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UPLIFT OF THE SØR RONDANE MOUNTAINS, EAST ANTARCTICA

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Abstract: The Sør Rondane Mountains, East Antarctica, originated from a marginal swell formed during the breakup of Gondwanaland, and the area is now in isostatic equilibrium. Altitude of the mountains is estimated to be controlled by the isostatic rebound associated with glacial denudation and ice loading. The denudational depth and extent of ice load subsidence were inferred from topographic cross sections of subglacial topography. The total mean erosion depth during 30 Ma since the preglaciation was approximated 1.3 km with an associated rebound of 1.1 km. As a result of the uplift, the summit altitude in the preglacial time is estimated to have been 2.1 km above the present sea level. The average rates of the denudation and surface uplift are 43 m/Ma and 37 m/Ma, respectively.

1. Introduction

Geomorphological evolution of the Sør Rondane Mountains has been interpreted from field surveys that were conducted between 1984-1991 (MORIWAKI *et al.*, 1991; MORIWAKI and HIRAKAWA, 1992). The reconstruction of the past ice sheet surfaces in different ages was attempted. These interpretations were based on glacial landforms from several high altitude areas above the present glacier surface. In Stage 5 (the oldest glaciation prior to 4 Ma), the Sør Rondane Mountains except for several high peaks were covered by an ice sheet (MORIWAKI and HIRAKAWA, 1992). If one considers that the past mountain altitudes approximated to the present one, the past ice sheet should have had an extremely large volume. On the other hand, recent studies on Antarctic ice sheet history suggest that the Pliocene East Antarctic ice sheet had been considerably smaller than the present ice sheet (MCKELVEY *et al.*, 1991).

To clarify the uplift of the Sør Rondane Mountains is important subject not only for studies of the landform history of the specific mountains, but also for studies of Antarctic ice sheet history. Although field data are very limited at present, uplift rates can be estimated to interpret geomorphology of the Sør Rondane Mountains. Uplift of mountains in Antarctica has been discussed in the Transantarctic Mountains, Prince Charles Mountains, and coastal highlands between 50°-64°E (TINGEY, 1985; WELLMAN, 1983, 1988; WELLMAN and TINGEY, 1981). These analogous situations provide constraints for interpreting the geomorphology of the Sør Rondane Mountains. In this study, the past altitudes of the Sør Rondane Mountains will be estimated based on the extent of denudational uplift since the preglacial time. This study will provide a starting-point for future studies of mountain uplift and its role in glacial history of the inland mountains in Eastern Queen Maud Land.

2. Study Area

The Sør Rondane Mountains form an eastern part of coastal mountain ranges, which extend intermittently over 2000 km in an east-west direction of Dronning Maud Land, East Antarctica. The Sør Rondane Mountains are located about 200 km inland from the coast (Fig. 1) and form a barrier damming the flow of inland ice, so that the region is on the steep margin of the Antarctic ice sheet, with ice altitudes decreasing from about 3 km in the south to the sea level within a 400 km distance. The mountains are separated by outlet-glaciers into several massifs composed of Precambrian and Palaeozoic metamorphic and granitic rocks. The coastal lowland on the north of the mountains is covered with the ice sheet being nourished by outlet glaciers from the southern ice plateau. According to the result of radio echo sounding (NISHIO and URATSUKA, 1991: Fig. 4-AB), moderate relief seems to have been shaped by areal scouring in the south and selective linear erosion in the north under the ice sheet.

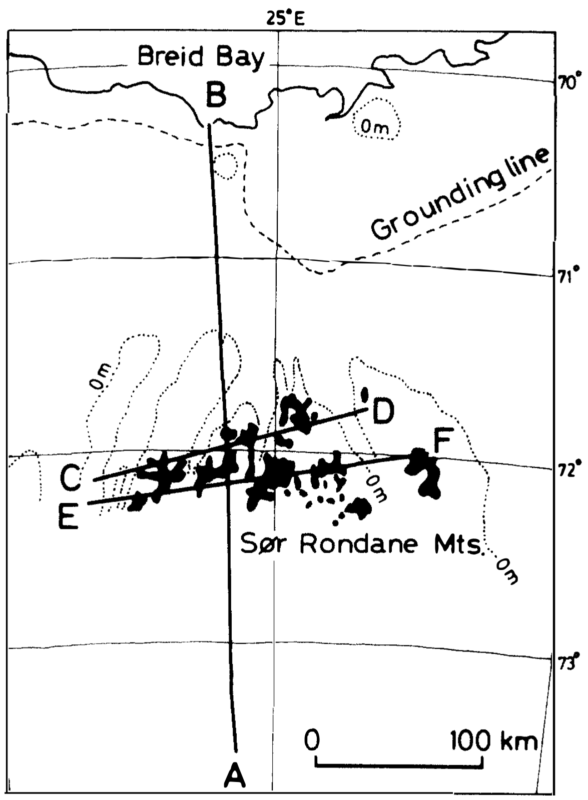

Fig. 1. Sør Rondane Mountains and the study area. Grounding line and 0-m contour indicate the bedrock topography (NISHIO and URATSUKA, 1991). Straight solid lines show the flight course of ice thickness measurements (A-B) and topographic cross sections (C-D and E-F).

3. Controlling Factors Determining the Altitude of the Mountains

Previous studies on geology, geomorphology, and geophysics of the Sør Rondane Mountains suggest the following three scenarios:

1) The coastal mountains in East Antarctica, including the Sør Rondane Mountains, are originated from marginal swells formed during the breakup of Gondwanaland in the Cretaceous (WELLMAN, 1983; YOSHIDA, 1991; OLLIER, 1991). Long and narrow highlands or plateaus with steep margins (Great Escarpment called by OLLIER, 1991) have existed since the time of breakup. The top surface of the plateau seems to be a past planation surface of Gondwanaland, and the escarpment is a remnant of the continental margin.

2) There is no evidence of both Mesozoic and Cenozoic orogenic movement of the Sør Rondane Mountains, in other words the Precambrian bedrocks underlying the region have not been subsequently reheated. In such Precambrian crustal areas, where the lithospheric strength is relatively high, WELLMAN (1988) indicated that the crust is in isostatic equilibrium for 300×300 km areas. In WELLMAN's context, the area including the Sør Rondane Mountains shown in Fig. 1 is now in isostatic equilibrium. Therefore, possible vertical movement of the land surface in the area depends upon both denudational uplift and subsidence by ice load.

3) Glacial denudation on gentle mountain slopes is thought to have started with the formation of the East Antarctic ice sheet in the middle Oligocene, about 30 Ma (WEBB, 1991). In the Miocene and the early Pliocene, strong glacial erosion occurred extensively on relatively low-relief lands. Subsequently, in the late Pliocene glacial erosion shifted to selective linear erosion (MORIWAKI and HIRAKAWA, 1992). The mountains were cut by outlet glaciers, creating intervening massifs separated by the troughs. In the Pleistocene, this glacial erosion was limited to deep troughs (MORIWAKI and HIRAKAWA, 1992).

As mentioned above, controlling factors of the isostatic movement of the Sør Rondane Mountains are limited. Assuming that the ice volume has not fluctuated since the ice sheet formation, a simple model can be applied to explain the relationship between preglacial and present altitudes. Altitude of preglacial land surface

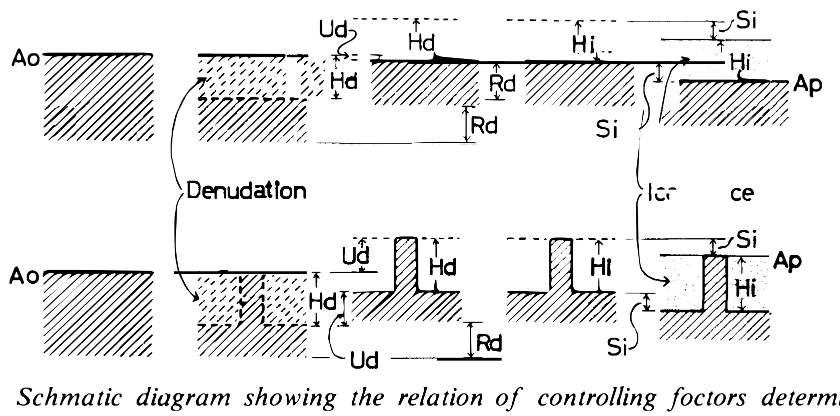

Fig. 2. Schmatic diagram showing the relation of controlling foctors determining the isostatic movements : $A_o = A_p - U_p + S_i$. Above: flat area, Below: mountain area. (See also the text.)

(A_0) is given by

$$A_0 = A_p - U_d + S_i,$$

where A_p is the present altitude of mountains, U_d is an amount of denudation uplift of surface, and S_i is an amount of subsidence of surface by ice sheet load. This relation is indicated in Fig. 2.

4. Results

4.1. Present altitude of the mountains or ground surface (A_p)

The highest peak of the mountains is nearly 3.0 km (2993 m) a.s.l. and most summits range from 1.5 to 2.8 km a.s.l. At the northern margin, mountains were severely eroded by outlet glaciers into spiky towers and small nunataks. While in the southern part, there are gentle summits which form concordant flat summit levels ranging from 2.6 to 2.8 km a.s.l. (Fig. 3). According to the geomorphological observations (*e.g.* MORIWAKI and HIRAKAWA, 1992), these gentle summits were eroded by local mountain glaciers, suggesting that the original planation surface (past Gondwana surface) is scarce.

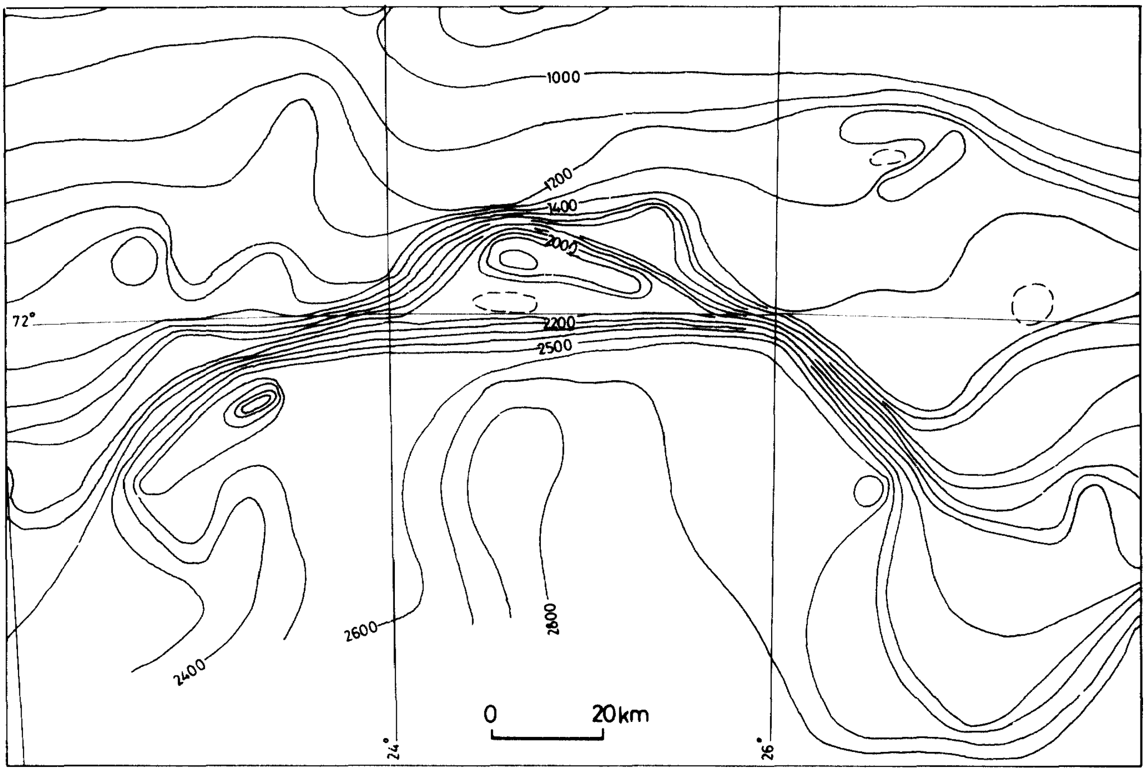

Fig. 3. Summit level contours in the area around the Sør Rondane Mountains, mapping highest point altitudes in each 15 × 15 km grid superimposed on topographic maps of 1 : 50000 scale (Geographical Survey Institute, Japan, 1989-1992) for the central part and 1 : 250000 (NORSK POLARINSTITUTT, 1988). Contour interval : 100 m, broken line shows depressed area. The area includes both ground and ice surface.

4.2. Subglacial and preglacial topography

To assess the total amount of denudation, information about subglacial topography or ice thickness is necessary. Unfortunately, information is restricted at present. An airborne radio echo sounding profile was obtained in a north-south direction crossing the mountains (NISHIO and URATSUKA, 1991: Fig. 4-AB). The cross-section shows typical features of the Great Escarpment on the margins of the past Gondwanaland (OLLIER, 1991). Gravity measurements revealed the subglacial topography along cross sections in many glacier troughs (VAN AUTENBOER and DECLEIR, 1978; DECLEIR *et al.*, 1989). Figure 4-CD and -EF sections display the estimated subglacial topography along two profiles parallel to the mountains axis compiled from results of these gravity measurements. Glacial troughs were deepened below sea level in the northern part of the mountains.

Preglacial denudation is assumed to have occurred by subaerial processes, although no information has been obtained on degree of erosion and relief. In the Prince Charles Mountains, however, WELLMAN and TINGEY (1981) suggests that the preglacial surface preserved by the Eocene lava flow shows very flat landforms. On

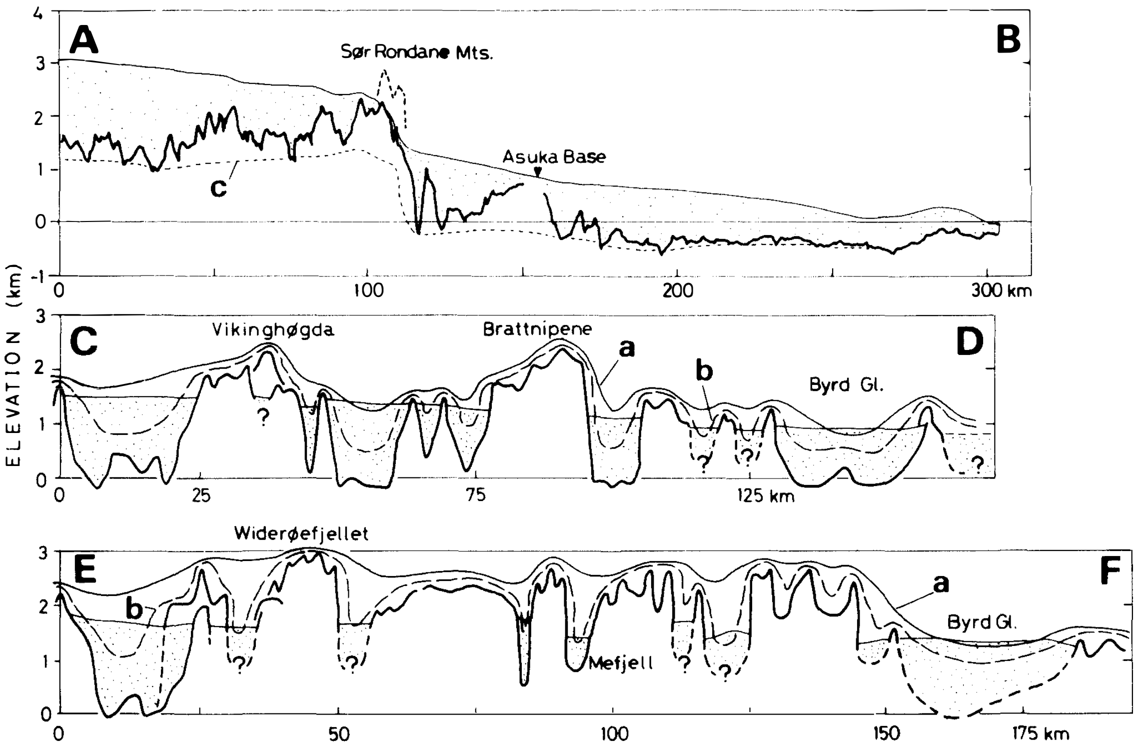

Fig. 4. Cross sections of glacier surface, subglacial topography, and inferred past topography. a : Preglacial topography, b : Topography in Stage 4 (MORIWAKI and HIRAKAWA, 1992), c : Assumed present lowest rock surface or glacial trough bottom. The A-B section shows the results of airborne radio echo sounding along the flight line between Breid Bay and the interior region (NISHIO and URATSUKA, 1991), and the ice thickness data in glacial troughs in C-D and E-F sections have been obtained from VAN AUTENBOER and DECLEIR (1978) and DECLEIR et al. (1989). The location of these sections are indicated in Fig. 1.

the other hand, WELLMAN (1983) shows that the preglacial relief ranged between 0.5-1.0 km in the southern highland to the south of the Napier Mountains. The present landforms of the Sør Rondane Mountains resemble the southern highland of the Napier Mountains, so that the WELLMAN's (1983) values are adopted to the preglacial relief of the Sør Rondane Mountains. Relatively high relief (0.5-1.0 km) in preglacial time was drawn as shown in Fig. 4. The curve b in Fig. 4 was drawn based on the geomorphological survey by MORIWAKI and HIRAKAWA (1992), and corresponds to the Stage 4 in MORIWAKI and HIRAKAWA (1992), which is assumed to be in the Miocene or the early Pliocene.

4.3. *Amount of denudational uplift of surface* (U_d)

Isostatic rebound resulted from denudation, that is denudational uplift of the crust, can be roughly estimated by the density contrast between the crust rocks and the underlying upper mantle. Rebound amount (R_d) is given by

$$R_d = H_d \ (d_c/d_m),$$

where H_d is the amount of denudation, d_m is the mantle density of 3.3 tm^{-3}, and d_c is the mean crustal density of 2.9 tm^{-3} (WELLMAN, 1988). The equation indicates that the rebound amount is 0.88 times the amount of denudation ($R_d = 0.88\ H_d$). However, the actual surface altitude caused by isostatic rebound, that is the amount of denudational uplift of actual surface, does not always coincide with the value of the rebound amount. For example, a flat surface eroded uniformly will be uplifted until it becomes to slightly lower level than the former level ($\times 0.88$) after denudation. On the other hand, in deeply dissected mountain areas, where denudation occurred mainly in valleys, peaks and divides rise above the initial surface level as HOLMES (1978, p. 373) illustrated. This difference is schematically indicated in Fig. 2.

The amount of denudation of the mountains was measured on the two east-west profiles indicated in Fig. 4 mean erosion depth was calculated from the area enclosed by estimated past profiles and present subglacial profiles (Table 1). Massifs and nunataks in the mountains represent crag-and-tail topography and streamlined plan shapes resulted from erosion by outlet glaciers flowing in a north-south direction. This topography means the northern and southern parts of the massif were eroded more largely than the mid parts. Therefore, for values of erosion depth of the entire massif, a half amount of value of erosion depth was added to the amount measured on the east-west profiles. The total mean erosion depth since the preglacial time appears to be around 1.3 km and the rebound amount attains over 1.1 km. The rebound caused also an equal amount of uplift to the high summits and ridges where the glacial erosion is thought to be relatively small.

Although there is no information on the denudational amount of the coastal lowland on the north of the mountains and of the inland plateau on the south, the present subglacial landforms suggest the mode and intensity of past glacial processes. The subgalcial topography of the coastal lowland shows relatively low relief (Fig. 4-AB) which is likely to be subjected to severe areal scouring. On the inland plateau, the terrain has been shaped by areal scouring in the south and selective linear erosion

Table 1. Estimated mean denudation depth and mean ice thickness measured in the Fig. 3.

	Denudation depth			Mean ice thickness	Subsidence
	a-b	b-present	a-present		
	(km)	(km)	(km)	(km)	(km)
C-D section	0.30	0.52	0.82	0.86	—
S-N addition	0.15	0.26	0.41	0.43	—
Total	0.45	0.78	1.23	1.29	0.39
E-F section	0.36	0.54	0.90	0.90	—
S-N addition	0.18	0.27	0.45	0.45	—
Total	0.54	0.81	1.35	1.35	0.40
Mean	0.5	0.8	1.3	1.3	0.4

a: Preglaciation, b: Stage 4 (Moriwaki and Hirakawa, 1992).

in the north under the ice sheet. The mean erosion depth by areal scouring in the coastal lowland is considerably larger than that on the inland plateau, because some parts of the inland ice sheet seem to be under the cold base condition (Moriwaki and Hirakawa, 1992). Because no estimated value of denudation depth has been obtained in these areas, the tentative values are adopted. For the coastal lowland, where intensive erosion occurred, the mean denudation depth is thought to have been 1.3 km, which is an equal value of that in the mountains. On the inland plateau, moderate erosion occurred on the order of 0.6 km, which is about a half value of that in the mountains (Table 2). Corresponding isostatic rebounds might be occurred, but the amount of surface uplift might differ according to the difference of the manner of glacial erosion and resultant landforms. In the coastal lowland, uplift amount of the higher surface is thought to be slightly larger than the calculated values because of its unevenness, then 0.0 km is adopted. In the inland plateau area, the uplift amount of higher places was estimated at a half value of the denudation depth (0.3 km), judging from the relatively moderate relief of the area (Table 2).

Table 2. Geomorphological characteristics associated with isostatic movement of the Sør Rondane Mountains and their vicinity.

	Sør Rondane Mts.		Inland plateau	Coastal lowland
	Preglacial	Stage 4	Preglacial	
	(km)	(km)	(km)	(km)
Present altitude (A_p)	2.8	2.8	1.5	−0.3
Mean denudational depth	1.3*	0.8*	0.6	1.3
Isostatic rebound (R_d)	1.1	0.7	0.5	1.1
Surface uplift by denudation (U_d)	1.1	0.7	0.3	0.0
Ice load subsidence (S_i)	0.4	0.0	0.6	0.4
Previous altitude (A_0)	2.1	2.1	1.8	0.1

* From Table 1.

4.5. Subsidence by ice loading (S_i)

The effect of ice loading can be estimated by the same way as the isostatic rebound. Subsidence of the regional mean rock surface by ice loading (S_i) is given by

$$S_i = H_i \ (d_i / d_{mc}),$$

where H_i is mean ice thickness, d_i is the ice density of 1.0 tm^{-3}, and d_{mc} is the mean crustal and mantle density of 3.1 tm^{-3}. Amount of subsidence is about 0.3 times the ice thickness ($S_i = 0.32\ H_i$).

The present ice thickness is shown in Fig. 4-AB. It is 1.5-2.0 km in the inland plateau and 1.0-1.5 km in the coastal lowland. This difference of ice depth may be reflected in the effect of ice loading : the amounts of estimated subsidence are 0.5-0.7 km on the inland plateau, and 0.3-0.5 km in the coastal lowland. In the mountains, ice depth was measured on the Fig. 4 (C-D, E-F sections) in the same way as the amount of denudation; mean ice depth was calculated from the dotted areas. The mean ice thickness is around 1.3 km, therefore the subsidence amount is nearly 0.4 km (Table 1).

4.6. Altitudes of the preglacial surface (A_0)

From the above arguments values of present altitude, denudational surface uplift, and ice load subsidence were obtained. Altitudes of the preglacial surface have been calculated from these values (Table 2). The past summit altitude of the Sør Rondane Mountains in the preglacial time is 2.1 km, and that in the stage 4 in the Miocene or Pliocene is also 2.1 km. Surface altitudes of the inland plateau and the coastal lowland in preglaciation are 1.8 km and 0.1 km (near sea level), respectively (Table 2).

5. Discussion

5.1. Rates of uplift

From the above-mentioned results, a model of "mountain building" can be illustrated. In short, the crust in the whole study area rose by isostatic rebound due to glacial denudation. While, the uplift of the ground surface was little to nothing in the coastal lowland where areal denudation occurred, and a moderate surface rise occurred on the inland plateau, where selective linear erosion and partly areal denudation occurred. On the other hand, the relatively large uplift occurred in the mountains where the summit areas were remained by differential glacial erosion. Similar models were proposed for some mountains in East Antarctica (WELLMAN, 1983; WELLMAN and TINGEY, 1981).

As mentioned previously, glacial erosion in the area probably started in the middle Oligocene, at about 30 Ma. The average rates of the denudation and surface uplift inferred for the Sør Rondane Mountains since the onset of glaciation is 1.3 km and 1.1 km during 30 Ma, or 43 m/Ma and 37 m/Ma, respectively. These are comparable to those obtained in the Prince Charles Mountains (WELLMAN and TINGEY, 1981) and in the Napier Mountains (WELLMAN, 1983).

5.2. Geomorphic evolution

SUGDEN and JOHN (1976) illustrates a model showing the evolution of an ice sheet and glacial landforms in a plateau near the periphery of a continent. This model, however, does not employ the isostatic crustal movement. Preferred geomorphological evolution of the Sør Rondane Mountains is shown in Fig. 5 which indicates the ice sheet development, geomorphic evolution, and isostatic movements of the area. The summit surface of the Great Escarpment was 2.1 km in the preglacial time (Fig. 5-A), and local ice caps and mountain glaciers covered the top in the early Oligocene (Fig. 5-B). Then the area was completely covered with the East Antarctic ice sheet in the middle Miocene. This enormous ice load isostatically depressed the rock surface, and the ice sheet caused denudation, accompanied subsequently by isostatic rebound. According to MORIWAKI and HIRAKAWA (1992), both extensive erosion and deep dissection occurred by a warm-based ice sheet in the mountains during the Stage 4 and 5 (Fig. 5-C). This caused in some summits to rise above the ice sheet and to shape to spiky towers as shown at present (Fig. 5-D).

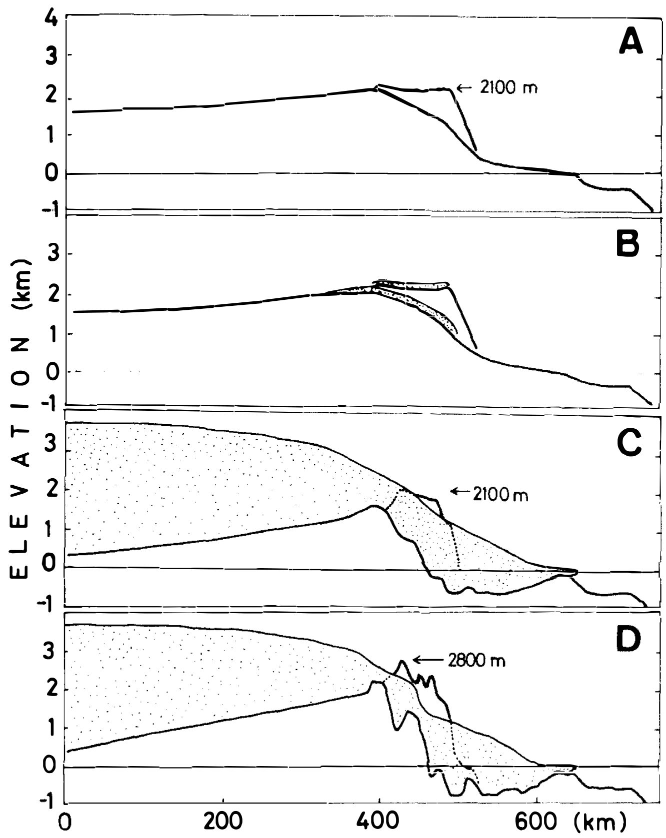

Fig. 5. Diagramatic cross sections of Sør Rondane Mountains showing the geomorphological history. A : Preglaciation time, Eocene (before 40 Ma), B : Onset of glaciation, Oligocene (before 30 Ma), C : Stages 4 or 5, late Miocene or Pliocene (5-10 Ma), D : Present day.

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