—研究論文—
Scientific Paper

Subglacial Topography in the Central Sør Rondane Mountains, East Antarctica: Configuration and Morphometric Analysis of Valley Cross Profiles

Frank PATTYN* and Hugo DECLEIR*

セールロンダーネ山地中央部における氷河の基盤地形：
氷河横断形状と形態解析

Frank PATTYN* · Hugo DECLEIR*

要旨：本論では，セールロンダーネ山地中央部で実施した氷河の基盤地形観測のすべてを概観する．データは重力計法と電波探査法によるもので，この地域で以前に行われたベルギー隊および JARE-28, -32 の観測で得られた氷河の横断方向，縦断方向のプロファイルである．こうしたデータを基に，この地域の氷河の基盤地形をまとめた．さらに，氷河横断基盤形状の形態学的特徴の解析方法を示すが，この方法はこれまでのべき級数法則よりも良い結果をもたらした．最後に，氷河横断基盤形状の本解析は，現在の基盤地形が氷河の削剥能力と関連した複雑な地形発達過程にあることを明らかにした．

***Abstract*:** In this paper an overview is given of all subglacial topography measurements carried out in the central Sør Rondane Mountains. Data of glacier valley cross and longitudinal profiles were gathered by gravimeter and radio echo sounding measurements during former Belgian expeditions and during the Japanese Antarctic Research Expeditions JARE-28 and JARE-32. Based on these data, a map of the subglacial topography in the central mountain area was compiled. Furthermore, a method is presented for analysing the morphometric characteristics of valley glacier cross profiles, which is shown to give better results than former power law equations. Finally, the morphometric analysis of the present glacierized valley cross profiles revealed a complex development regime, linked with the erosion potential of the glacierized area.

1. Introduction

The Sør Rondane is a typical coastal margin mountain range in Eastern Dronning Maud Land, stretching over a distance of 220 km in east-west direction, situated approximately 200 km from the coast, with the highest elevation being 3000 m a.s.l. (Fig. 1). This range forms part of a chain of mountains surrounding the East Antarctic continent from the Borg Massif in Western Dronning Maud Land (5°W–73°S) to the Yamato Mountains in Eastern Dronning Maud Land (35°E–72°S), which forms the border with the ice sheet of Enderby Land and the Shirase Drainage Basin. Most of the ice flow coming from the polar plateau is drained along both sides

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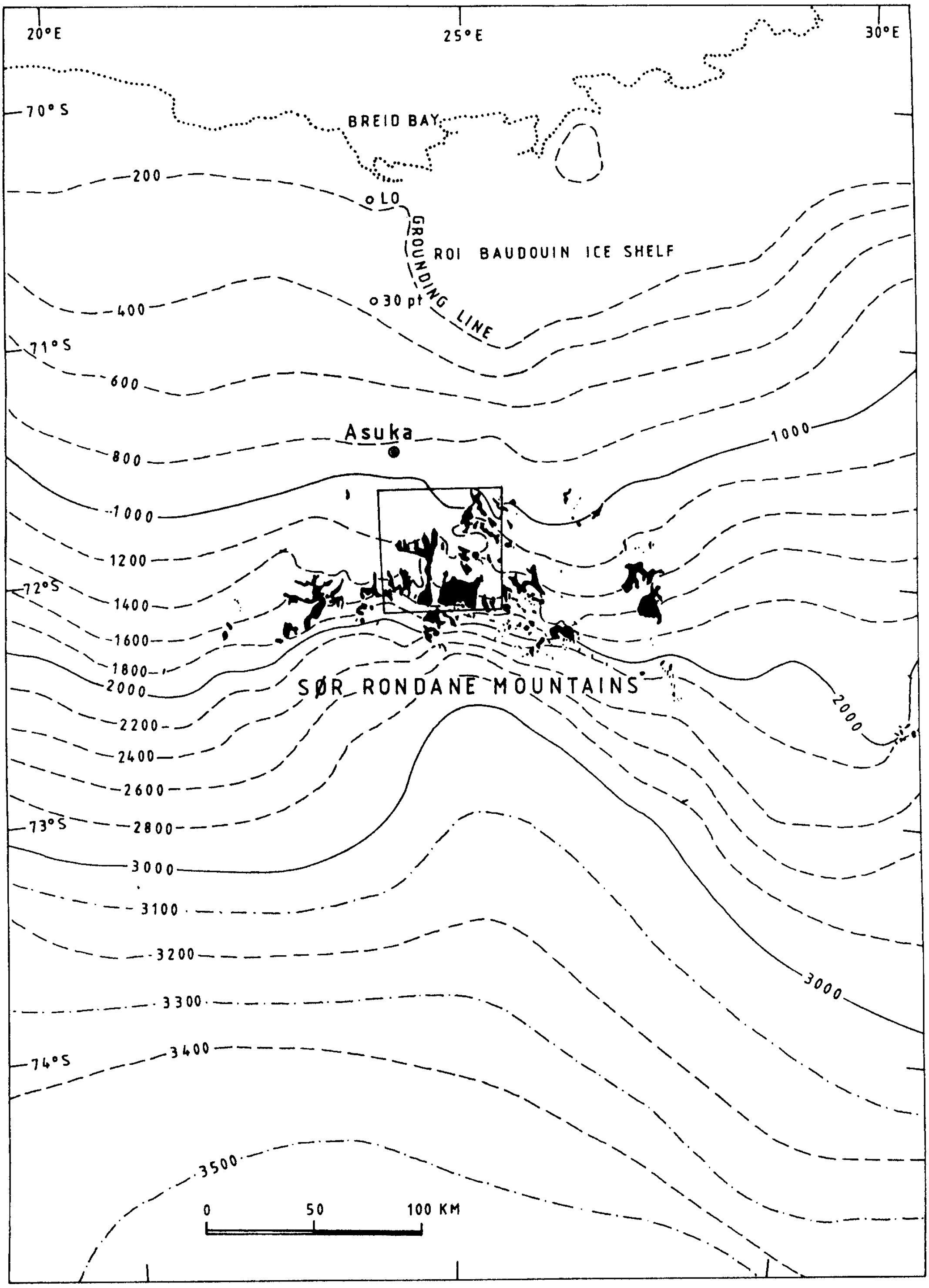


Fig. 1. Topographic map showing the Sør Rondane Mountain range, the inland ice slope and Breid Bay.

of the Sør Rondane, giving rise to two large outlet glaciers, Hansenbreen in the west and Byrdbreen in the east. A small number of outlet glaciers cut through the range, all characterized by varying ice fluxes (Gunnestadbreen, Jenningsbreen, Gjelbreen, ...). From airborne radio echo sounding (NISHIO and URATSUKA, 1991) it was found that the subglacial topography south of the mountain range lies well above sea level (1000 to 2000 m a.s.l.), while between the Sør Rondane and the coast the ice sheet has a marine configuration with a bedrock elevation ranging from −100 to −300 m a.s.l. Therefore, at the coast, the ice sheet is drained into a well-developed ice shelf (Roi Baudouin Ice Shelf).

The first ice thickness measurements in the Sør Rondane Mountains were carried out by VAN AUTENBOER and BLAIKLOCK (1966) and VAN AUTENBOER and DECLEIR (1974, 1978). They applied the gravimetric method in an approach to estimate the total glacier ice discharge through the range. However, due to the closing down of the Belgian Base Roi Baudouin in 1967, research activities in this area were adjourned. With the establishment of the Japanese Asuka Station in 1985, 50 km north of the Sør Rondane Mountains, both airborne and oversnow radio echo sounding surveys were carried out for the first time during the austral summer 1986–87, in the coastal area and south of the mountain range (NISHIO and URATSUKA, 1991). Also ice thickness measurements were carried out in the central mountain area, both by radar and gravimetry (DE VOS and DECLEIR, 1988; DECLEIR *et al.*, 1989). Finally, during JARE-32 (austral summer 1990–91) both gravimetric measurements and radio echo sounding surveys were carried out in the central Sør Rondane on outlet glaciers as well as on local glaciers and valley glaciers (PATTYN *et al.*, 1992, 1993). This paper reports on the ice thickness measurements carried out so far in the central Sør Rondane and on a preliminary analysis of the subglacial topography. A new method for analysing the valley cross sections is also presented.

2. Ice Thickness Measurements: Methodology

2.1. *The electromagnetic method*

The usual equipment for radio echo sounding of ice have been powerful instruments, intended for installation in aircrafts or oversnow vehicles. However, in mountain areas and on steep and rough glacier surfaces neither of these instruments can be used. The Scott Polar Research Institute developed a small Radio Echo Sounder for operation on a wooden Nansen sledge. Both transmitter and receiver, power supply and recording equipment were fitted in one alluminium case, allowing room for both instrumentation (including two antennas) and operator on the small sledge. The electromagnetic signal was transmitted at a frequency of 160 MHz. The performance of the sounder was limited to 1000 m, limiting its use to depth profiles on smaller outlet glaciers and valley glaciers in the mountain range. For the deeper parts of the glaciers the gravimetric technique was applied. The gravimetric method is also successful on moraine covered ice surfaces, where the electromagnetic method fails and which are inaccessible for snow sledge or snow scooter.

2.2. *The gravimetric method*

VAN AUTENBOER and BLAIKLOCK (1966) and VAN AUTENBOER and DECLEIR (1974, 1978) applied the gravimetric method in the Sør Rondane in an approach to estimate the total glacier discharge through the range. Also during the summer field seasons of JARE-28 (DE VOS and DECLEIR, 1988; DECLEIR *et al.*, 1989) and JARE-32 (PATTYN *et* *al.*, 1992, 1993) similar measurements were carried out with a Worden (JARE-28) and a LaCoste & Romberg (JARE-32) gravity meter. Cross sections of the most important outlet glaciers were constructed using Talwani's method (TELFORD *et al.*, 1976) for modelling two-dimensional gravity anomalies. According to this method, the ice thickness in a valley cross section is calculated by an iterative procedure in which the computed gravity effect of a model cross section is compared with gravity values measured on the glacier surface. Generally the areal integral representing the gravity effect of the two-dimensional ice mass is replaced by a line integral which is then numerically solved by a polygonal approximation of the periphery of the ice body. In this case the upper vertices of the polygon—which correspond with the gravimeter observation stations—are known, while the lower vertices—vertically beneath the same observation points—relate to the unknown subglacial bedrock (Fig. 2). It is also possible to compose the unknown ice mass of a set of rectangular vertical ice prisms extending from the bedrock to the ice surface. Each prism then has a gravimeter station as boundary along the cross profile (Fig. 2).

Thus, the gravimetric method for ice thickness determination—unlike the electromagnetic method—requires difficult modelling and is not always unambiguous. DECLEIR *et al.* (1989) presented a comparison between radar and gravimetric soundings of two glacier cross profiles. Taking the radar thicknesses as a standard, it appeared that the gravity method highly depends on the number of gravity stations and on the modelling procedure employed. They also inferred an underestimate of 10% in ice thicknesses obtained in previous studies in Sør Rondane (VAN AUTENBOER and DECLEIR, 1974, 1978). In this respect, the recommendations of DECLEIR *et al.* (1989) were followed and all gravimeter soundings in this study were analysed and modelled in the same (unambiguous) way. In case of sufficiently high density of gravity stations the prism method was used, because it is less liable to instability.

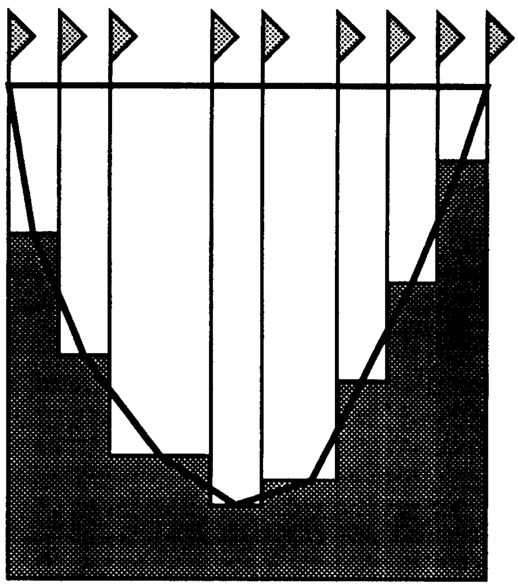


Fig. 2. Polygonal and prism approximation of glacier cross section.

However, when only very few gravity stations are available (5 to 10) the prism method overestimates the ice thickness because it produces an exaggerated vertical wall effect near the side of the glacier (Decleir *et al.*, 1989). In that case the polygonal method was applied.

3. Subglacial Morphology in the Central Sør Rondane

Figures 5 to 13 display all valley cross and longitudinal profiles measured in the central Sør Rondane Mountains, grouped per glacier type (outlet glaciers, local glaciers) and name. Table 1 shows for each profile its major characteristics, while the geographic distribution of the profiles is given in Fig. 3. Most of the ice thickness profiles are glacier cross profiles, but a few longitudinal lines were also run. The map of the subglacial relief (compiled after these measured data) is shown in Figs. 4a and 4b.

In the central part of the Sør Rondane, Jenningsbreen and Gjelbreen have cut a 40 km long U-shaped valley through the mountain range. 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).

The north-south longitudinal profiles of Gjelbreen and Jenningsbreen are displayed in Figs. 5a and 5b respectively. For Jenningsbreen we have drawn the profile along a meridian following De Breuckbreen, across a small ridge A and then continuing the middle and upperpart of Jenningsbreen. Both Gjelbreen and Jenningsbreen display in their upperpart (near the ice fall), proceeding northwards, a rapid thickening of the ice characterized by a steep bedrock slope of 120 m/km (the "trough head"), with the bedrock dipping below sea level some 10–20 km north of the ice fall. The maps (Figs. 4a and 4b) underscore the subglacial relief as an ice covered fjord landscape. The cross profiles of the oulet glaciers Jenningsbreen and Gjelbreen (Figs. 6a–f and 7a–e) display mostly the characteristic U-shape of the ice filled valleys 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).

The Sør Rondane is thus divided in a number of massifs, separated by the U-shaped valleys, creating a complex landscape, characterized in the first place by selective linear erosion. The intervening massifs on the other hand are marked by small valley glaciers and local ice caps. The dominant flow direction of the larger outlet glaciers as well as the local valley glaciers, is from south to north, even though some deep glaciers (Nipebreen and Mefjellbreen) between Gjelbreen and Komsbreen flow in a marked east-west direction. The cross sections of Mefjellbreen (Figs. 8a–c) clearly reveal a north-south tending subglacial ridge, linking the high grounds of Mefjell (in the south) with the Berckmanskampen and Menipa area (in the north). This subglacial ridge is probably a remnant of the ridge between the south-north flow direction of the present local glaciers and dry valleys of Mefjell, cascading into Gjelbreen and Komsbreen respectively. In a later stadium, an increased ice flow from

Table 1. Overview and origin of the longitudinal and cross profiles of glaciers in the central Sør Rondane Mountains, Antarctica. References: VAN AUTENBOER (VAN AUTENBOER and DECLEIR, 1974, 1978), JARE-28 (DE VOS and DECLEIR, 1988; DECLEIR et al., 1989), JARE-32 (PATTYN et al., 1992, 1993, and unpublished).

Profile	nr.	Measurement	Profile type	Source	Fig.
Gjelbreen 1	G1	Radar	longitudinal	JARE-32	5a
De Breuckbreen 2	DB2	Radar	longitudinal	JARE-32	5b
Jenningsbreen 1	JB1	Gravimeter	cross	JARE-32	6a
Jenningsbreen 2	JB2	Gravimeter	cross	JARE-32	6b
Jenningsbreen 3	JB3	Gravimeter	cross	JARE-32	6c
Jenningsbreen 4	JB4	Gravimeter	cross	JARE-32	6d
Jenningsbreen Central	JC	Gravimeter	cross	JARE-28	6e
Jenningsbreen-Ellisbreen	JE	Gravimeter	cross	VAN AUTENBOER	6f
Gjelbreen North	GN	Gravimeter	cross	VAN AUTENBOER	7a
Gjelbreen Central	GC	Gravimeter	cross	JARE-28	7b
Gjelbreen Lunckeryggen	GS	Gravimeter	cross	JARE-28	7c
Gjelbreen 2	G2	Radar	cross	JARE-32	7d
Gjelbreen 3	G3	Radar	cross	JARE-32	7e
Mefjellbreen West	MW	Gravimeter	cross	JARE-28	8a
Mefjellbreen Central	MC	Gravimeter	cross	JARE-28	8b
Mefjellbreen East	ME	Gravimeter	cross	JARE-28	8c
Nipebreen	N1	Gravimeter	cross	VAN AUTENBOER	9
De Breuckbreen 1	DB1	Radar	cross	JARE-32	10a
De Breuckbreen 3	DB3	Radar	cross	JARE-32	10b
Goosenbreen	GO1	Radar	cross	JARE-32	11
Berckmans 1	BM1	Radar	cross	JARE-32	12a
Berckmans 2	BM2	Radar	longitudinal	JARE-32	12b
Berckmans 3	BM3	Radar	cross	JARE-32	12c
Pilten 1	P1	Radar	cross	JARE-32	13a
Pilten 2	P2	Radar	cross	JARE-32	13b
Pilten 3	P3	Radar	longitudinal	JARE-32	13c

the east (Byrdbreen) probably linked the eastern and western part of Mejellbreen, leading to an undisturbed east-west ice flow.

4. Glacial Valley Morphometry

In order to analyse the valley form development of glaciers, mathematical techniques are often used. The form of a valley cross profile can be described by its form ratio (*FR*) and shape factor (*f*), and by a power law equation. The form ratio (*FR*) is then defined as:

$$FR=\frac{D}{2W}, \tag{1}$$

with D the height of the trimline above the deepest part of the valley and W the half width of the valley. In addition to the form ratio, the more complicated shape factor can be introduced, defined as the cross-sectional area (Ar) divided by the height of the trimline above the deepest part of the valley (D) multiplied by the trough perimeter (Pe) (NYE, 1965), or :

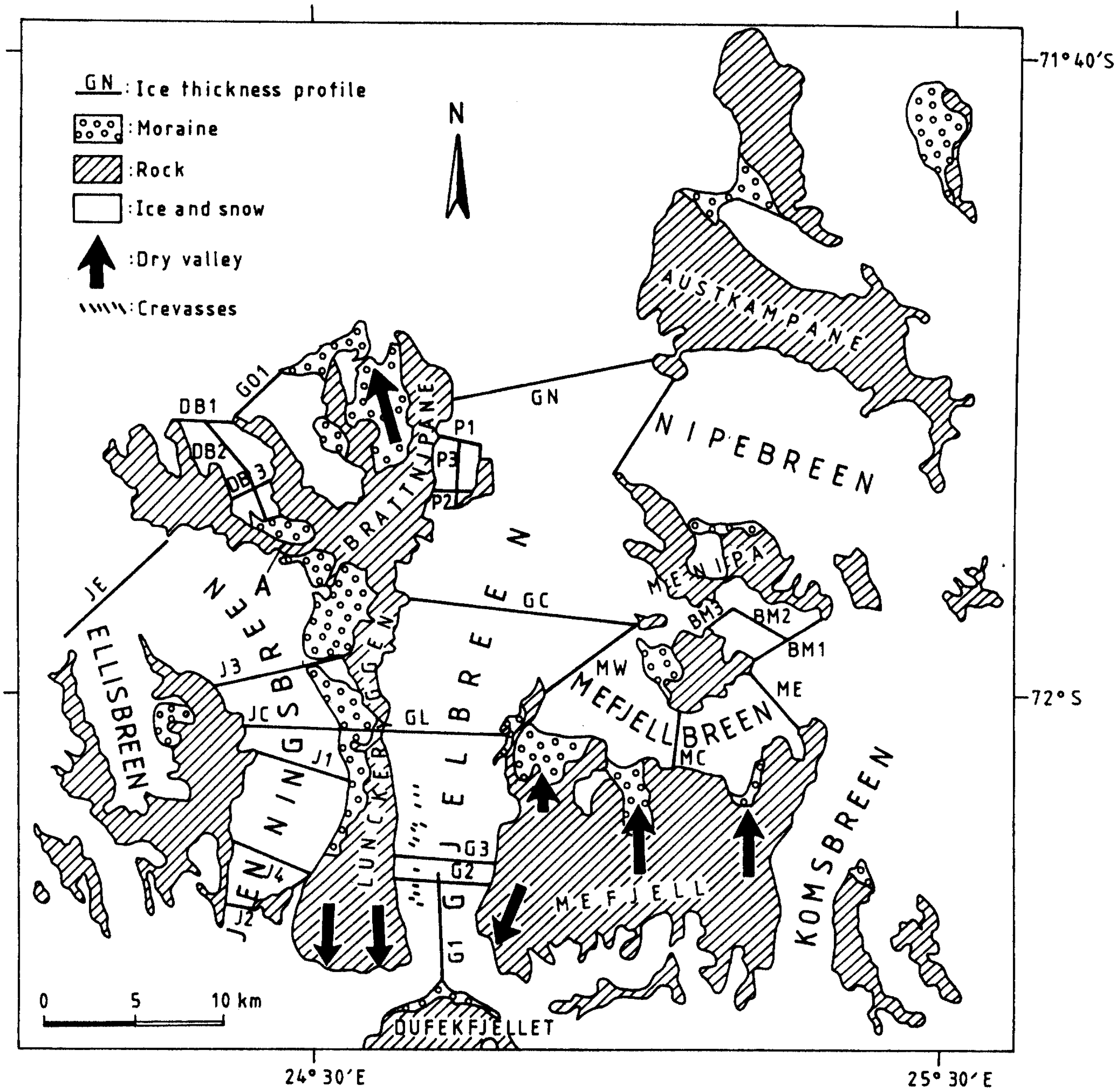


Fig. 3. Detailed map of the central part of the Sør Rondane (inset of Fig. 1) displaying the main outlet glaciers and the measured ice thickness profiles.

$$f=\frac{Ar}{D\,Pe}\,. \tag{2}$$

Finally, the power law equation can be written as:

$$y=ax^{b}, \tag{3}$$

describing a curve where y is the vertical and x the horizontal distance from the origin, placed in the central lower part of the valley, to a point on the curve (GRAF, 1970). The power curve (3) can assume a wide variety of forms determined by the values given to a and b. For $b > 1.0$ the resulting curve is concave upwards to a degree that increases with higher values of b. For $b < 1.0$ the curve will be convex

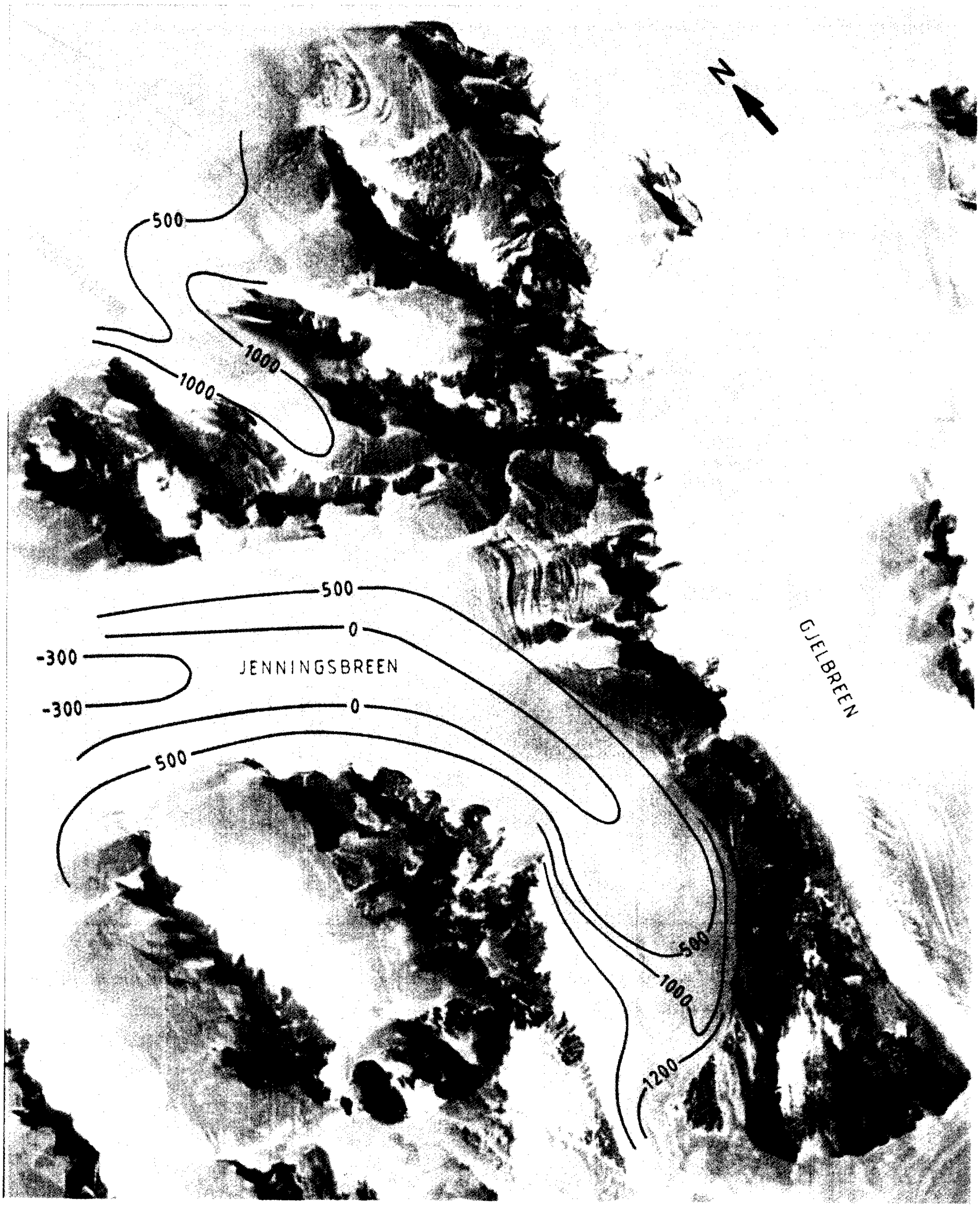


Fig. 4a. Landsat Thematic Mapper satellite image of Jenningsbreen and De Breuckbreen displaying subglacial bedrock contours in m.

upwards. In short, the coefficient *b* serves as *a* good measure for describing the valley form, while the form ratio (1) gives a complete, quantitative and dimensionless representation of the geometry of the cross section.

The use of the power law equation endured much criticism in the last decade,

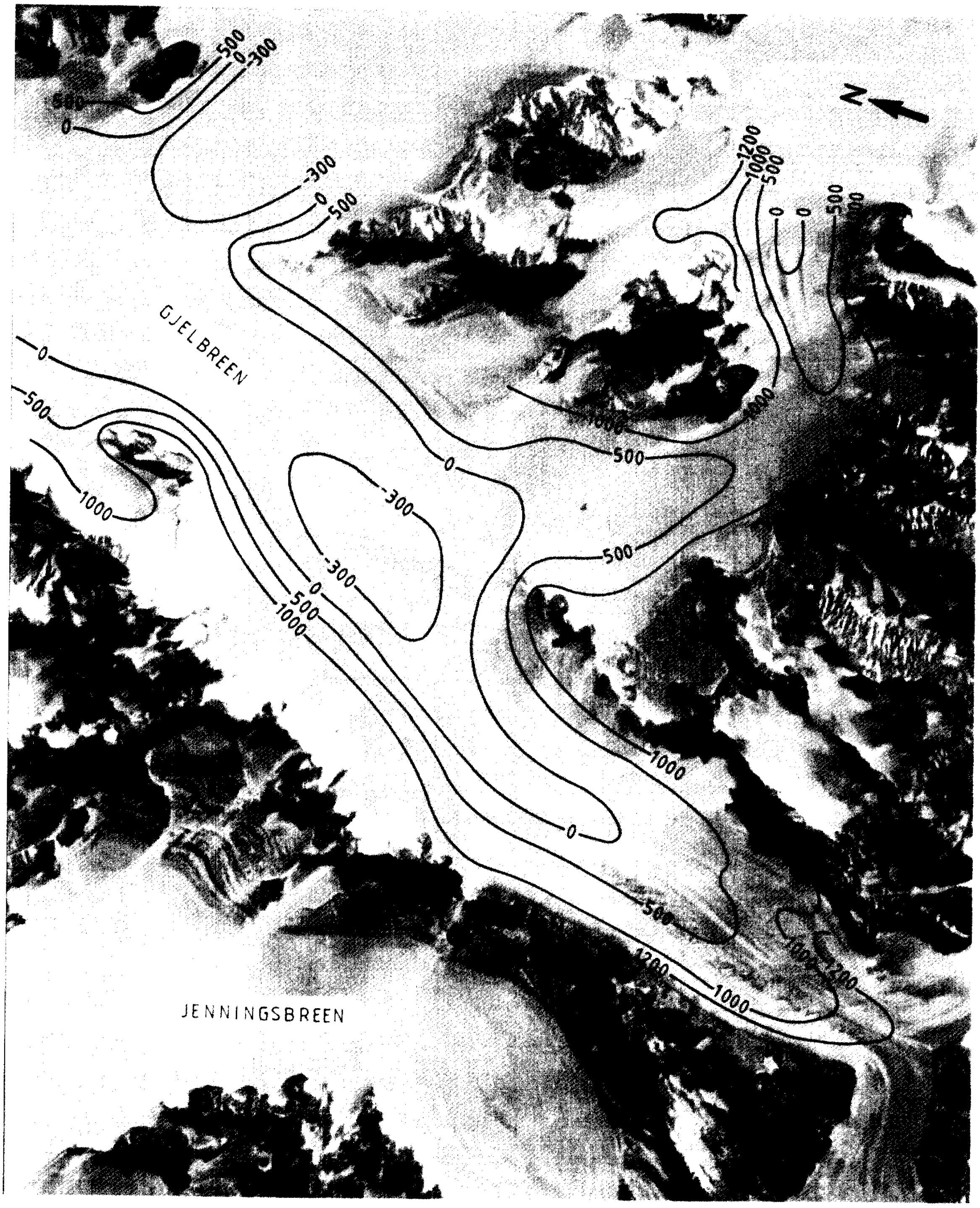


Fig. 4b. *Landsat Thematic Mapper satellite image of Gjelbreen and Mefjellbreen displaying subglacial bedrock contours in m.*

especially through the papers of WHEELER (1984), HARBOR (1990), and HARBOR and WHEELER (1992). The power law equation is generally obtained through a linear regression analysis in its logarithmic form:

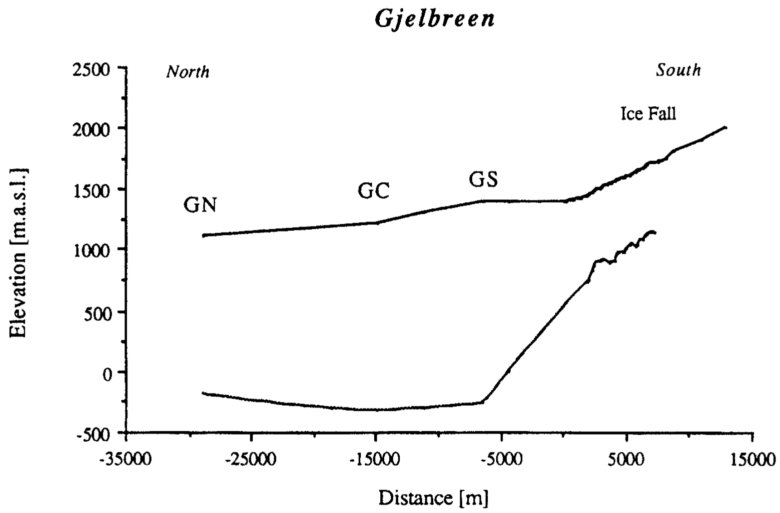


Fig. 5a. Longitudinal profile on Gjelbreen

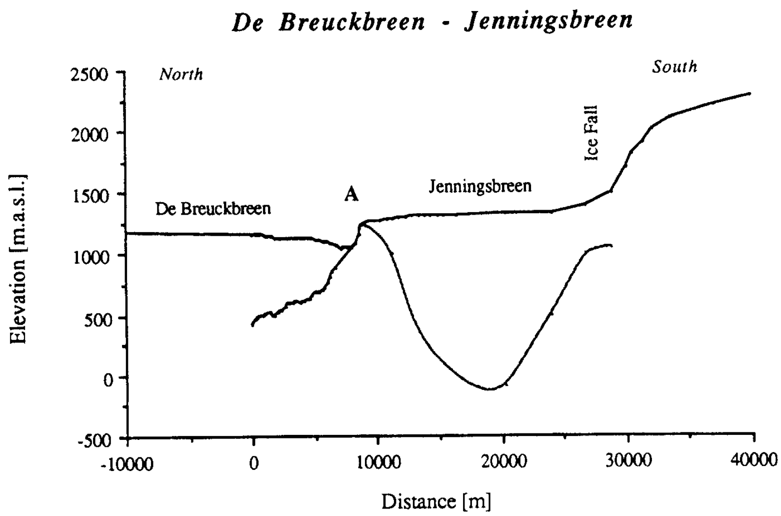


Fig. 5b. Longitudinal profile on Jenningsbreen and De Breuckbreen

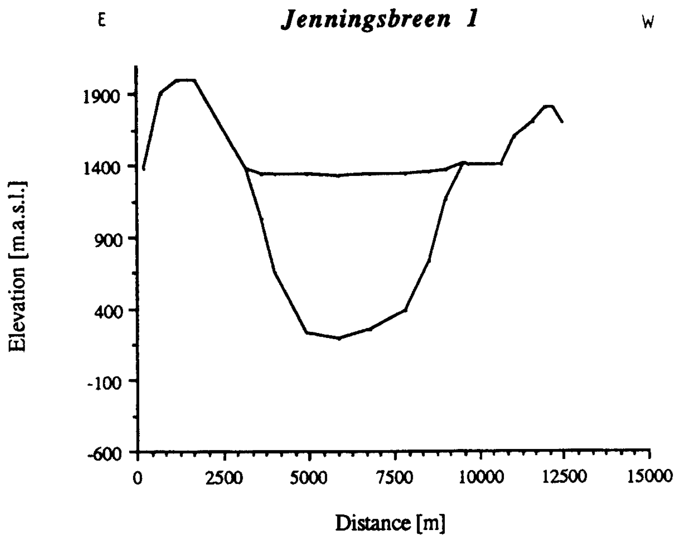


Fig. 6a.

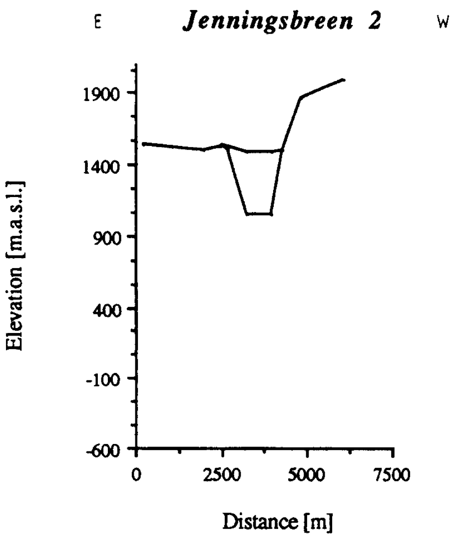


Fig. 6b.

Figs. 6a–f. Glacier cross profiles on Jenningsbreen.

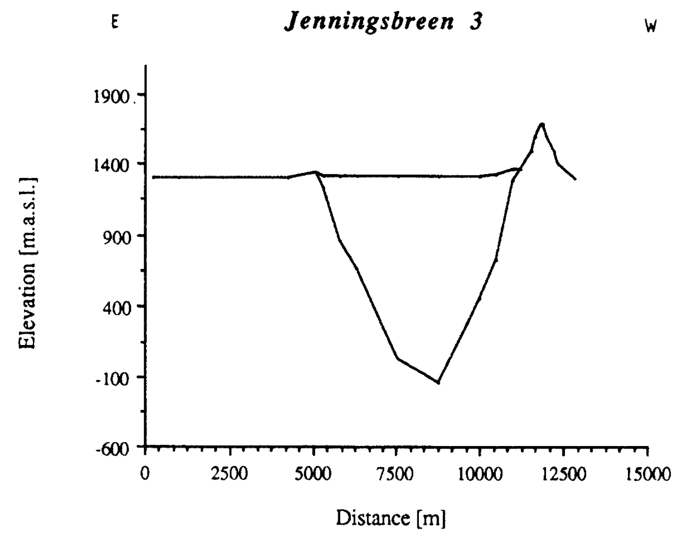


Fig. 6c.

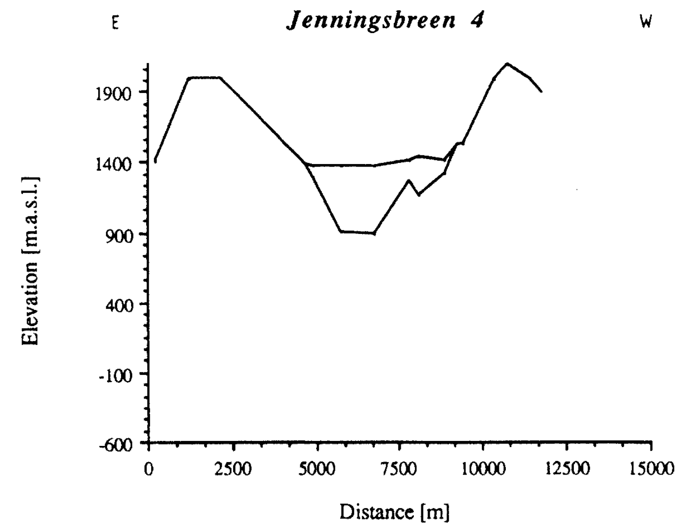


Fig. 6d.

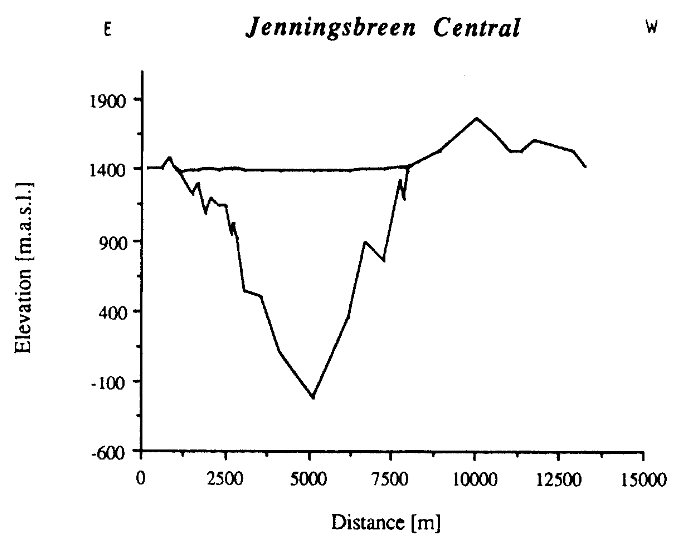


Fig. 6e.

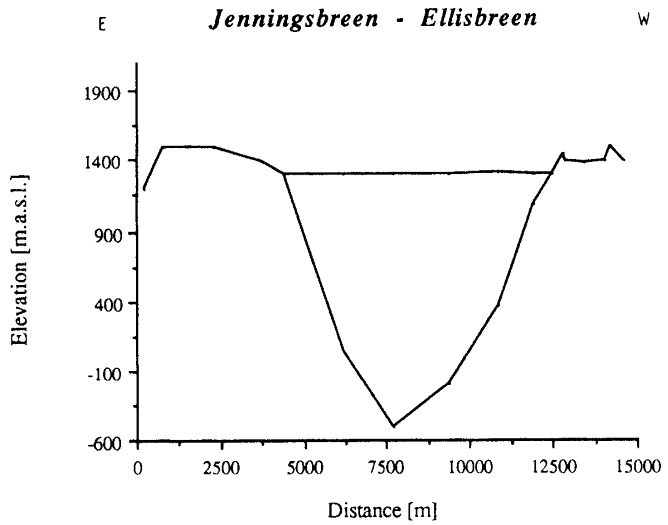


Fig. 6f.

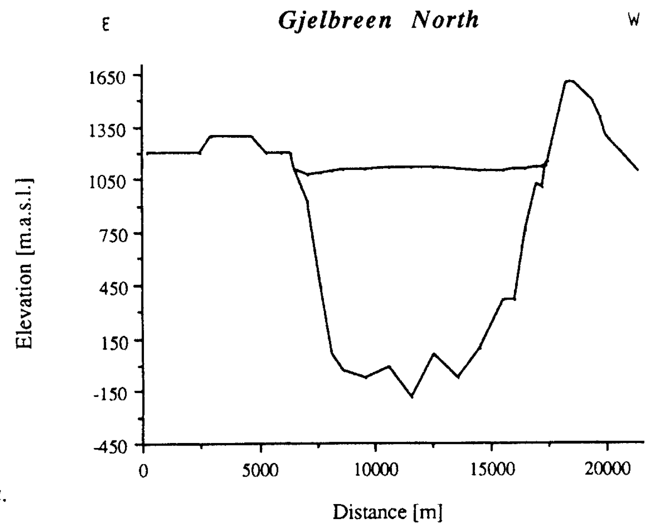


Fig. 7a.

Figs. 7a–e. Glacier cross profiles on Gjelbreen.

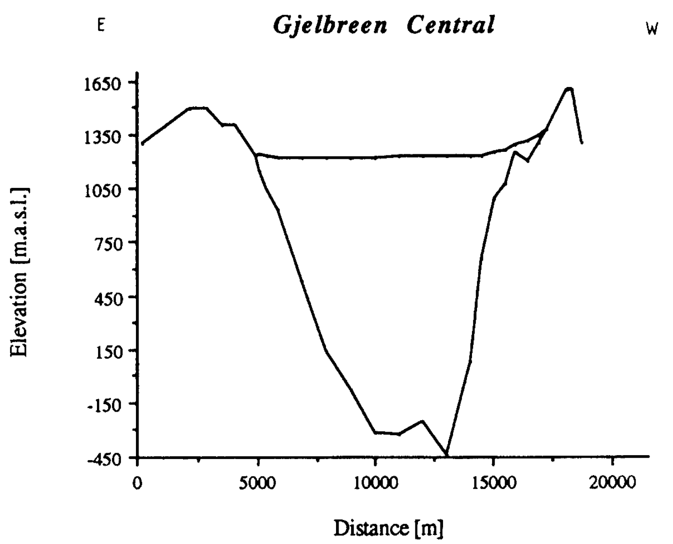
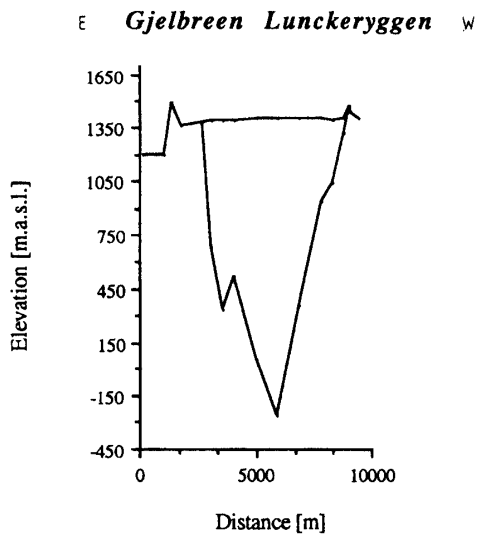


Fig. 7b.



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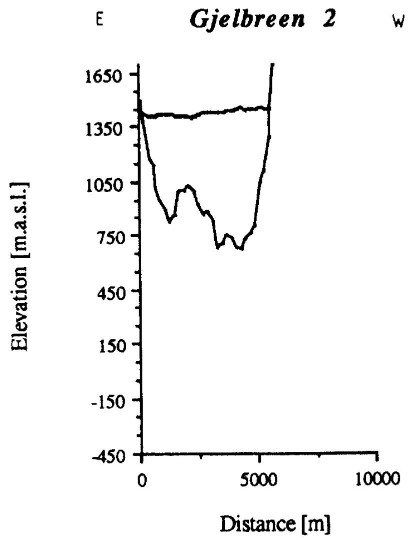


Fig. 7d.

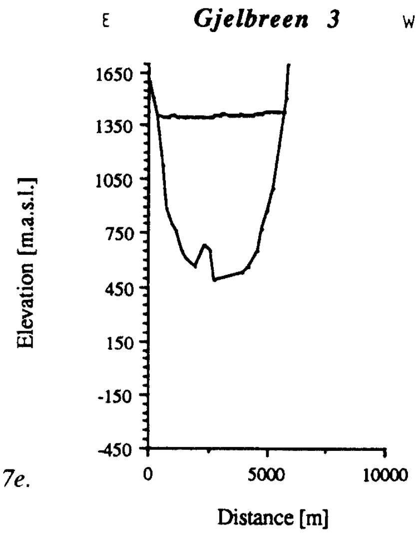


Fig. 7e.

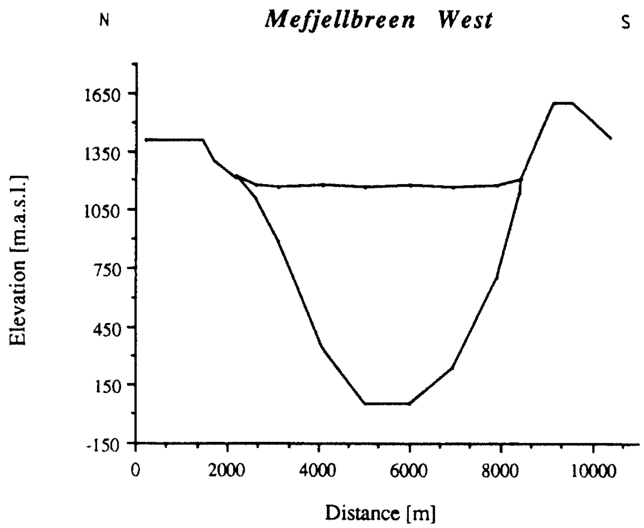
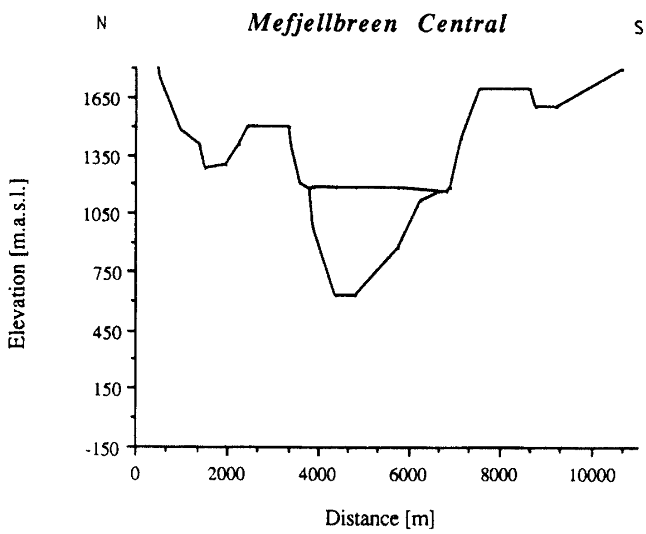


Fig. 8a.

Figs. 8a–c. Glacier cross profiles on Mefjellbreen.



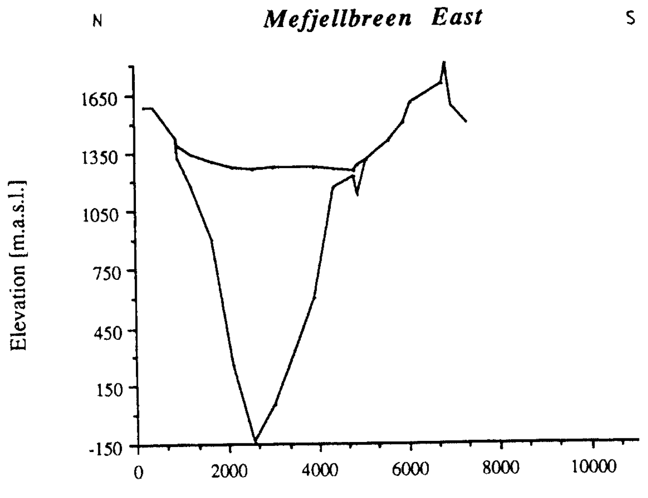


Fig. 8c.

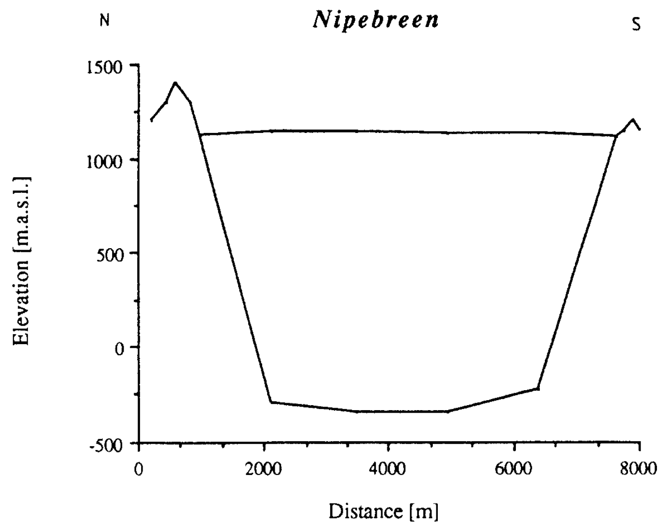


Fig. 9. Glacier cross profile on Nipebreen.

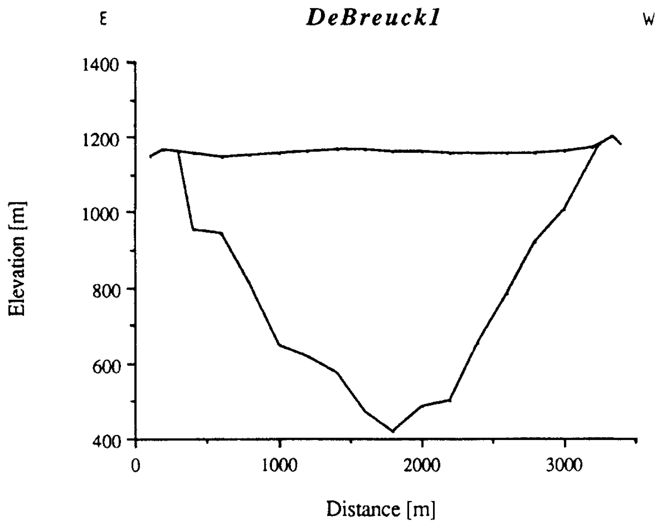


Fig. 10a.

Figs. 10a–b. Glacier cross profiles on De Breuckbreen.

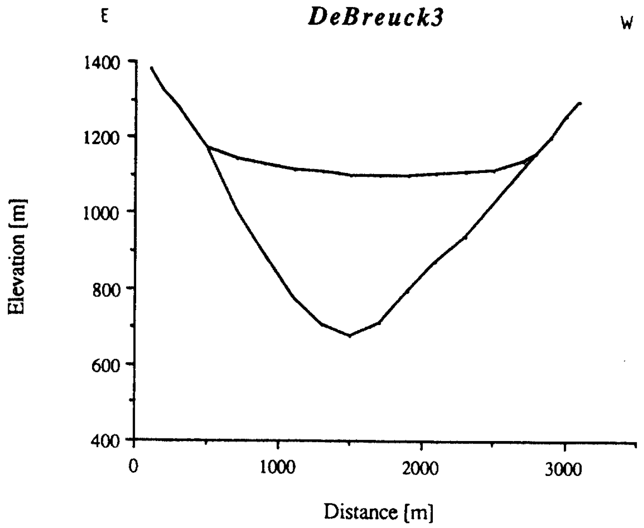


Fig. 10b.

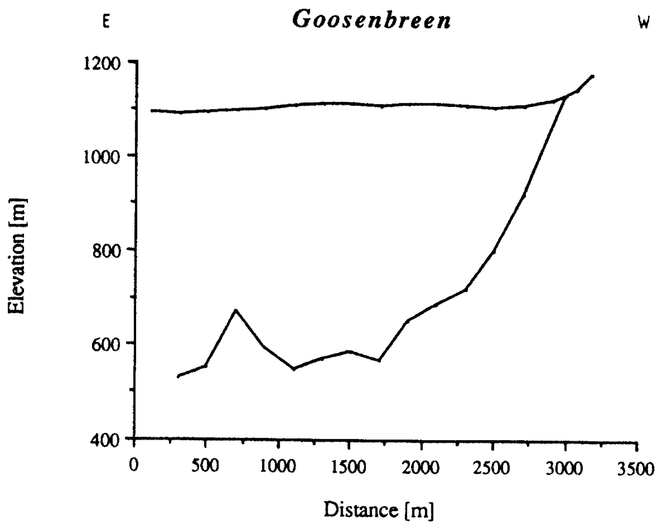


Fig. 11. Glacier cross profile on Goosenbreen.

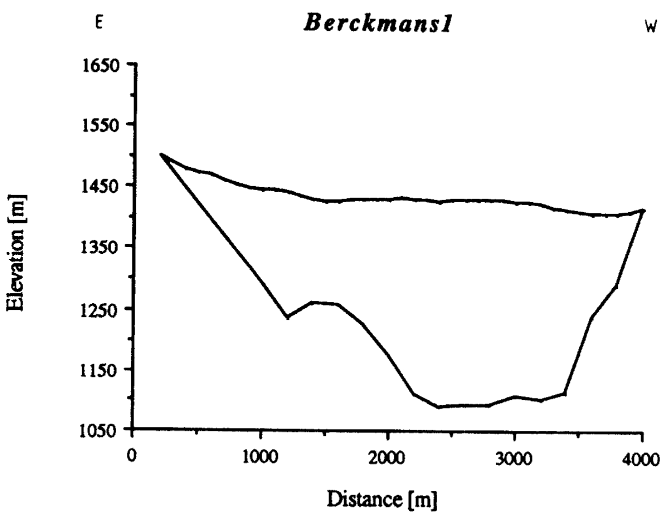


Fig. 12a.

Figs. 12a–c. Longitudinal and glacier cross profiles on small glacier in-between Berckmanskampen and Menipa (tentatively called Berckmans).

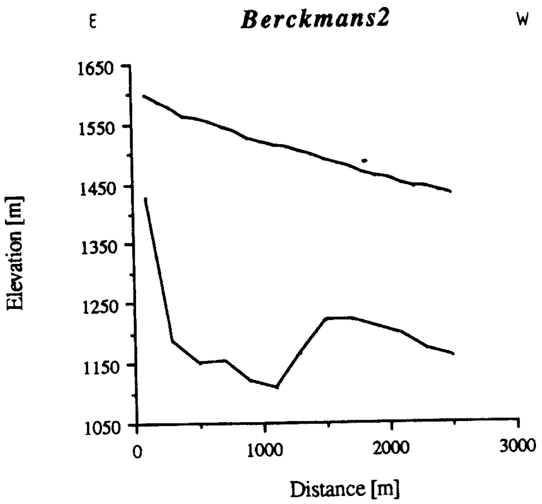


Fig. 12b.

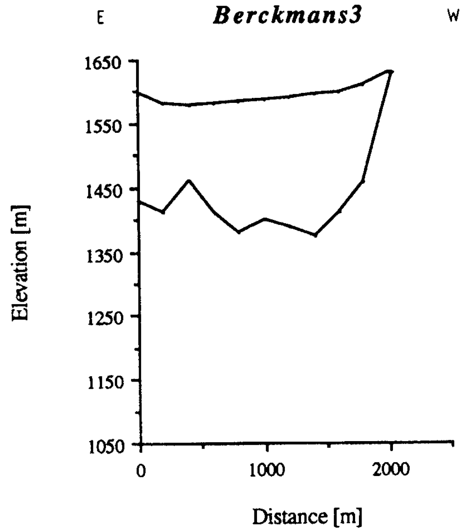


Fig. 12c.

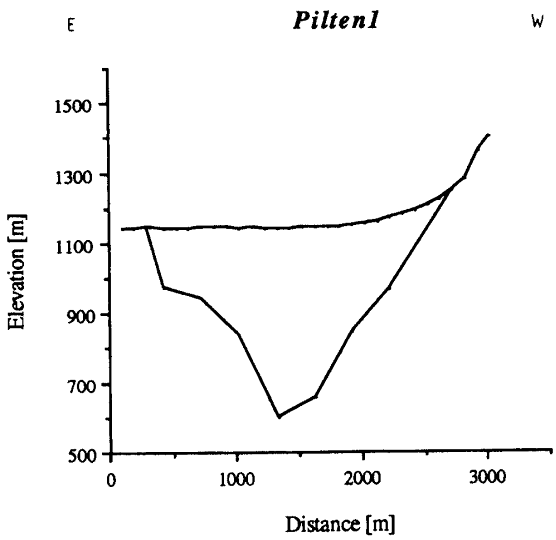


Fig. 13a.

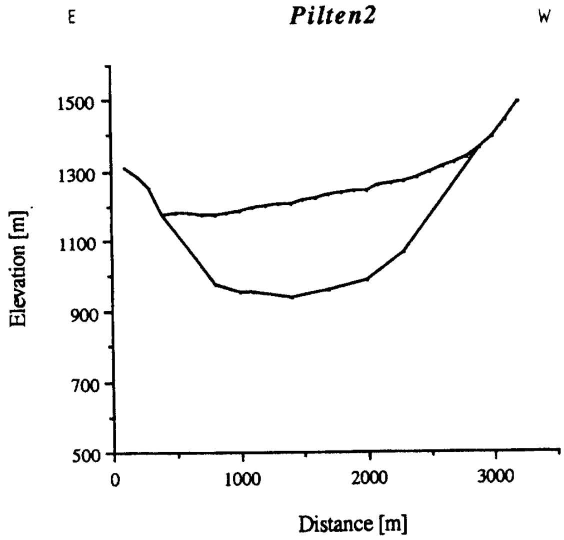


Fig. 13b.

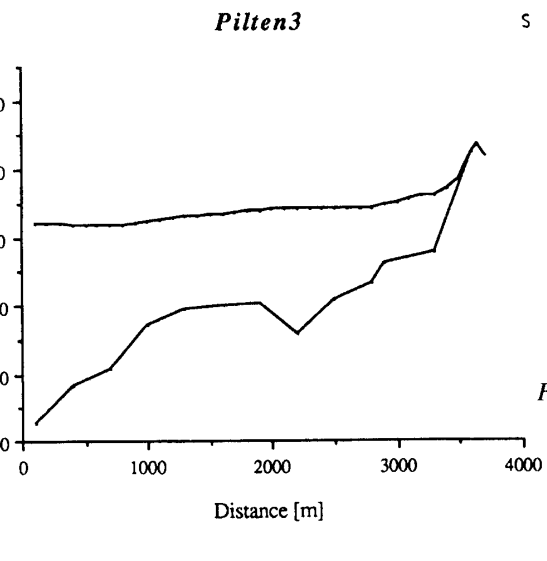


Fig. 13c

Figs. 13a–c. Longitudinal and glacier cross profiles on small glacier in-between Pilten and Brattnipane (tentatively called Pilten).

$$\ln(y)=\ln(a)+b\ \ln(x). \tag{4}$$

Equation (4) cannot be used with negative values of x (horizontal distance), so that the curves have to be fitted to the two sides of the valley as separate operations or the negative x-values have to be mirrored with regard to the origin (central midpoint of the valley floor). The curve has also no turning point and is constrained by the zero datum of the ordinate which it is unable to cross (WHEELER, 1984). Here we arrive at the key issue, *i.e.* the datum problem. There is no clear definition whether the valley floor or sea level should be used as the datum for the coordinate system. In fact, large differences occur in the b-value when applying both datums, because the value of b is a function not only of the cross profile form, but also of the origin of the coordinate system adopted for the analysis. Furthermore, a bias is introduced by the logarithmic transformation used in deriving the power law eq. (4). HARBOR and WHEELER (1992) show that observations close to the center of the valley exert the strongest influence on the regression coefficients. Finally, another bias is introduced due to glacial and post-glacial deposits masking the lower part of the erosional profile. In order to circumvent these errors, WHEELER (1984) proposes the use of a curve that is itself immune to such consideration. This requirement is fullfilled by the quadratic equation. (in this case a parabola):

$$y=P+Qx+Rx^2. \tag{5}$$

The curve generated in this way is not constrained in either direction and can be used to describe the complete cross section with one, symmetrical curve. This curve can extend below zero-datum although its turning point can be shifted in both the x- and y-direction with respect to the valley mid-point. However, the interpretation of the coefficients of (5) is less straightforward than is the case with the power law equation. Here, the shape of the valley cross profile is forced to take the parabolic form, which is assumed to be the case of a glacial valley. The correlation coefficient should then give a measure of the deviation from this ideal profile form, from which it becomes difficult to extract relevant information concerning the valley shape. In order to overcome these difficulties we opted for a new description of the power law equation, less straightforward to solve, but more consistent in the analysis. Therefore the power eq. (3) is rewritten in the form:

$$y-y_0=a(x-x_0)^{\mathrm{b}}. \tag{6}$$

When $x_0 = y_0 = 0$, eq. (6) reduces to the power law eq. (3) and when $b = 2$, the parabolic form (5) is obtained. Equation (6) is solved by the method of general least squares adjustment. To assure the symmetrical shape around the central valley axis and to remove domain errors introduced by the power equation, eq. (6) is rewritten as:

$$F=y-y_0-a.\ \exp\{b\ \ln|x-x_0|\}=0. \tag{7}$$

Since eq. (7) is non-linear it is linearized by means of a Taylor expansion, whereby only the first derivative terms are retained. The general solution then becomes (see for instance MIKHAIL and GRACIE (1981) for more detail):

$$\Delta = [B^T(A\ Q\ A^T)^{-1}B]^{-1}B^T(A\ Q\ A^T)^{-1} f^0, \tag{8}$$

with A the observation matrix containing the derivatives of eq. (7) to x and y, B the coefficient matrix containing the derivatives to a, x_0 and y_0, Q the covariance matrix, which, for the sake of simplicity is taken as the unity matrix, f^0 the solution vector, and Δ the adjustment to the unknown coefficients a, x_0 and y_0. Introducing a first estimate for the unknown coefficients, eq. (8) is solved iteratively until Δ equals 0. It was found that a convergence of the solution of the least squares adjustment depended heavily on a good estimation of the coordinates of the origin. Moreover, to ensure a more stable solution, the coefficient b was not treated as an unknown. The adjustment is therefore repeated for a wide range of b-values (ranging from 0.1 to 5, with a step of 0.01), whereby the highest correlation coefficient between the observed and calculated profiles corresponds to the best b-value. That the method is unambiguous is shown in Fig. 14, where the correlation coefficient is given for the glacier profile J1 (Jenningsbreen) as a function of b. Except for the profile of Nipebreen, where b-values tend to exceed 5 and where the method failed, all other estimations of Sør Rondane cross profiles showed a comparable smooth correlation coefficient function as the one generated in Fig. 14.

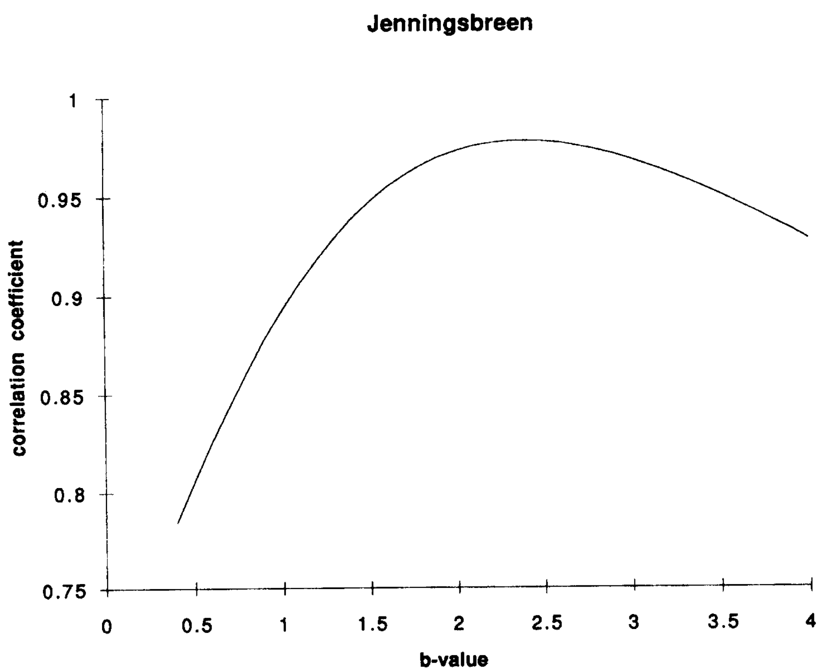


Fig. 14. *Calculation of the best fit of the power law curve to the valley cross profile of Jenningsbreen. The correlation coefficient is calculated for b-values ranging from 0.1 to 4.0 with a step of 0.01.*

5. Morphometric Analysis of the Sør Rondane Cross Profiles

Table 2 displays for all the Sør Rondane cross profiles the valley width, the central valley depth, the form ratio, the coefficient R and the correlation coefficient for the parabolic eq. (5), the coefficient b and the correlation coefficient for the power law eq. (6), and the shape factor f. All the Sør Rondane cross profiles used in this analysis are at present filled with ice. In a way this is unique, since most of the morphometric studies today are applied on deglaciated valleys, of which the cross-profiles are disturbed by post-glacial deposits and valley wall weathering under ice free conditions. Three cross profiles (BM1, J4 and MC) were omitted in the analysis, because of their low correlation coefficient when retreiving the b-value and their irregular shape of the valley walls.

The values of the form ratios of the valleys in the central Sør Rondane have a limited range showing rather constant values, since there is a tendency that deeper valleys become wider (Fig. 15), as explained in Pattyn *et al.* (1992). However, small form ratios are found at profiles GN and GC. These broader valley profiles (more than 10 km in width) are due to the confluence or junction with adjacent glaciers or tributaries (Mefjellbreen and Nipebreen).

It is observed that the trough head of the valley of Jenningsbreen (Table 2, profiles listed from south to north) is marked by a high value of b (1.53–2.38), pointing to a U-shaped valley, while low values (0.49–1.47), resulting in a V-shaped profile, characterize the main stream of the glacier. The same picture is found at Gjelbreen, with b-values of 4.02 at the trough head, low values in the central part (1.22), but then again increasing after the confluence of the adjacent tributaries

Table 2. Morphometric parameters of valley cross profiles in Sør Rondane: 2W=glacier width, D=glacier central depth, FR=form ratio, R=quadratic coefficient of the parabolic equation, r(R)=correlation coefficient of the parabolic equation, b=b-value of the power law equation, r(b)=correlation coefficient of the power law equation, f=shape factor.

Glacier	$2W$	D	FR	R(10-4)	$r(R)$	b	$r(b)$	f
J2	1793	503	0.28	6.55	0.90	1.53	0.91	0.29
J4	4549	646	0.14	0.93	0.79	0.44	0.86	0.31
J1	6366	1225	0.19	1.25	0.97	2.38	0.98	0.33
JC	7102	1651	0.23	0.98	0.77	0.49	0.93	0.22
J3	6146	1507	0.25	1.56	0.97	1.47	0.98	0.28
JE	8115	1819	0.22	1.06	0.98	1.45	0.98	0.28
G3	5300	894	0.17	1.16	0.87	4.02	0.96	0.35
GL	6326	1739	0.27	1.36	0.89	1.22	0.91	0.26
GC	11012	1685	0.15	0.54	0.93	1.81	0.93	0.30
GN	10940	1333	0.12	0.42	0.91	3.20	0.94	0.33
B1	2950	762	0.26	3.17	0.96	1.48	0.97	0.28
B3	2300	497	0.22	3.42	0.94	1.25	0.96	0.28
P1	2430	648	0.27	3.33	0.89	0.85	0.96	0.26
P2	2600	424	0.16	2.19	0.98	2.48	0.98	0.35
MC	2913	551	0.19	2.33	0.75	0.95	0.82	0.28
ME	4199	1566	0.37	2.99	0.85	0.81	0.93	0.22
MW	6259	1171	0.19	1.24	0.97	1.58	0.97	0.31
BM1	3800	410	0.11	0.85	0.81	1.21	0.86	0.33

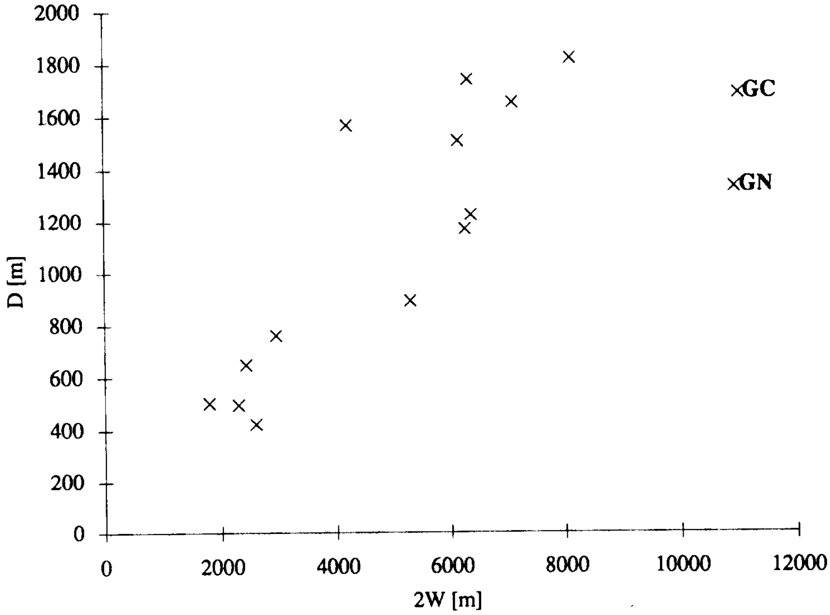


Fig. 15. Scattergram of valley width versus depth for Sør Rondane valley cross profiles.

Mefjellbreen and Nipebreen (1.81–3.20). A possible explanation for this phenomenon is that high erosion rates at the bottom of the ice fall, due to an increase in velocity and basal shear stress, and at the confluence area in the north, due to an increase of ice mass, account for a more profound valley development towards a U-shaped profile. However, as stated by AUGUSTINUS (1992) a lithological influence on the erosional process is not excluded: geological observations in the central Sør Rondane (ISHIZUKA and KOJIMA, 1987) show some differences in lithology and geologic structure of metamorphic rocks of the northern and southern part of the range. Along the boundary, they defined the Main Shear Zone (MSZ), running through the central part of Jenningsbreen and Gjelbreen, over the col at the northwest tip of Mefjell, continuing to the east through the central part of Mefjellbreen. Oddly, the four cross profiles that coincide with the MSZ, which is only 1 km wide (JC, GL, MC and ME) are characterized by very low b-values and hence shape factors, which could point to a high erosion resistance of the MSZ. Such an inference requires however more extensive field work.

In order to evaluate the inter-relationships between the morphometric parameters such as width, depth, form ratio, shape factor and the regression coefficients R and b, a correlation analysis was performed (Table 3). The highest correlation (0.89) is found between the shape factor and the b-values, indicating a positive trend (Fig. 16). Also a high correlation is found between the R-coefficient of the parabolic equation and the valley width. This is rather obvious, since the parabolic equation (and R in particular) can be written as a function of valley width (PATTYN *et al.*, 1992). A negative trend is observed between the b-values and the form ratio, though not so significant as the correlation between the b-values and shape factor. HIRANO and ANIYA (1988) found that such a '*b-FR*' diagram depicts the developmental process of

Table 3. Correlation matrix for the morphometric valley shape measures: n.s. = not significant at 90 per cent level, p=level of significance, 2W=glacier width, D=glacier central depth, FR=form ratio, R=quadratic coefficient of the parabolic equation, r(R)=correlation coefficient of the parabolic equation, b=b-value of the power law equation, f=shape factor.

	2*W*	*D*	*FR*	*R*	*r*(*R*)	*b*	*f*
2*W*	1.00						
D	0.77	1.00					
	$p < 0.001$						
FR	−0.53	0.06	1.00				
	$p < 0.05$	n.s.					
R	−0.80	−0.65	0.57	1.00			
	$p = 0.0003$	$p < 0.01$	$p < 0.05$				
r(*R*)	−0.03	−0.21	−0.32	−0.04	1.00		
	n.s.	n.s.	n.s.	n.s.			
b	0.25	−0.21	−0.69	−0.31	0.24	1.00	
	n.s.	n.s.	$p < 0.005$	n.s.	n.s.		
f	0.12	−0.39	−0.79	−0.22	0.54	0.89	1.00
	n.s.	n.s.	$p = 0.0005$	n.s.	$p < 0.05$	$p = 0.0001$	

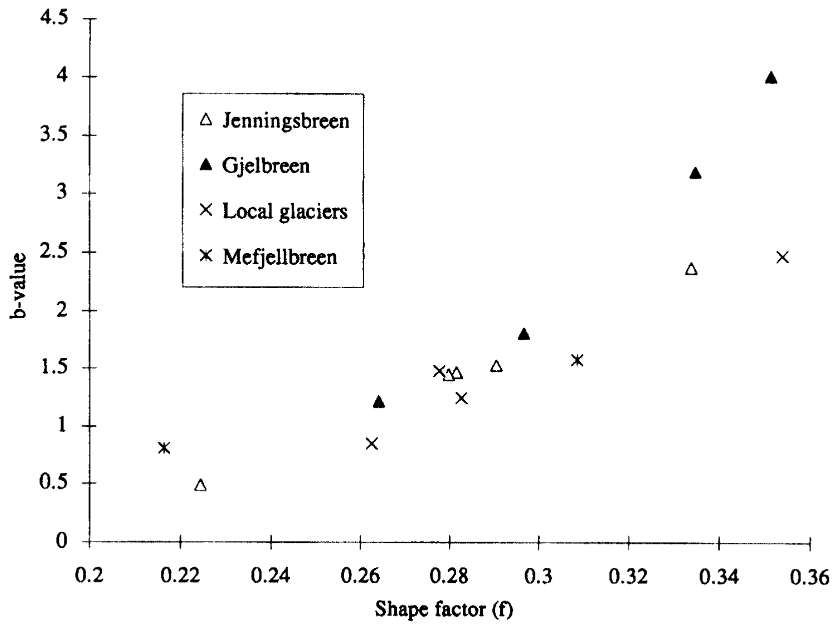


Fig. 16. Scattergram of shape factor versus b-value of the power law equation for Sør Rondane valley cross profiles.

glaciated valley morphology, *i.e.* the succession from V-valley to U-valley. From valley cross profiles in the Canadian Rockies, Patagonia and Antarctica they suggest two types of cross-profile development of the glacial valley. One type of the development (Rocky Mountain Model, RMM) is from a shallow, wide V-shaped valley to a deep U-shaped valley, and another type (Patagonia-Antarctica Model, PAM) is from a rather steep and narrow V-shaped to a wide, broad U-shaped valley.

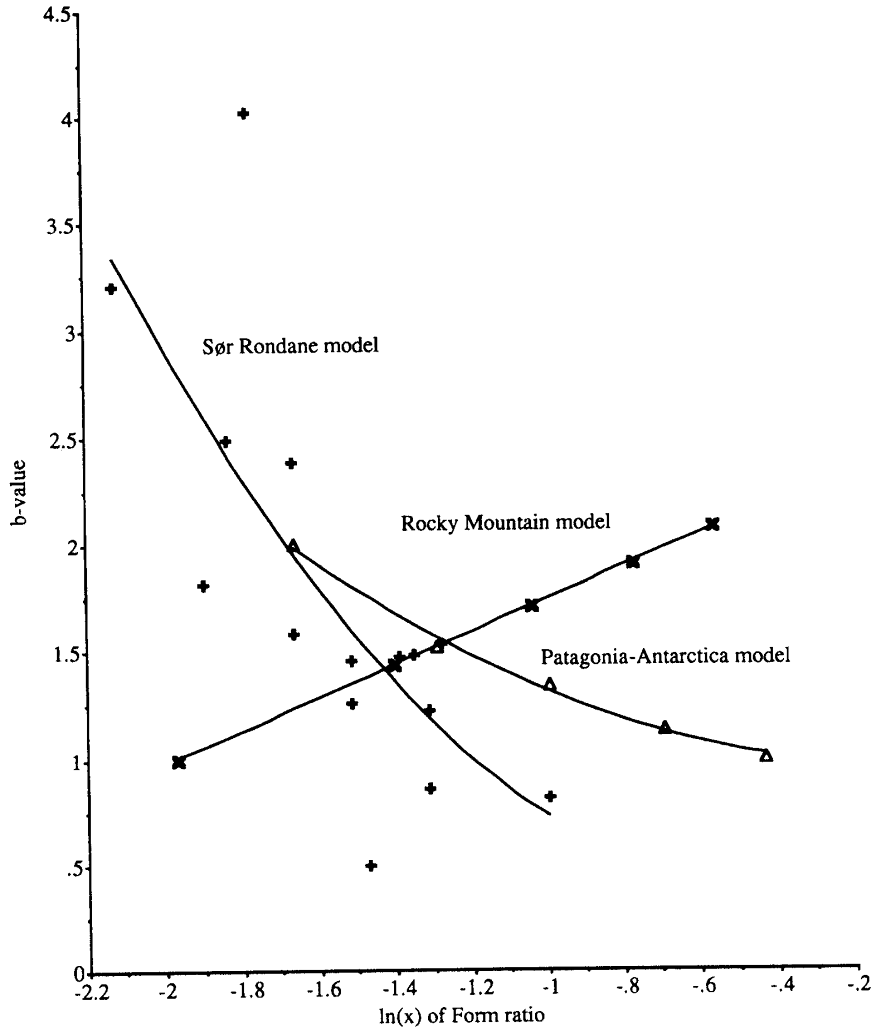


Fig. 17. Comparison of the 'FR-b'-diagram of Sør Rondane valley cross profiles with the Rocky Mountain model (RMM) and the Patagonia-Antarctica model (PAM) of HIRANO *and* ANIYA *(1988).*

The Rocky Mountain model depicts a deepening development of the glacial valley, while the Patagonia-Antarctica model portrays a widening, rather than deepening, process of glacial development and hence suggesting a different mechanism of glacier erosion from the alpine type. Whether this '*b-FR*' diagram represents the form development of glacial cross profiles is a subject of much debate. According to HARBOR (1990) such an analysis is hindered by the fact that one cannot observe the evolution of a single cross profile (from V to U) over time, so that the spatial variations in form are linked to some surrogate measure over time or extent of glacial erosion. Nevertheless, morphometric data from the Southern Alps in New Zealand (AUGUSTINUS, 1992) confirm the 'alpine type' Rocky Mountain model of HIRANO and ANIYA (1988). Also, the morphometric data of the glacier profiles in Sør Rondane confirm more or less the Patagonia-Antarctica model (Fig. 17). Thus, bearing in mind Harbor's critic, the '*b-FR*' diagram for the Sør Rondane glacier profiles can be interpreted as follows: (i) V-shaped valleys tend to be smaller in width than U-shaped valleys, and (ii) considering the overdeepened bedrock profile of the outlet glaciers, the valley width of the pronounced U-shaped valleys is remarkably high.

6. Conclusions

Based on gravimetric surveys and radio echo sounding measurements carried out in the Sør Rondane Mountains, a map of the subglacial relief is compiled, emphasizing an overdeepening of the central outlet glaciers, with bedrock elevation well below sea level. The data of the cross profile measurements was further used for a morphometric analysis, thereby calculating the form ratio, shape factor and the spectral coefficients of the power law and parabolic equation. The use of a robust power law, with respect to the datum, allowed for a more accurate determination of the power law coefficient b, which describes the valley cross profile form. Nevertheless, morphological data alone are insufficient to assess a morphological development over time. HARBOR (1990) argues that in order to understand the evolution of glacial cross profiles, the attention should also focus on (i) the flow pattern through the glacier cross section, (ii) the glacial erosion process, (iii) the pattern of bedrock resistance to erosion, (iv) the evolution in valley slopes above the glacier, and (v) the temporal variation in ice occupation. Since the major outlet glaciers in the central Sør Rondane, *i.e.* Jenningsbreen and Gjelbreen, are characterized by a complex morphological evolution from the trough head to the mountain exit area, the control of ice dynamics and bedrock weathering resistance cannot be neglected, as shown by the morphometric characteristics of the valleys after confluence with adjacent tributaries and by the valley morphology coinciding with the Main Shear Zone (MSZ). However, a detailed analysis of the ice-bedrock relationship (glacial erosion) and extensive field work is necessary to arrive at more solid conclusions. Following the analysis of HIRANO and ANIYA (1988), the relationship between b-values and the form ratio is in overall agreement with their proposed Patagonia-Antarctica model.

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