GLACIATION OF THE CENTRAL PART OF THE SØR RONDANE, ANTARCTICA: GLACIOLOGICAL EVIDENCE

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Abstract: Ice thickness measurements, carried out by radio echo sounder and gravimeter in the central part of the Sør Rondane Mountains, reveal a subglacial topography of the outlet glaciers and its tributaries characterized by U-shaped valley profiles and overdeepened bedrock. Mass-flux measurements highlight the reduced flow of at least one glacier (Jenningsbreen). This glacier is in the process of being cut off from the main ice supply and may serve as an example for the deglaciation process. An interesting feature of this deglaciation is that, once decoupled from the main ice supply, this glacier is probably characterized by an increased lowering of the ice surface gradient due to the ablation which is characteristic for the upper part of the present outlet glacier. In the end this will result in a southward flow of which examples can be found elsewhere in the mountains. On the basis of the field evidence a numerical flow line model is presented to simulate the behaviour of the outlet glaciers of the central part of the mountains during the last glacial maximum. These experiments show an increase of 300–400 m in ice thickness under realistic assumptions for mass balance, temperature and sea level. Some of the higher glacial levels are then attributed to an environment characterized by a higher accumulation predating the last glacial maximum.

Key words: glaciation, deglaciation, glacial geology, glacier modelling, subglacial topography

Introduction

With respect to the glacial history of Antarctica, we witness at present a shift of interest from a (deep) sea bottom view towards a land based view. Indeed, much new evidence is now emerging from continental Antarctica, both from investigations of the continental shelf and from studies of the ice-free areas. These ice-free areas can be subdivided into low lying coastal oases and marginal mountain areas. While the former originated from a general recent (Holocene) ice sheet retreat, the latter is witness of a much longer glacial history. This history can be followed through study of the morphological features and sediments laid down by the waxing and waning of outlet glaciers traversing the mountain massifs. The paleogeography reconstructed from the glacio-geological evidence is physically constrained by the laws of glacier dynamics. In this respect it is interesting to note that recently a three-dimensional ice cap model has been put forward enabling for the first time a realistic simulation of the Antarctic ice sheet as a whole during the last glacial-interglacial cycle (Huybrechts, 1990a). On a larger scale, for the simulation of outlet glaciers, we still have to rely on numerical flow line models (Pattyn *et al.*, 1989).

In this paper we will focus on the Sør Rondane, a typical coastal margin mountain range in Dronning Maud Land and present a glaciological view on the glaciation and deglaciation of the range. Glacio-geological and geomorphological evidence of this area were first described by Van Autenboer (1964), Souchez (1966) and later mainly by Japanese workers e.g. Iwata (1987), Hirakawa *et al.* (1988), Hayashi and Miura (1989), Hirakawa and Moriwaki (1990). The glaciological observations presented here have been collected through participation in the Japanese Antarctic Research Expedition, JARE 28 (1986–87), JARE 31 (1989–90) and JARE 32 (1990–91).

Mass Balance and Temperature Regime of the Sør Rondane Area

The Sør Rondane Mountains are a 220 km long, east-west lying mountain range with the highest elevation of 3000 m, situated at a distance of 200 km from the coast and damming the main ice flow coming from the polar plateau (Fig. 1). The damming effect can clearly be seen by (i) the stepwise topography of the glaciers ("ice fall") at the southern rim where they cut through the Sør Rondane (ii) the divergent ice flow pattern imposed by the mountains and (iii) the considerable reduced ice mass transport through the mountains. The latter has been established by Van Autenboer and Decleir (1974, 1978) by measuring the ice discharge through each of the outlet glaciers over the entire northern margin of the mountains. The total ice discharge, assuming a density of 917 kg/m^3, corrected for a probable underestimation of 10% (Decleir *et al.*, 1989) amounts to 1.76 km^3/y or a mean mass flux (defined here as the discharge of ice per km) of 0.01 km^2/y, calculated over the 180 km long northern boundary of the mountains. This mass flux is especially low if we compare this value with the mean mass flux for the periphery of Antarctica as a whole which we estimate at 0.15 km^2/y (assuming a total net accumulation of 2000 km^3/y of ice on a circular continent with mean radius of 2105 km). As noted by Van Autenboer (1964) and Nishio *et al.* (1984), the reduced ice flow through the mountains is probably the reason for the sheltered and crevasse-free inland ice slopes, northwards from the mountains until the coast and ending in the relatively narrow Roi Baudouin Ice Shelf.

The mass flux of the ice shelf at the coast is much larger than the 0.01 km^2/y flux through the mountains. By comparing the coastal configurations on a 1937 and 1960 map, Van Autenboer (1964) estimated the ice shelf velocities at about 300 m/y. These values agree with the results (veloci-

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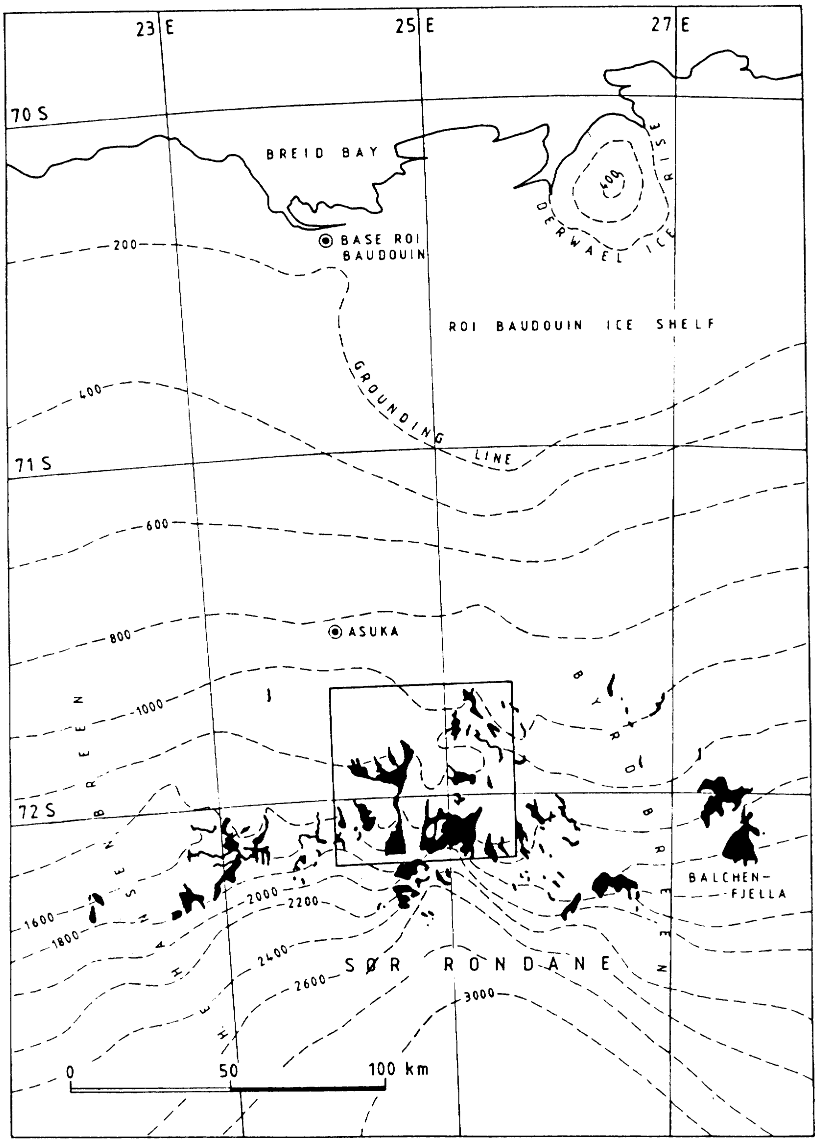


Fig. 1. Topographic map showing the Sør Rondane Mountain Range, the inland ice slope and Breid Bay.

ties ranging between 250 and 300 m/y) from the repeated survey (1965–1967) of a geodetic network spanning the area between Base Roi Baudouin and Derwael Ice Rise (Derwael, personal communication). Assuming a mean frontal ice thickness of 200 m a mean mass flux of 0.06 km^2/y is inferred. This leaves us with a horizontal mass divergence of $0.06 - 0.01 = 0.05$ km^2/y, which -in case of equilibrium- must be compensated by a mean net mass accumulation of 0.30 m/y of ice over the 170 km inland ice slope and ice shelf.

The areal distribution of the mass balance is however difficult to assess. A net accumulation of 0.43–0.50 m/y of ice was measured at Base Roi Baudouin (situated on the ice shelf some 15 km from the coast) by drilling (Tongiorgi *et al.*, 1961) and from stake measurements (De Breuck, 1961). A few kilometers further south, on the first slopes beyond the grounding line (i.e. the transition zone between the grounded ice sheet and the floating ice shelf), a higher accumulation of 0.67 m/y was measured as should be expected in this coastal area where the precipitation is mainly caused by cyclonic activity. Further inland on the regular rising inland slope and away from the moist bringing source, both Belgian and Japanese stake measurements indicate a rapid decrease of the accumulation to reach a few centimeters at the foot of the Sør Rondane (altitude of 1000 m).

For the model calculations in this paper we will assume - below 1000 m- a linear decrease of both mass balance M and temperature T with altitude z.

$$M = 0.50 - 0.00025\,z$$

for $z < 1000$ m

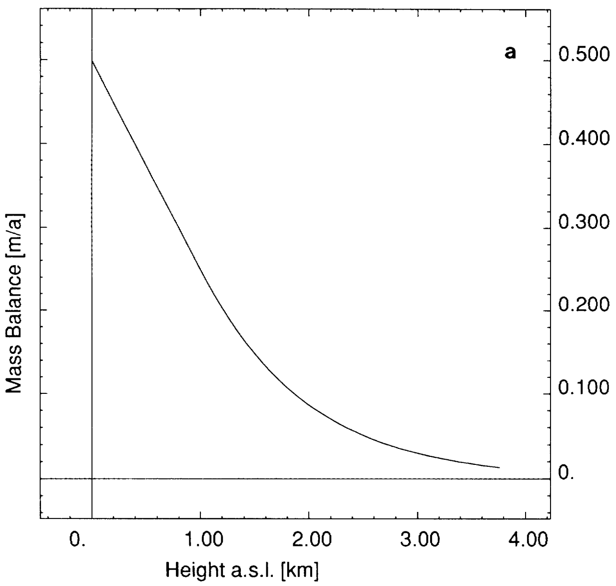
$$T = 257.4 - 0.0051\,z$$

with z in m, M in m/y of ice, and T in K. The lapse rate for temperature has been taken from Fortuin and Oerlemans (1990) and the sea level temperature adapted to the mean temperature measured at Base Roi Baudouin. This gives for the coastal area north of the Sør Rondane an average mass balance of 0.38 m/y which is to be compared with the 0.30 m/y inferred from the horizontal mass divergence. From this we infer the relatively independent existence of the coastal ice sheet and ice shelf from the continental outflow, sheltered as it is by the damming mountains. It should be added immediately that Nishio *et al.* (1984) estimated, on the basis of a more recent hydrographic survey, much lower ice shelf velocities and a discharge at the coast which exceeds the one through the mountains by a factor of two only.

Further inland, beyond the mountains, the linear decrease in temperature with altitude is much greater (Fortuin and Oerlemans, 1990), while for the mass balance an exponential decrease, typical for the cold polar deserts, was assumed

$$M = 0.687 \exp(-0.00106\, z) \qquad \text{for } z \geq 1000 \text{ m}$$

$$T = 266.4 - 0.0143\, z$$



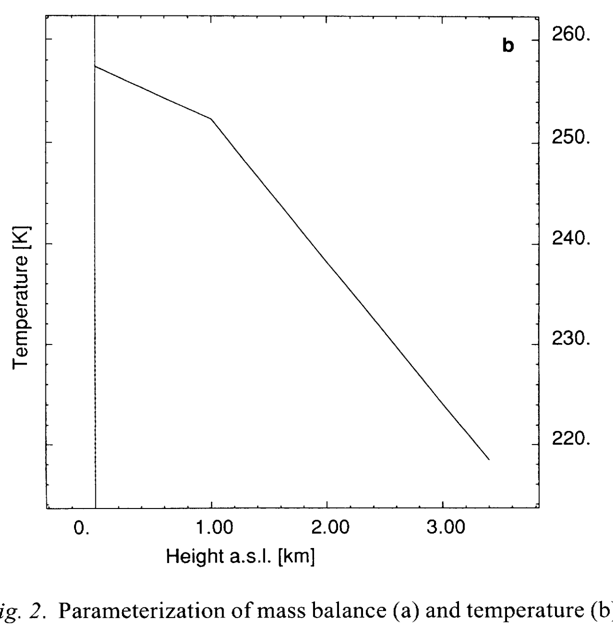


Fig. 2. Parameterization of mass balance (a) and temperature (b) in function of height above sea level.

The adapted temperature and mass balance profiles are shown in Fig. 2. According to this mass balance parameterization, the whole region belongs to the accumulation area ($M > 0$). However, within the mountains, especially near the lee side of the wind-protecting ridges, extensive blue ice zones exist. Based on satellite remote sensing data, the blue ice coverage of the upper part of Jenningsbreen and Gjelbreen was estimated at 45–50% of the glacier surface. Non-systematic stake measurements, in the upper part of the glacier valley, indicated in 80% of the cases a negative mass balance (ablation values ranging between 0 and 0.30 m/y).

Glaciation of the Sør Rondane

Two relatively large glaciers, H.E. Hansenbreen and Byrdbreen, near the western and eastern margins of the Sør Rondane respectively, discharge about 73% of the total ice transport through the mountains, underscoring the damming effect of the central part of the mountains. Together with the remaining outlet glaciers they divide the Sør Rondane into a number of massifs separated by deep U-shaped valleys, creating a complex landscape characterized in the first place by selective linear glacial erosion. The intervening massifs are indeed marked by cold based small valley glaciers and local ice caps probably exerting little erosion. A number of other features, however, testify here of a former higher and much more active glacier stand of which the south-north tending dry valleys, parallel to the present active outlet glaciers, are the most prominent.

The extreme eastern part of the mountains, with peaks never exceeding 2000 m, was completely overridden by the ice as shown by the presence of erratics, the flat topography of ridges and valleys, glacial striae etc. (Van Autenboer 1964; Hayashi and Miura, 1989). Moreover, the degree of weathering, especially in the North Balchenfjella area with the extensive occurrence of glacial polished and smooth surfaces, led Hayashi and Miura (1989) to conclude to a relatively recent retreat of the ice cover. Long lasting ice-free conditions on the other hand are inferred in the western part of the mountains where, under influence of the subaerial weathering, the retreat of the glaciated steep walls resulted in rectilinear debris-covered slopes (Richter slopes) where the angle of the rectilinear sections are governed by the repose angle of the debris (Iwata, 1987). In the central part of the mountains, the full scale of alpine morphology (arêtes, pyramidal peaks, cirques) witness of a former wetter and warmer glacial conditions probably preceded the present-day dry- and cold-based conditions. On the basis of morphological evidence as well as the presence of tills, erratics, striae, roches moutonnées etc., Hirakawa *et al.* (1988) were able to reconstruct the elevation of the maximum stage of glaciation and conclude in general to a 400 m higher glacier stand, locally reaching 1000 m, at the foot of the ice fall.

We will now investigate in more detail the glaciological conditions pertaining to this maximum glacial extent by means of modelling. To do this we will first describe the present characteristics of the glaciers and of the subglacial relief.

Subglacial Relief of the Central Part of the Sør Rondane

In the central part of the Sør Rondane (Fig. 3) Gjelbreen

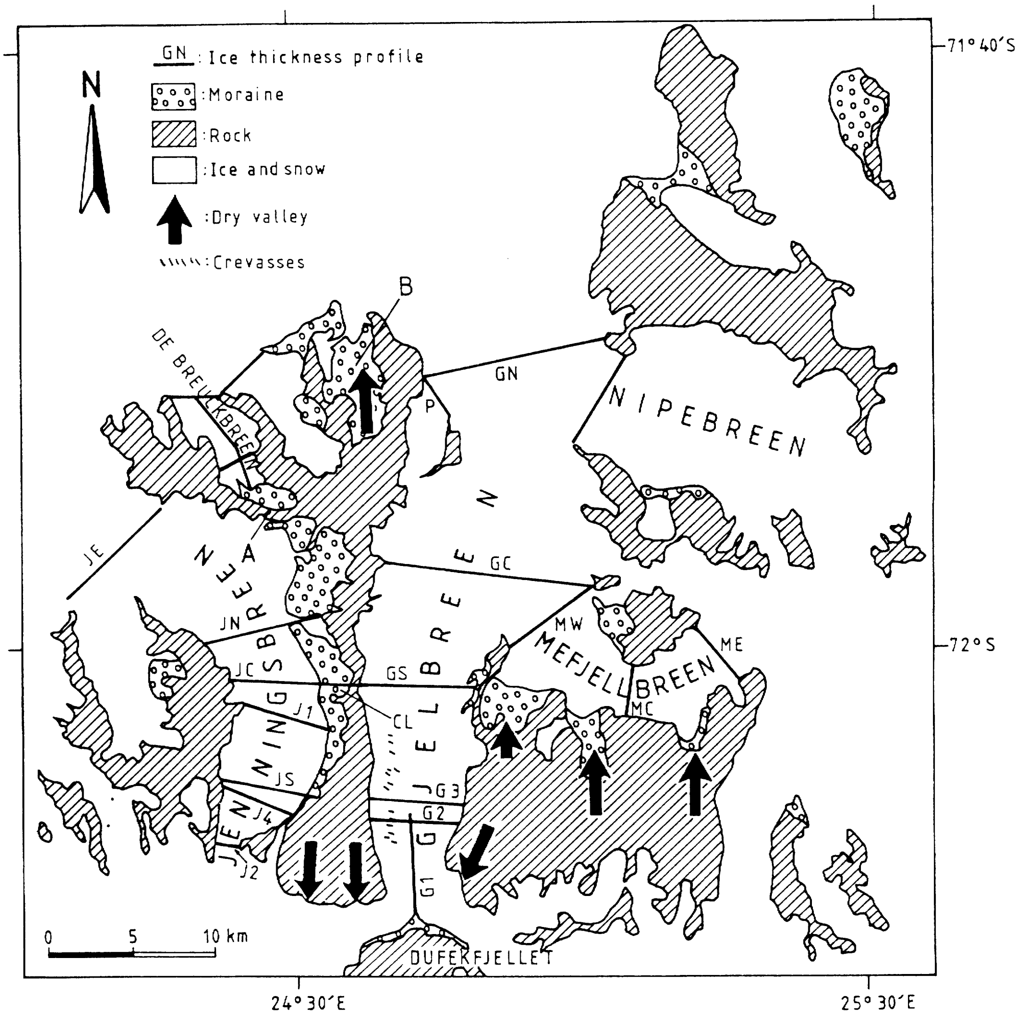


Fig. 3. Detailed map of the central part of the Sør Rondane (inset of Fig. 1) displaying the main outlet glaciers and measured ice thickness profiles.

and Jenningsbreen have cut a 40 km long U-shaped valley through the mountains, descending from an altitude of 2000 m to reach 1100 m at the northern edge of the mountains. At the southern entrance the ice is funnelled through a fairly narrow gorge (3 km wide), spilling over and cascading down the trough head to flow northwards in a widening valley (10–12 km wide in the exit area).

Glacier cross section

A selection of cross profiles, based on ice thickness measurements carried out either by radio echo sounder or by gravimeter during JARE 28, 31 and 32 are presented in Fig. 4 (see Fig. 3 for situation). Both glaciers display the characteristic U-shape of the ice filled valley, often with a slope break separating the region with present subaerial weathering from the area with glacial erosion. Such breaks are also observed in the dry valleys adjacent to Jenningsbreen and Gjelbreen revealing a former higher glacier stand (e.g. Hirakawa and Moriwaki, 1990).

For Jenningsbreen, which lacks tributaries, one expects a gradual evolution of the cross profile from head to exit. In order to investigate this we calculated Graf's form ratio

$$F_R = \frac{D}{2W}$$

where D is the depth and W the half width of the valley, as well as the coefficient a of the best fitting parabola $y = ax^2$ (with x the cross distance reckoned from the midpoint of the valley and y the height above the valley floor). It has indeed been shown (Hirano and Aniya, 1988) that such a parabola gives a very good approximation to a cross profile resulting from glacial erosion related to continental ice sheets.

Our results (Table 1) indicate a mean value for the form ratio of $F_R = 0.22$ (excluding the value for cross profile J4), which is close to the value found for glaciers in Antarctica by Hirano and Aniya (1988). More surprising is the constancy of these F_R-values for Jenningsbreen, although the valley depths vary between 400 and 1700 m (Fig. 4a). This suggests that the deepening of the glacial trough is proportional to the widening of the valley in such a way that the form of the cross profile is preserved. Adopting a parabolic cross section one expects thus each cross profile to be described by

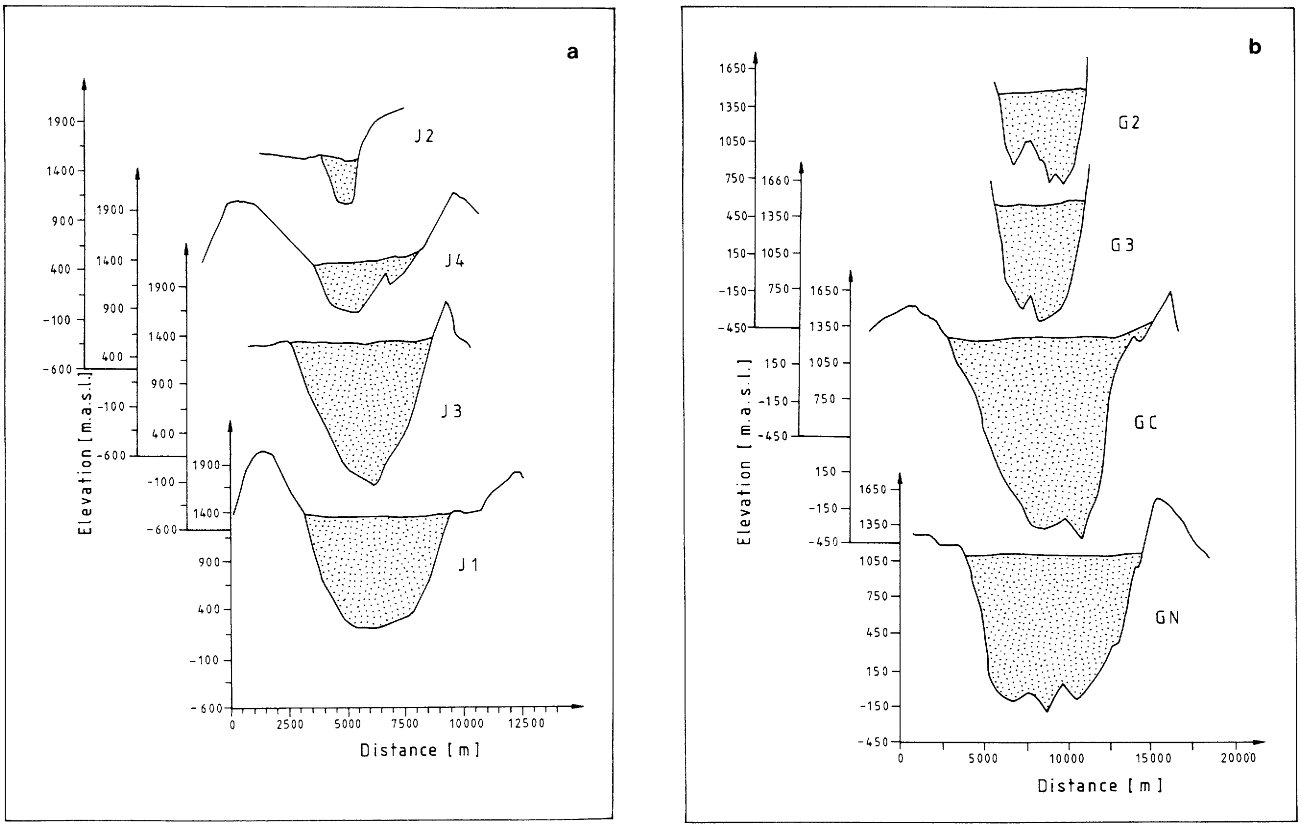


Fig. 4. Cross-profiles (from south to north) of Jenningsbreen (a) and Gjelbreen (b).

$$y = ax^2 \quad \text{where} \quad a = \frac{2F_R}{W}$$

so that the slope is given by

$$\frac{dy}{dx} = 4F_R \frac{x}{W}$$

This means that at certain distances from mid-valley ($x = W$, $x = W/2$, $x = W/4$, ...) the same slope value is kept along the glacier. Only profile J4, at the foot of the ice fall, shows an anomalous low form ratio but this can be explained by the fact that the actual channel occupies here only the eastern part of the cross section.

The analysis is less straightforward for Gjelbreen which is to be considered as a composite glacier (Fig. 4b and Table 2). Profile G2 can best be visualized as the juxtaposition of two parabolic profiles which, taken individually, would result in a higher value for the form ratio than the one given in the table. Further down the valley (GS) the subglacial ridge disappears and the form ratio increases to $F_R = 0.26$, to decrease again for profile GC ($F_R = 0.13$, north of the junction with Mefjellbreen) and profile GN ($F_R = 0.12$, north of the junction with Nipebreen).

In summary then we find that the glacial erosion results in standard parabolic valley forms with a form ratio $F_R = 0.22$, which is preserved along the glacier valley. Broader U-profiles, with much lower form ratio's, are due to the confluence or junction with adjacent glaciers or tributaries.

Table 1.

	$F_R = D/2W$	a (Least squares)	$a = 2F_R / W$
J2	0.24	$(8.8 \pm 1.4) \times 10^{-4}$	4.5×10^{-4}
J4	0.15	$(2.1 \pm 0.09) \times 10^{-4}$	4.1×10^{-4}
J1	0.18	$(1.2 \pm 0.08) \times 10^{-4}$	1.7×10^{-4}
JL	0.22	$(1.6 \pm 0.17) \times 10^{-4}$	1.2×10^{-4}
J3	0.23	$(1.6 \pm 0.10) \times 10^{-4}$	1.4×10^{-4}
JE	0.22	$(1.1 \pm 0.09) \times 10^{-4}$	1.1×10^{-4}

Table 2.

	F_R	a (Least squares)
G2	0.14	$(1.8 \pm 0.19) \times 10^{-4}$
G3	0.17	$(2.0 \pm 0.14) \times 10^{-4}$
GS	0.26	$(1.4 \pm 0.18) \times 10^{-4}$
GC	0.13	$(0.5 \pm 0.04) \times 10^{-4}$
GN	0.12	$(0.4 \pm 0.04) \times 10^{-4}$

Longitudinal profiles

The south-north longitudinal profiles of Gjelbreen and Jenningsbreen are given in Fig. 5. For Jenningsbreen we have drawn the profile along a meridian following De Breuckbreen, across a small ridge A and then continuing the middle and upper part of Jenningsbreen (Fig. 3).

Jenningsbreen and Gjelbreen display in their upper part a rapid thickening of the ice near the trough head, characterized by a steep bedrock slope of 120 m/km, and with the bedrock dipping below sea level some 10–20 km north of the icefall. As Gjelbreen becomes shallower further north this local overdeepening might be related to the confluence with Mefjellbreen. However, former transfluence of Jenningsbreen via the low col on the ridge separating Jenningsbreen and Gjelbreen (CL in Fig. 3) is not to be excluded in this area. As this overdeepening seems to be a common characteristic of nearly all the outlet glaciers of the Sør Rondane (Van Autenboer and Decleir, 1974), the emerging picture is that of an ice covered fjord landscape. The fingerlike massifs with indentations open to the north, are filled up with ice reaching the bottom well below sea level. Some indentations penetrate deep within the massif until its southern part.

Glacier Simulation

Model formulation

With the construction of the longitudinal and cross profiles we have set the scene for an experiment in which we will try to simulate the present behaviour of the outlet glaciers in this part of the Sør Rondane. A flow line was constructed perpendicular to the surface contours from the ice divide towards the coast, along the glaciers. Bedrock topography in the mountains was obtained from the cross-sectional and longitudinal glacier profiles. Based on airborne radio echo sounding profiles, flown by Japanese scientists (Nishio and Uratsuka, 1991), we were able to prolong the bedrock topography northwards and southwards of the range. Given thus the subglacial and surface topography, as well as

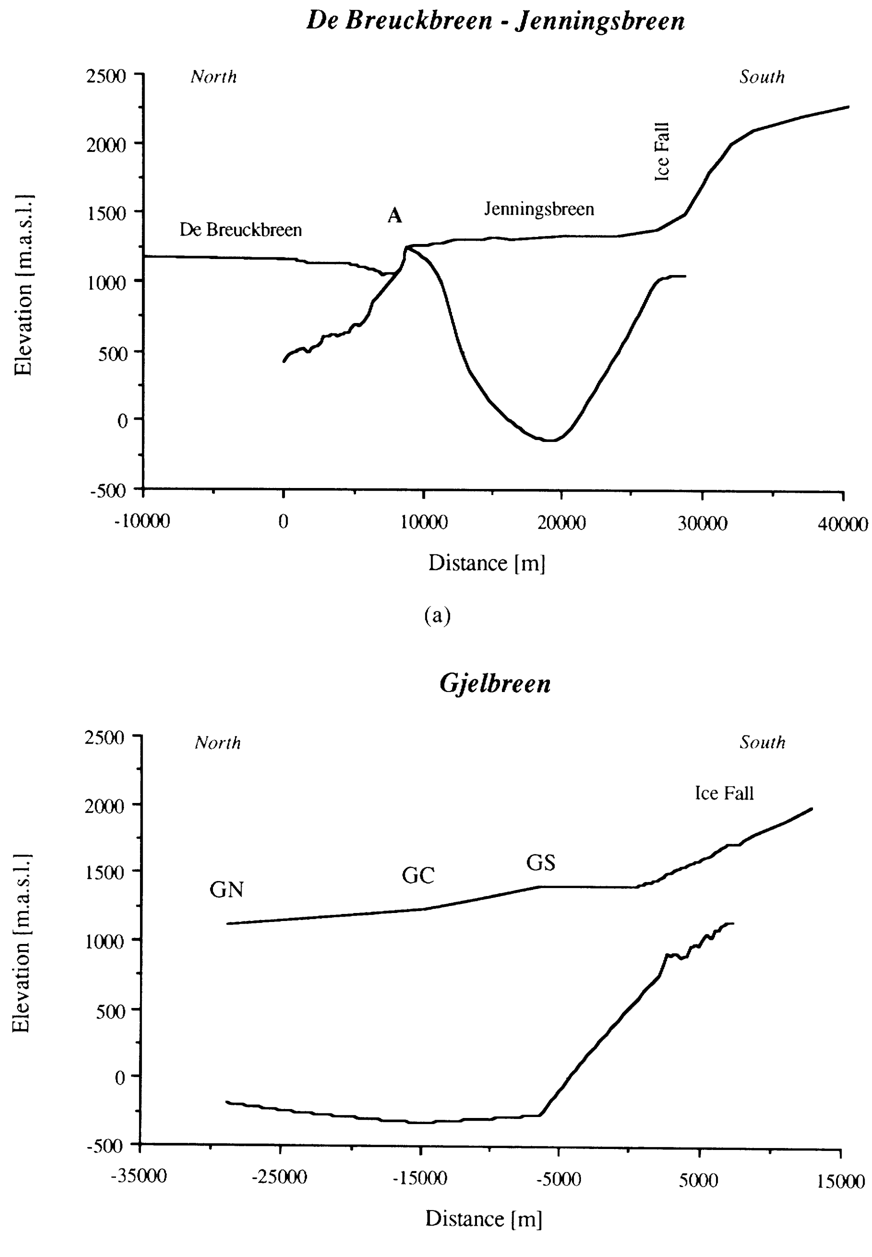


Fig. 5. Longitudinal profiles of Jenningsbreen (a) and Gjelbreen (b). The capital letters on the ice surface of Gjelbreen indicate the position of the cross profiles (see Fig. 3).

the data for surface temperature and mass balance (Fig. 2), along a complete flow line it is possible to apply a numerical ice flow model similar to the one used by Oerlemans and Van der Veen (1984).

This model was first applied to preliminary data in the Sør Rondane by Pattyn *et al.* (1989), where a complete overview of the employed modelling technique can be found. The results obtained with the present study may differ somewhat from Pattyn *et al.* (1989), but new input (subglacial topography) and reference material (present ice thickness and surface velocities), and a better mass balance distribution function account for these. A fixed grid in space is used, with 120 grid points along the x- axis and grid spacing of 5 km. The first grid point is situated on the polar plateau at an elevation of 3500 m a.s.l. The last grid point 50 km north from the present ice shelf front near the border of the continental shelf, so that the grounding line can move freely over the shelf. Sensitivity experiments (Pattyn *et al.*, 1989) already pointed out that a shift of the ice divide further inland did not alter the model results, but imposed larger demands on the computation time. The continuity equation, defining conservation of the mass, relates the local rate of change of ice thickness with the flow field. The latter is found by substituting the basic stress equilibrium in a Glen-type flow law. In the mountain range, where the ice sheet transforms into a valley glacier a shape factor is introduced to account for the effect of the side drag on the valley walls (Paterson, 1981). Since the cross sections of the valleys in the Sør Rondane have a parabolic shape, this factor is found to vary between 0.6 and 1.0, depending upon the form ratio of the valley profiles (a higher form ratio implies more side drag and thus lower velocities).

The model also allows for the temperature dependence of the flow properties of ice. This is important since a 10°C temperature shift implies an order of magnitude change in strain rates. The thermodynamics were incorporated using a method described by Oerlemans and Van der Veen (1984) by expanding the vertical temperature profile as a second degree polynomial in terms of z, the height above the bedrock. The coefficients are then obtained from the lower (geothermal heat flux) and upper (surface temperature) boundary condition and the vertically integrated heat equation (Oerlemans and Van der Veen, 1984).

Simulation of the present-day glaciation (Jenningsbreen and Gjelbreen)

Figures 6a and 6b show the reference run for Jenningsbreen and Gjelbreen where the model has been allowed to relax to a stationary state and where the model tuning parameters (one for the ice sheet, one for the ice shelf; Pattyn *et al.*, 1989) have been adjusted to obtain the present ice thickness distribution. It can be seen that a reasonable fit has been obtained between the modelled and actual ice surface in accordance with a realistic mass balance and temperature regime as prescribed, and taking into account an ablation value of 0.10 m/y in the upper regions of the mountain valleys.

The ice surface profile of the outlet glaciers show a marked concavity in the otherwise convex topography of the East Antarctic ice sheet and reflects the effect of the mountains. The concavity is due to (i) the stiffening and piling up of cold ice behind the mountains, (ii) the softening and increased velocity above and beyond the rock threshold where the ice channels into the outlet valleys, (iii) the decreasing velocities in the northern and overdeepened part of the fjordlike inlets and (iv) the increased mass balance on the inland ice slope. The softening of ice is due to the drastic lowering of the ice sheet over the mountains and the related increase in surface temperature and ice velocities. In the ice fall area the pressure melting point is reached so that basal sliding and active erosion might be held responsible for the cutting of the gorge.

An independent check on these simulations is feasible by comparing the computed with the measured surface velocities of the glaciers (Table 3). For Gjelbreen, the measurements in the exit area are in agreement with the simulated values. The calculated velocities for Jenningsbreen ("outlet" in Table 3) are however much too high. As explained by Pattyn *et al.* (1989) this can be overcome by modelling Jenningsbreen as a "local" glacier (Table 3), where the lower velocities are in accord with the reduced ice discharge. This is accomplished by situating the origin of Jenningsbreen not on the plateau ice divide but on a small ice dome just behind the mountain range. Evidence for this can be found in the reduction in ice supply as observed in the ice fall region of Jenningsbreen where, among others, a ridge forming part of the trough head has recently come ice free.

We can thus conclude that, in addition to creating a concave ice topography, the damming effect of the Sør Rondane results in a very reduced ice flow (Gjelbreen) and even in a glacier which is at present in the process of being cut off from the main ice supply (Jenningsbreen).

Simulation of the last glacial maximum

Although we ignore at present the precise environmental conditions reigning in the Sør Rondane during the last glacial maximum, the proximity to the central ice divide suggests the use of the climatic forcing similar to the one found for the Vostok ice core. In the model, surface temperature is perturbed to changes in background temperature (assumed uniform over the ice sheet) and local changes in surface elevation. Changes in surface temperature also affect accumulation rates in different climates. Following Lorius *et al.* (1985) the accumulation rate $M(t)$ in the past is calculated from its present value $M(0)$ time the ratio of the derivatives of the saturation vapour pressure with respect to $T_f(t)$ and with respect to $T_f(0)$, where T_f is the temperature of formation of precipitation:

$$M(t) = M(0)\exp\left[22.47\left[\frac{T_0}{T_f(0)} - \frac{T_0}{T_f(t)}\right]\right]\cdot\left[\frac{T_f(0)}{T_f(t)}\right]^2$$

with $T_f(t) = 0.67\ T_s(t) + 88.9$, where $T_s(t)$ is the surface temperature and $T_0 = 273.15$ K (see also Huybrechts, 1992). Applying this formula for a temperature decrease of 11°C during the last glacial maximum (Jouzel *et al.*, 1987), we find for each point of the glacier profile (assuming that the temperature decrease with altitude, Fig. 2) an accumulation rate which is roughly 50% of the present one.

Imposing these climatic conditions, together with an estimated 150 m sea level drop, the glacier topography of Jenningsbreen and Gjelbreen was again simulated and compared with the reference run for the present climate. Table 4 gives the difference in altitude at some key points of the profiles and Fig. 6c shows the result for Gjelbreen. The discrepancies between the differences for Gjelbreen and Jenningsbreen (Table 4) might be due to the different methods employed to obtain them. For comparison of the glacial simulation with the present one we were forced to treat Jenningsbreen as an outlet glacier. As pointed out in the above section, although the resulting topography is an acceptable one, a physical inconsistency arises generating exaggerated velocities. The results for Gjelbreen seem therefore a better basis for comparison with paleo-glacial reconstructions based on field evidence as well as for general discussion.

Figure 6d shows for Gjelbreen another ice age experiment forced only by a drop of 150 m in sea level and −11 °C in sea level temperature (preserving thus the present mass balance). A substantial increase in ice thickness is noted on the plateau due to the reduced deformation of the colder ice, while a very large (>1000 m) increase is observed north of the mountains due to the grounding of the ice sheet and the advance of the grounding line until the edge of the continental shelf. This increase in thickness due to the shift in grounding line is apparently dammed by the mountain range. The resulting profile still contains the conspicuous concavity notwithstanding the appreciable ice cover over the mountains. This continued existence of the concavity characterizing the

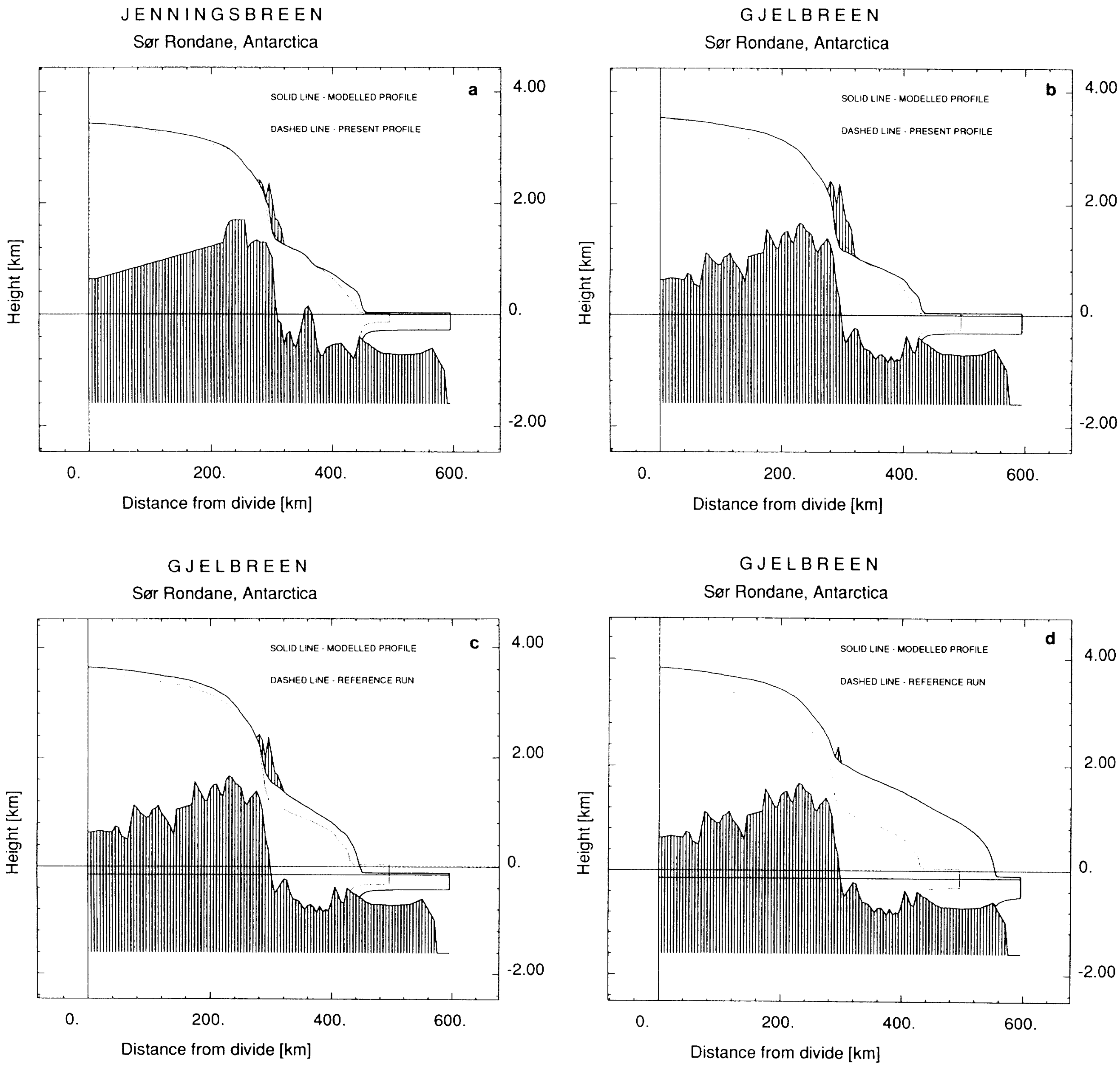


Fig. 6. Model simulations of present topography of Jenningsbreen (a) and Gjelbreen (b) and glacial simulations of Gjelbreen (c) and (d). See text for more information.

Table 3. Velocities in m/y.

	Gjelbreen		Jenningsbreen		
	calc.	observed	outlet (calc)	observed	local (calc)
Max slope	69	—	102	16	28
Middle glacier	33	—	40	1	9
Exit area	13	10	20	1	5
Ice shelf	286	300	366	300	306

Table 4. Altitude differences of central glaciers during glacial stage (height in m).

	Gjelbreen	Jenningsbreen (outlet)
Plateau	+110	+115
Dufek Massif	+128	+126
Ice slope	+390	+232
Middle glacier	+348	+225
Exit area	+207	+214
Piedmont area	+215	+147

profile of outlet glaciers seriously undermines older results based on the use of parabolic and elliptical profiles for reconstructing the paleo-surface of outlet glaciers (e.g. Denton and Hughes, 1981; Mayewski and Goldthwait, 1985).

Coming back to the more realistic ice age simulation of Fig. 6c (Table 4), where the reduced accumulation rate due to the lower temperatures is taken into account, we notice an advance of the grounding line of 20 km and an increase in altitude of the polar plateau of some 100 m. This relatively moderate steady state response of the Antarctic ice sheet during the last glacial maximum is in accord with a more advanced three-dimensional modelling for Antarctica as a whole (Huybrechts, 1990a, b, 1992). In the mountain range, our glacial simulation for Gjelbreen produces a somewhat steeper glacier surface as compared to the present reference run (Fig. 6c), varying from 215 m in the north to 390 m in the south (Table 4). In this ice age model, based on physical grounds, there is no room for a drastic increase in surface elevation at the foot of the ice fall (>1000 m) as obtained by Hirakawa and Moriwaki (1990) in their reconstruction of the former ice surface. These authors also report in their study, which is completely based on geomorphological and glacio-geological evidence, an average increase in ice thickness of 400 m. As this is of the same order of magnitude as our findings, it may be concluded with some confidence, that in general the morphology and glacial sediments point to a 300–400 m higher glacial stand characterizing the last glacial maximum. In our opinion, the evidence for a higher ice surface is witness for an environment predating the last glacial maximum and characterizing much wetter environmental conditions (confer but not necessary equal to Fig. 6d).

Aspects of deglaciation

The polycyclic origin of the present glacial morphology in East Antarctica urges us to think about the role of the frequent and undoubtedly long-duration periods of deglaciation (at least compared to the last glacial maximum, see Porter, 1989). The Sør Rondane forms a small ablation island within the continental-sized accumulation area of the East Antarctic ice sheet. A lowering of the ice sheet topography results in reduced mass transport to the glaciers traversing the mountains. Both the modelling experiment as well as the small ice discharge values (Van Autenboer and Decleir, 1978) indicate that such is the case for Jenningsbreen. Speculating then on the future behaviour of this glacier, it is quite possible that, once completely severed from the ice supply coming from the south, the substantial ablation in the upper part of the valley will cause a continued lowering of the ice surface of upper Jenningsbreen. In the end, the ice flow will reverse (north to south), the glacier being fed from the north by ice, circumventing the central part of the mountains. In our opinion, several small scale examples of such scenario can actually be observed in the Sør Rondane.

In the Brattnipene area for instance, De Breuckbreen lies in the northward extension of Jenningsbreen and was previously fed by this glacier, the ice flowing over the low col A (Fig. 3) on the ridge separating the two glaciers (Van Autenboer, 1964; Hirakawa *et al.*, 1988). Once cut off from the Jenningsbreen supply, the surface of De Breuckbreen lowered by some 200 m, reversing the flow as can actually be observed (Fig. 5b). The inverted moraines both on the ice and in the dry valley of Brattnipene (Pierrebotnen on the Norwegian map, near B on Fig. 3) are another witness (but now the final phase) of this deglaciation process of which many other examples can be found in the Sør Rondane.

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