3 \tag{3}$$

where A is the temperature-dependent flow law parameter. So, the vertically-integrated velocity in the grounded ice sheet is a local quantity and depends non-linearly on ice thickness and surface slope. In this approach, no distinction has been made between ice flow due to internal deformation and basal sliding. However, with respect to the depth-averaged velocity there is probably not so much difference, since the bulk of the movement occurs in the lower layers anyway.

In the freely floating ice shelf approximation, on the other hand, basal shear stresses are very small and the driving stress is now balanced by gradients in normal stress deviators (that represent ice shelf "stretching"). From the condition that the net normal force on the ice shelf should equal hydrostatic pressure exerted by the sea water at the ice shelf front, velocities can be integrated along the flowline to give (see e.g. Oerlemans and Van der Veen 1984, p. 62):

$$\dot{\varepsilon}_{xx} = \frac{\partial U}{\partial x} = A\overline{\tau'^3_{xx}} = A\left[\frac{1}{4}\rho g\left[1 - \frac{\rho}{\rho_w}\right]H\right]^3 \tag{4}$$

where ρ_w is the density of sea water [1028 kg m^{-3}] and $\overline{\tau'_{xx}}$ the vertical mean longitudinal stress deviator. This formulation requires the ice velocity to be known at the grounding line. At this location, where neither the ice sheet nor the ice shelf approximation holds, and all stress components can be shown to be important, an expression for the vertically integrated velocity U can be derived from the flow law and the complete stress equilibrium. Following an analysis by Van der Veen (1985), it reads:

$$U = AH\left[\frac{5}{2}(\tau'^2_{xx} + D^2) + \frac{5}{3}(\tau'^2_{xx} + 3D^2)\tau_d + \frac{15}{4}D\tau_d^2 + \tau_d^3\right] \tag{5}$$

where D is the longitudinal stress gradient ($=2\partial/\partial x\,(H\tau'_{xx})$) and τ'_{xx} follows from ice-shelf dynamics as given in eq. (4). In the present approach, we opted for a stress transition zone of 10 km landinward, in which τ'_{xx} becomes zero. This is a value suggested by

Herterich (1987) and did not appear not to be crucial to the model outcome. Boundary conditions for the continuity calculations are zero ice thickness gradients at both the ice sheet and ice shelf ends. Ice-shelf calving is not considered, so the ice shelf extends all the way to the model grid edge. This has no influence on the position of the grounding line.

The model also accounts for the temperature dependence of the flow properties of ice. This is important since a 10° C temperature shift in the basal deformation layers implies an order of magnitude change in strain rates. A more rigorous approach here would be to calculate the fully coupled temperature and velocity fields in a two-dimensional vertical plane, as was done in Huybrechts and Oerlemans (1988), but this option was rejected because it would put too heavy demands on computational resources, in particular for the fine grid presently in use. Instead, thermodynamics were taken care of by a more crude method, that nevertheless accounts for the complete physics in an essentially correct way. As described in Oerlemans and Van der Veen (1984, pp. 88—89), the ice temperature has been expanded in powers of h′, the height above bedrock, and three terms have been retained:

$$\Theta(x,z,t) = \Theta_0(x,t) + h'\Theta_1(x,t) + h'^2\Theta_2(x,t) \tag{6}$$

The spectral coefficients Θ_0, Θ_1, Θ_2 of this second-order polynomial then follow from the lower and upper (the mean annual air-temperature) boundary conditions, and from the vertically integrated heat equation, where the two-dimensional ice velocity components in the advective terms are specified by parabolic functions. The lower boundary condition incorporates the effects of geothermal heating and dissipation and is expressed as a temperature gradient:

$$\left(\frac{\partial T}{\partial z}\right)_b = -\frac{\psi}{k} - \frac{U\tau_b}{k} = -\frac{1}{k}[\psi + \psi_c] \tag{7}$$

with ψ the geothermal heat flux [42 mWm^{-2}], $\psi_d = U\tau_b$ the contribution due to dissipation and k the thermal conductivity of ice [66 MJ $m^{-1}K^{-1}a^{-1}$]. However, an additional remark is in order here, as high velocities and surface slopes in the marginal ice sheet region appeared to lead to excessively high values of the deformational heat flux and consequently, instabilities in the temperature calculation. This complication has been avoided by setting an upper limit to the basal temperature gradient and spreading the left-over dissipative heat along the vertical.

Since most of the velocity shear is found in the lower layers of the ice sheet, it is natural to express the flow parameter in function of basal temperature, that is done by means of an Arrhenius equation (e.g. Paterson 1981):

$$A(T^*) = a \exp\left[\frac{-Q}{RT^*}\right] \tag{8}$$

where T* is absolute ice temperature corrected for the dependence of the melting point on pressure ($T^* = T + 8.7\ 10^{-4}H$, where T is expressed in K), Q the activation energy for creep (60 kJ mol^{-1} for $T^* < 263.15$ K and 139 kJ mol^{-1} above 263.15 K), R the universal gas constant (8.314 $Jmol^{-1}K^{-1}$) and $a = 1.71\ 10^{-6} Pa^{-3}$ $year^{-1}$ for $T^* < 263.15$ and $a = 8.21\ 10^9\ Pa^{-3}$ $year^{-1}$ above 263.15 K. In the ice shelf, a linear temperature profile was put forward. With the bottom temperature equal to the freezing point of sea water, the corresponding flow parameter is then computed with the mean shelf temperature.

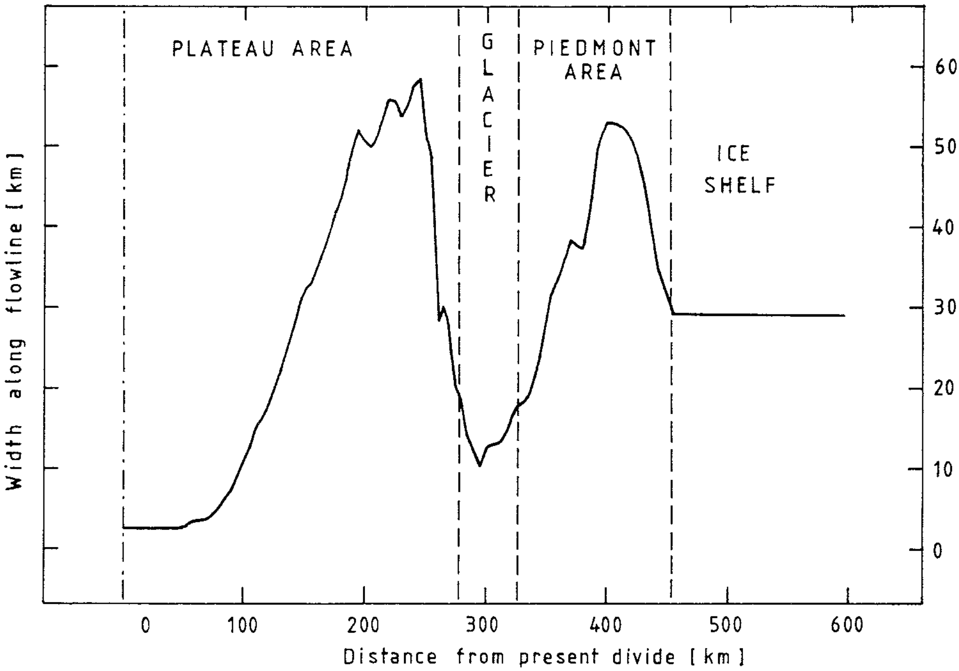

Fig. 2: Width of Gunnestadbreen along the flowline. The width distribution of Jenningsbreen resembles this one to a large extent

3. DATA

Basic input data for the model are surface and bed topography. Two flow lines perpendicular to the contour lines were drawn along Gunnestadbreen and Jenningsbreen, two drainage glaciers of the central part of the Sør Rondane (fig. 1). Bed topography has only been measured at a limited number of locations in the mountain range (Van Autenboer and Decleir 1974; De Vos and Decleir 1988) and at the grounding line (F. Nishio, personal communication). However, since surface elevations are read from a topographical map, remaining ice thicknesses 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 (e.g. Paterson 1981, p. 86). These yield stresses are then obtained by inter/extrapolating values calculated at measured sites along the whole glacier length. On the plateau south of the ice fall, where surface gradients become very small, the bedrock slope was taken linearly downwards to the center of the ice sheet, as suggested by the Drewry maps (1983). The resulting glacier geometries are shown in fig. 3. They both show a stepwise surface profile with steep slopes in the mountain range. The bedrock topography suggests an overdeepened fjord landscape: a southern threshold where the ice sheet transforms into a valley glacier and where ice thicknesses are relatively small, ending in a subglacial trough system partly below sea level north of the mountain range. This general picture agrees well with radio echo sounding results (F. Nishio, personal communication). The width distribution along the Gunnestadbreen and Jenningsbreen flowline is displayed in fig. 2.

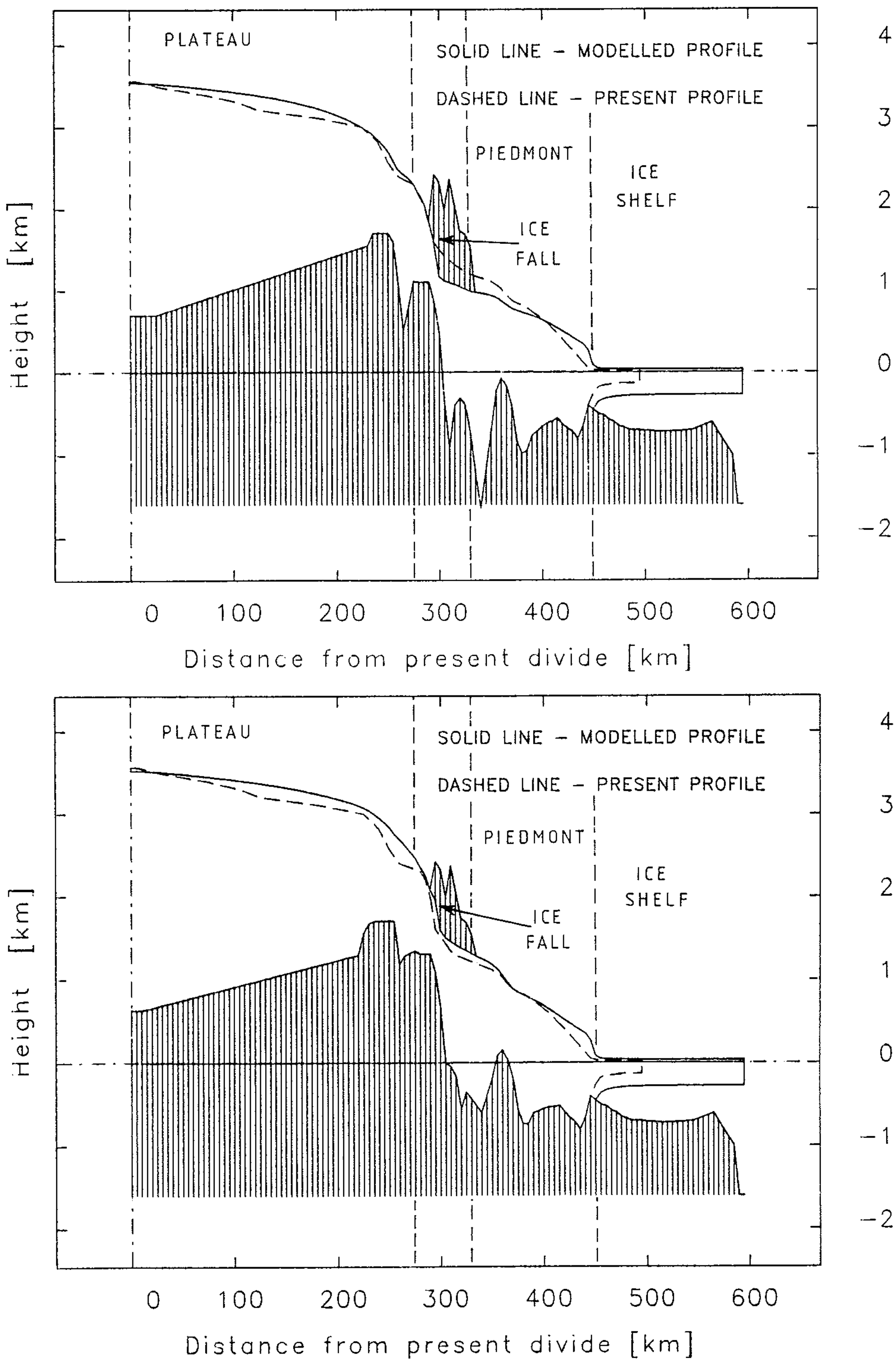

Fig. 3: Glacier geometries of Gunnestadbreen (upper panel) and Jenningsbreen (lower panel), with model output for present conditions (solid lines)

Relations for surface temperature and accumulation rate were obtained as simple functions of elevation and temperature, respectively. Data were selected from a compilation for the entire Antarctic ice sheet made at SPRI, (supplied to us by D. Drewry) and include the 24° E central Sør Rondane traverse. These parameterizations read:

$$T_{surface} = 261.4\,\mathrm{K} - 0.012(H + h) \tag{9}$$

$$M = 6.15\;10^{-8}\left[\frac{T_{surface}}{16.85}\right] \tag{10}$$

which are functional relations to be expected. For every 10 K lowering of surface temperature, which is a typical glacial-interglacial shift, the accumulation rate is then roughly halved. Although the present glacier surface in the mountain range is to some extent characterized by bare ice fields, indicating that the accumulation regime is locally disturbed by sublimation, wind scouring (due to katabatic winds) and perhaps some surface melting as well, ablation has not been considered. In this respect, it remains highly uncertain whether these conditions also prevailed during colder climates and/or higher glacier stands. Also, prescribing zero accumulation or even ablation rates of 10 cm a^{-1} (Van Autenboer 1964; Nishio et al. 1986) in the mountain range, turned out hardly to change the qualitative behaviour of the model (in particular the *relative* response).

4. REFERENCE RUN

If one wants to investigate changes in glacier geometry, the first thing to do is defining a reference run. Starting from the observed ice sheet and imposing different environmental conditions is a less meaningful way to proceed, because the input data are most likely not in full internal equilibrium with the model physics. As a consequence, it becomes hard afterwards to distinguish between the "natural" model evolution and the "real" ice sheet response. This may be due to shortcomings in the description of ice mechanics or insufficient data coverage, or it may indicate that the present ice sheet is just not in steady state. It is probably a combination of all these factors. Since it is not known how far the present glaciers are out of steady state, the pre-exponential constants a in eq. (8) have been chosen such that the modeled glacier geometries closely resembled the presently observed states. The resulting flow law parameters then appeared to be a factor 7 lower (taken for both glaciers) than the ones given in Paterson (1981). This represents an ambiguity that cannot be avoided, but the relative response (our prime interest, in fact) appeared not to depend crucially on this choice.

The resulting ice sheets, after the model has been allowed to relax to a stationary state for 200,000 years (the interglacial reference runs) are displayed in fig. 3 (solid lines). In general, the agreement with the field data is quite satisfactory, and the model is certainly capable of reproducing the main characteristics of these glaciers, such as the stepwise surface profile when the glacier enters the mountain range. An independent check on the model results may then be provided by the velocity and temperature profiles. Fig. 4 shows the calculated depth-averaged horizontal velocities and the basal temperature distribution in the reference run for Gunnestadbreen. Basal temperatures are shown to rise from cold values between −25° C and −10° C below the Polar Plateau to near pressure melting beyond the ice fall. These values suggest basal sliding to make up for a significant part of the vertically-integrated ice mass flux, once the mountain range is surpassed. In view of a similar geometry for Jenningsbreen, the corresponding basal temperature profile essentially shows the same picture.

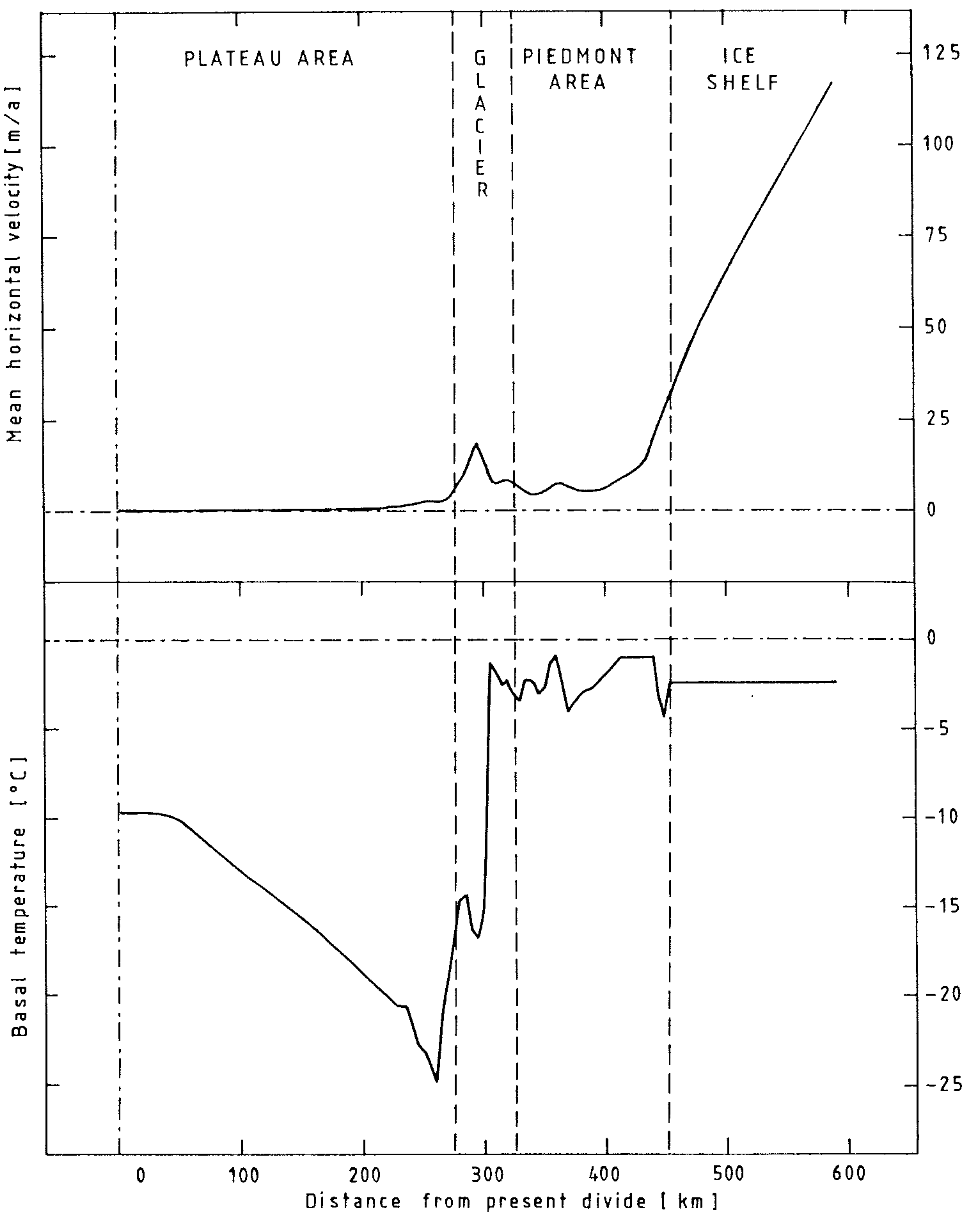

Fig. 4: Depth-averaged horizontal velocity (upper panel) and basal temperature (lower panel) in the interglacial reference run for Gunnestadbreen

The surface velocities in the central part of the Sør Rondane are low as compared to the larger outlet glaciers east and west of the range (Van Autenboer and Decleir 1974, 1978). Observed velocities for Gunnestadbreen have been measured to be in the range of 10—30 m a^{-1}, which agrees well with the modeled values. Measured velocities of Jenningsbreen, on the other hand, are found to be extremely low (less than 1 m a^{-1}). In view of the similar geometric characteristics of this glacier, the modeled velocities (up to ± 30 m a^{-1}) are obviously at odds with the observed ones. Van Autenboer and Decleir (1978) suggested that in view of a general reduction in glacierization, Jenningsbreen is presently cut off from the main ice supply and is thereby not actively fed by the plateau ice sheet. A small ice cap south of Walnumfjellet, still in contact with the main ice sheet, was thought to feed Jenningsbreen. In order to investigate this point somewhat further, an additional experiment was carried out, in which Jenningsbreen was modeled as a local glacier, starting from a local ice divide near to the present glacier head. Modeled surface velocities for a steady state situation are now much lower (typically of the order of 3 m a^{-1}) and agree better with the observed values. From this it must be concluded that Jenningsbreen is at present indeed a local glacier, or alternatively, has to be grossly out of equilibrium. However, if the glacier surface elevation would increase (glacial stage) Jenningsbreen would not be a local glacier, but would indeed become a real outlet glacier such as Gunnestadbreen.

5. SENSITIVITY STUDY

Let's turn to the central question now of what factors may be decisive to explain a surface elevation shift of some 300—350 m in the central Sør Rondane mountain area, as suggested by glacial-geological evidence (Hirakawa et al. 1988). As mentioned in the introduction, we forced the respective reference runs by shifts in environmental conditions, thought to be representative for "glacial" conditions (e.g. Jouzel et al. 1987). The following scenario's were tested: (i) a general lowering of accumulation rate with 50 %, (ii) a decrease in temperature at sea-level with 11° C, (iii) a lowering of sea-level with 150 m, (iv) a shift of the ice divide 200 km towards the interior, and several combinations of these elements. The model then ran some 200,000 years forward in time, when a clear stationary state was established.

Results of these sensitivity experiments for Jenningsbreen are shown in fig. 5. As can be judged from this figure, by far the most decisive single environmental parameter is a temperature drop of 11 K, that stiffens the ice and consequently, results in a buildup (fig. 5, curve 2). This effect is further enhanced by lowering sea-level, that results in a small additional seaward migration of the grounding line by some 20 km (fig. 5, curve 5). However, colder climates in the Antarctic are believed to be accompanied by reduced accumulation rates, mainly because colder air can carry less precipitable moisture. As became clear in the Vostok deep ice-core results (e.g. Lorius et al. 1985), this dependence is particularly apparent over the East Antarctic Plateau. Although such a relation is less obvious in coastal areas, where accumulation processes are not quite the same as on the plateau, curves 4 and 7 (fig. 5) show the model outcome when accumulation rates have been reduced by 50 %. This represents a counteracting effect, although of lesser amplitude than the temperature response, and the shift in surface elevation now amounts to 150—200 m.

A possible explanation for these lower elevation shifts, as compared to Hirakawa's values, may be due to the fact that the grounding line does not seem to move much, even in case sea level is lowered by 150 m. This may be due to shortcomings in the model or indicate that the present flowline approach is not accurate enough in view of the rugged subglacial bed topography. However, it seems equally likely that the imposed late-Quaternary glacial-interglacial shift is just not of sufficient amplitude to produce important grounding line migration. To investigate this point somewhat further, the grounding line was prescribed to take its maximum possible position, i.e. at the edge of the continental shelf. Such a displacement by about 120 km of the grounded ice sheet may have resulted, for instance, from tectonic or isostatic uplift in former times. Additionally, we imposed a shift of the ice divide 200 km further inland, that may have followed changes in the flow pattern on the East Antarctic ice sheet dur

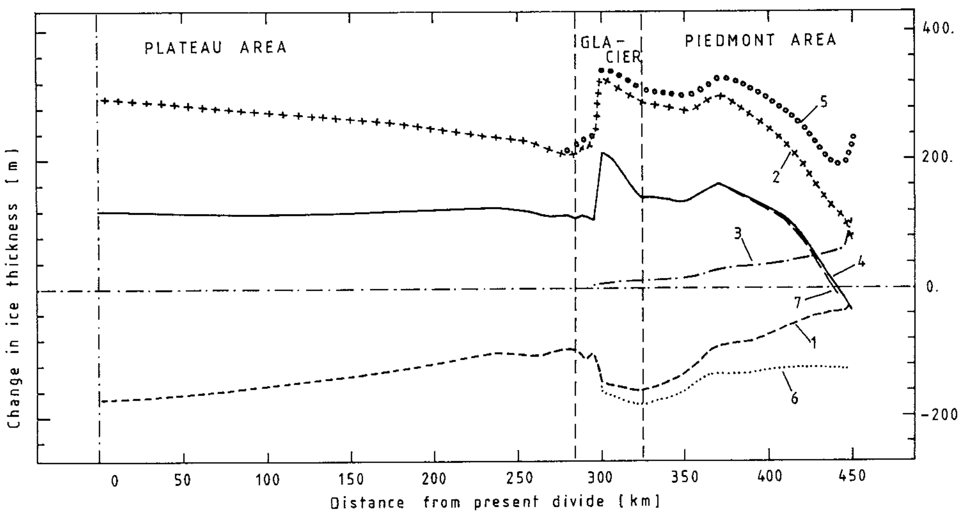

Fig. 5: Jenningsbreen flowline: absolute changes in ice thickness according to the different sensitivity experiments; (1) ΔM, (2) ΔT, (3) ΔHsl, (4) ΔT ΔM, (5) ΔT ΔHsl, (6) ΔM ΔHsl, (7) ΔT ΔM ΔHsl, with ΔM a mass balance lowering of 50 %, ΔT a decrease in temperature with 11 K and ΔHsl a sea-level drop of 150 m

ing large glaciations. Results of this admittedly rather speculative experiment are compared with our "standard glacial run" in fig. 6. It is interesting to note that a displacement of the ice divide influences the ice sheet topography only marginally, and that the effect fades out when the mountains are approached (fig. 6, curve 1). On the other hand, the position of the grounding line has a marked influence (fig. 6, curve 2). In particular, downstream the mountain range and the ice fall, important changes of 500 m and more do show up now, and not surprisingly, their magnitude increases with decreasing distance from the grounding line. Fig. 7 shows this extreme glacial situation covering the greater part of the Sør Rondane and leaving only the highest peaks of the mountain range as isolated nunataks. So, from this experiment it must be concluded that a larger elevation shift, more in accordance with the field data, can only be explained, when an additional forward grounding-line migration is assumed.

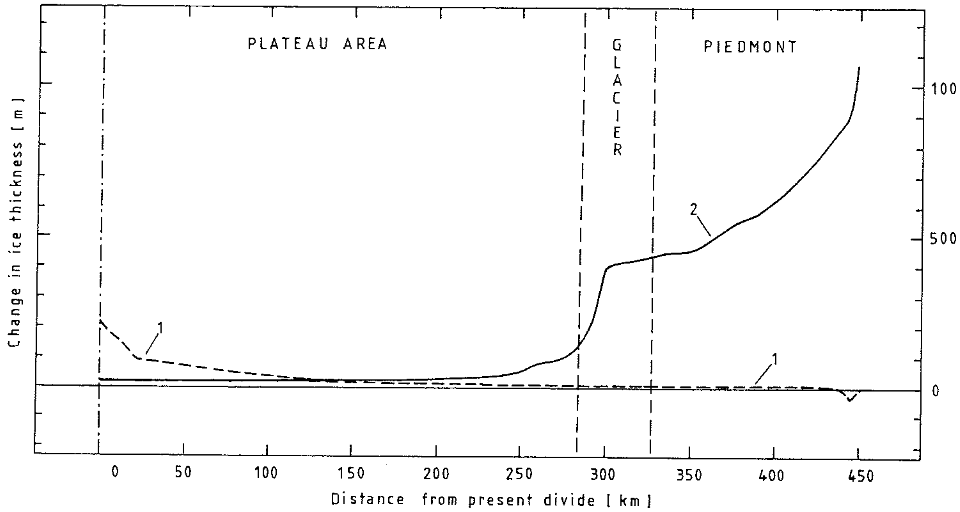

Fig. 6: Jenningsbreen flowline: shift in surface elevation relative to the reference run (1): ice divide displaced 200 km inland (2) grounding line at the edge of the continental shelf

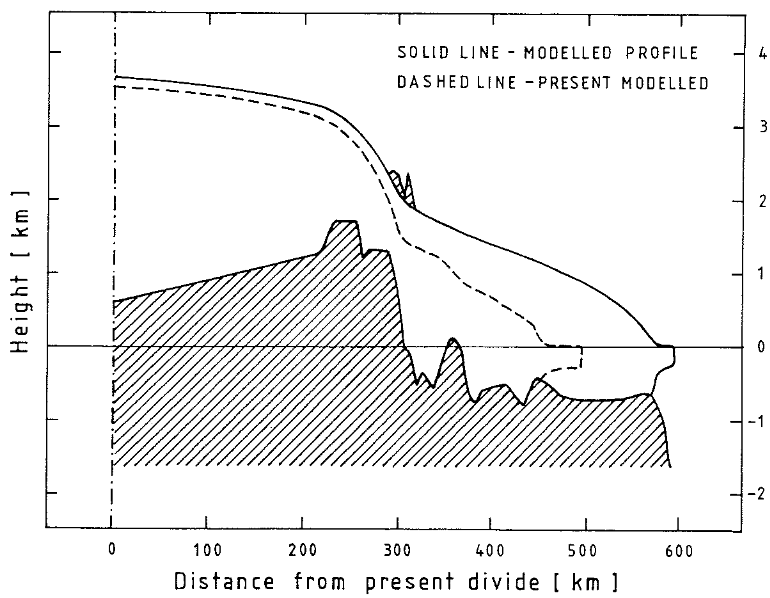

Fig. 7: The Jenningsbreen flowline in a "maximal glaciation" model run

6. DISCUSSION

On the basis of a large amount of geomorphological and glacio-geological evidence (ice smoothed surfaces, roches moutonnées, erratics, moraine cover, . . .) Hirakawa et al. (1988) present a palaeogeographical reconstruction of a "maximum" glacial stage in the central Sør Rondane. In fig. 8 both Hirakawa's profiles (present and glacial) and our model results for Jenningsbreen (in the "standard glacial run", fig. 5, curve 7 and in an earlier Cenozoic glacial stage whereas the grounding line reaches the edge of the continental shelf, fig. 6, curve 2) are shown together. Hirakawa's profiles show the conspicious advance of the ice fall near the head of Jenningsbreen associated with a reduced rise in topography on the lower part as compared to the relative important rise of the surface upstream the ice fall. In contrast with this reconstruction, our results of the earlier Cenozoic glaciation show the surface elevation rise to be more important below the ice fall than on the adjoining plateau. In fact, on a physical base it cannot be justified that the thickening below the ice fall would be less pronounced than a thickening upstream. High surface and bedrock gradients at the ice fall imply small ice thickness changes due to large surface velocities, while a migration of the grounding line towards the north results in an ice thickening downstream. A large grounding line migration (fig. 6, curve 2) results in important changes of 500 m and more downstream the mountain range and the ice fall, while the ice sheet at the plateau is influenced only marginally. This means that the Sør Rondane damps ice changes at the plateau. Another discrepancy with Hirakawa's results concerns the advancement of the ice fall. The fact that our model does not account for a larger displacement may be due to the large surface gradients.

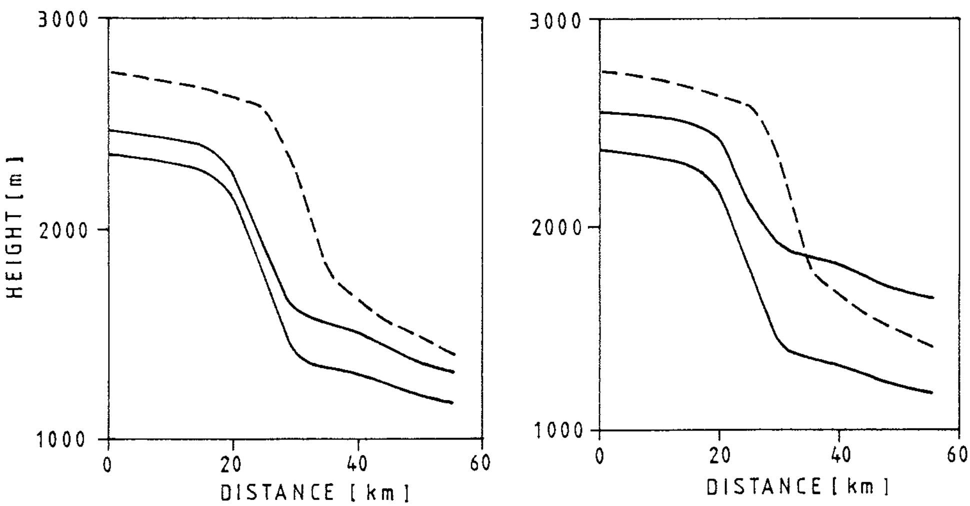

Fig. 8: A comparison between Hirakawa's palaeogeographical reconstruction (dashed line) with our model results (solid lines) at maximum glaciation (right panel) and in the standard glacial run (left panel)

7. CONCLUSIONS

In this study, an attempt was made to interpret observed glacier fluctuations in the central Sør Rondane mountains by numerical modeling of its glacier system. In spite of some of the approximations involved, the model was shown to be able to account for major characteristics of the present glacier geometry. A comparison with a palaeogeographical reconstruction of a maximum glacial stage in the Sør Rondane revealed that deposits were not from a late Quaternary glaciation but should be from a full grown ice sheet, i.e. an earlier Cenozoic glaciation.

As a final remark, however, it should be stressed that our results relate to steady state conditions and involve environmental scenario's of a rather schematic nature. This may be seen as a drawback, but since the interest was in an order of magnitude estimate, a more detailed approach would probably not be justified here. Also, we have no reasons to believe that this approach determines our model outcome in the sense that results could become entirely different.

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