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Features of ice sheet flow in East Dronning Maud Land, East Antarctica

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Abstract: The Japanese Antarctic Research Expeditions (JAREs) have done glaciological studies on ice sheet dynamics and surface mass balance in East Dronning Maud Land, mainly around the Shirase Glacier drainage basin, during more than 30 years. The surface mass balance, obtained mainly by the snow stake method, was more than 250 mm/a in the coastal region, less than 50 mm/a in the inland region higher than 3500 m in altitude, and about 100 mm/a on average in the five drainage basins in East Dronning Maud Land. The ice flow velocity was observed around East Dronning Maud Land in three observation periods: on a route transversal to the Shirase Glacier flow in 1969 to 1974, along a route longitudinal to Shirase Glacier and a transversal route from Mizuho Station (70°42′S, 44°17′E, 2250 m a.s.l.) to the Sør Rondane Mountains area in 1982 to 1987, and along a route from S16 (69°02′S, 40°03′E, 554 m a.s.l.) near the coast to Dome Fuji Station (77°19′S, 39°42′E, 3810 m a.s.l.) in 1992 to 1995. Assuming steady ice flow, the balance velocity is calculated by integrating the surface mass balance in the upstream area from a specific point to the flow origin between adjacent stream lines. From the relation between balance velocity and basal shear stress, the basal sliding area was specified.

key words: ice sheet flow, balance velocity, surface mass balance, basal sliding

1. Introduction

The Japanese Antarctic Research Expeditions (JAREs) have performed continuous glaciological studies on the surface mass balance and the ice sheet dynamics along numerous traverse routes in East Dronning Maud Land over more than 30 years (Fig. 1). The main purpose of these studies is to elucidate the features of the surface mass balance and the behavior of the ice sheet flow corresponding to the ice dynamics of Shirase Glacier. The measurements of the ice sheet dynamics were done in three periods, as described below.

First, in the Glaciological Research Program on Mizuho Plateau (1969–1975), a 250 km long triangle chain was set for ice flow measurement along a route transverse to Shirase Glacier from the Yamato Mountains to the central part of the Shirase drainage basin at 72°S during 1969 to 1974 (Naruse, 1978a, 1979). In this measurement, thinning of the ice sheet by 70 cm/a was observed (Mae, 1977; Mae and Naruse, 1978).

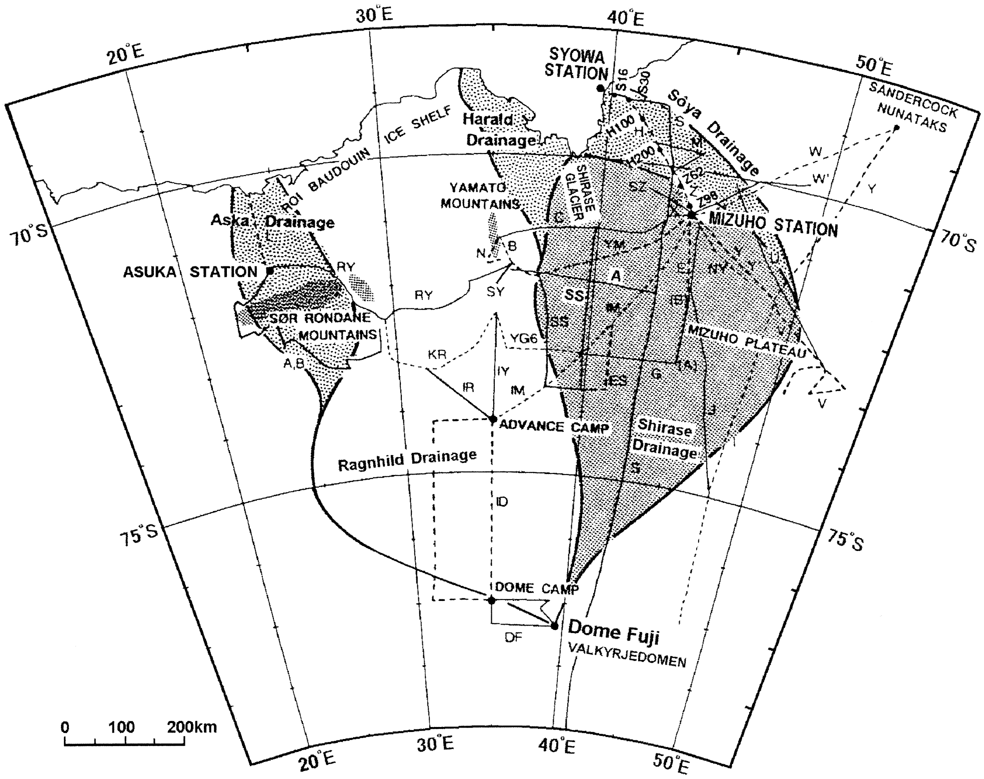

Fig. 1. Observation area and traverse routes in East Dronning Maud Land.

Second, in the East Queen Maud Land Project (1982–1986) the study area was expanded from 25°E near Sør Rondane to 45°E around Mizuho Station (70°42′S, 44°17′E, 2230 m a.s.l.). Survey points were set along a longitudinal route on Shirase Glacier and a transversal route from Mizuho Station to Asuka Station (71°31′S, 24°08′E, 930 m a.s.l.) near the Sør Rondane Mountains area, and their location was positioned twice for the ice flow velocity by the satellite doppler positioning system of NNSS (Navy Navigation Satellite System) (Nishio *et al.,* 1989). In this extended area the main survey objects were measurements of ice flow, surface and sub-glacial topography, and medium depth ice coring at Mizuho Station.

During the Deep Ice Coring Project at Dome Fuji (1991–1997), during which a deep ice core was obtained of 2500 m at Dome Fuji Station, the surface flow velocities were measured by GPS (Global Positioning System) along the 1000 km route from S16 (69°02′S, 40°03′E, 554 m a.s.l.) to Dome Fuji Station (77°19′S, 39°42′E, 3810 m a.s.l.) through Mizuho Station.

2. Surface mass balance

Japanese Antarctic Research Expeditions (JAREs) have observed the surface mass

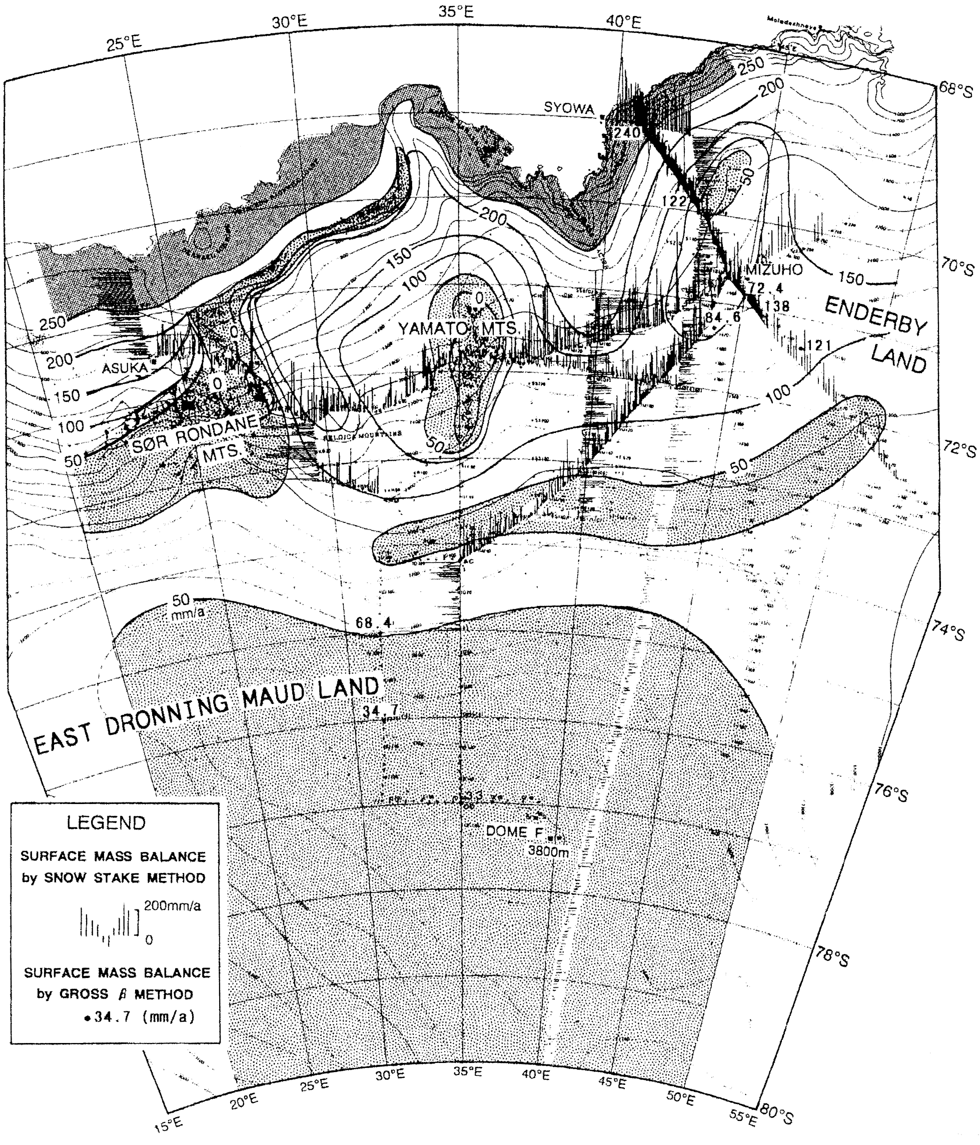

Fig. 2. Surface mass balance in East Dronning Maud Land.
Solid lines are isolines of surface mass balance (mm a^{-1} water equivalent). Bars show snow-surface mass balance measured by the snow stake method. Values at solid circles are surface mass balance obtained from Gross Beta activity and tritium profiles (Takahashi et al., 1994).

balance along many traverse routes in the area from West Enderby Land to East Dronning Maud Land, East Antarctica, since 1968. Along the route between Syowa Station (69°00′S, 39°35′E, 20 m a.s.l.) and Mizuho Station, the surface mass balance has been annually measured by snow stakes since 1968. During the Glaciological Research Program on Mizuho Plateau from 1969 to 1975, surface mass balance observations were carried out from the Yamato Mountains area in the west to the Sandercock Nunataks in the east. In the Glaciological Research Program in East Dronning Maud Land from 1982 to 1987, the observations were carried out over a wider area from the Sør Rondane Mountains area to Mizuho Plateau including the highest point in East Dronning Maud Land, Dome F (77°22′S, 39°37′E, 3810 m a.s.l.). All these surface mass balance data have been compiled to obtain the distribution of surface mass balance in East Dronning Maud Land as shown in Fig. 2.

The surface mass balance generally decreased with distance from the coast, which was more than 250 mm/a in the coastal region and less than 50 mm/a in the inland region above 3500 m in altitude. At Mizuho Station (2230 m a.s.l.) the components of surface mass balance were studied, where the sublimation was about 50 mm/a, precipitation was between 140–260 mm/a, and the mass removal from the surface by drifting snow was estimated as about 100 mm/a, which agrees with the surface mass balance estimated at 70 mm/a from the grain growth rate (Takahashi *et al.,* 1994).

Around the mountainous area, the surface mass balance was small and in extreme cases negative. In the Yamato Mountains area (71°10′–72°50′S, 35°–36°E, 1600–2500 m a.s.l.), a large extent of bare ice is exposed, where the surface mass balance is largely negative due to sublimation. The cause of this bare ice field is explained by mass removal caused by increasing katabatic wind in the relatively steep surface area (Takahashi *et al.,* 1988). The same effect is seen in the inland area between 3000 m and 3200 m a.s.l., where the surface mass balance was relatively small, less than 50 mm/a, compared to the surrounding areas. In this area the katabatic wind is strong, and redistribution of snow occurs by the acceleration of drifting snow (Takahashi and Watanabe, 1997).

3. Measurements of ice sheet flow

3.1. Triangle chain survey

Surveys of a triangulation chain were carried out in 1969 and 1973–1974 (Naruse, 1978a, b, 1979). The triangle chain was set along Route A from the Yamato Mountains to the central part of the Shirase drainage basin at a distance of 250 km along 72°S; it was composed of 162 triangles by 164 stations. The first two stations were fixed on ice-free rock near the Yamato Mountains as a base line. The length of a triangle side was about 3 km. The survey was principally conducted by angle measurements of the triangle chain with Wild T2 theodolites. From the difference of position in the two surveys in 1969/70 and 1973/74, horizontal and vertical components of surface velocities were measured at 140 stations.

3.2. NNSS survey

From 1982 to 1987, the observations of surface flow velocities and strain rates were done mainly along a flow line of Shirase Glacier on 39.5°E (Route SS) and along a traverse route parallel to the 2000 m contour line from Mizuho Station to the Belgica Mountains *via* the Yamato Mountains (Nishio *et al.,* 1989). Flow velocities were

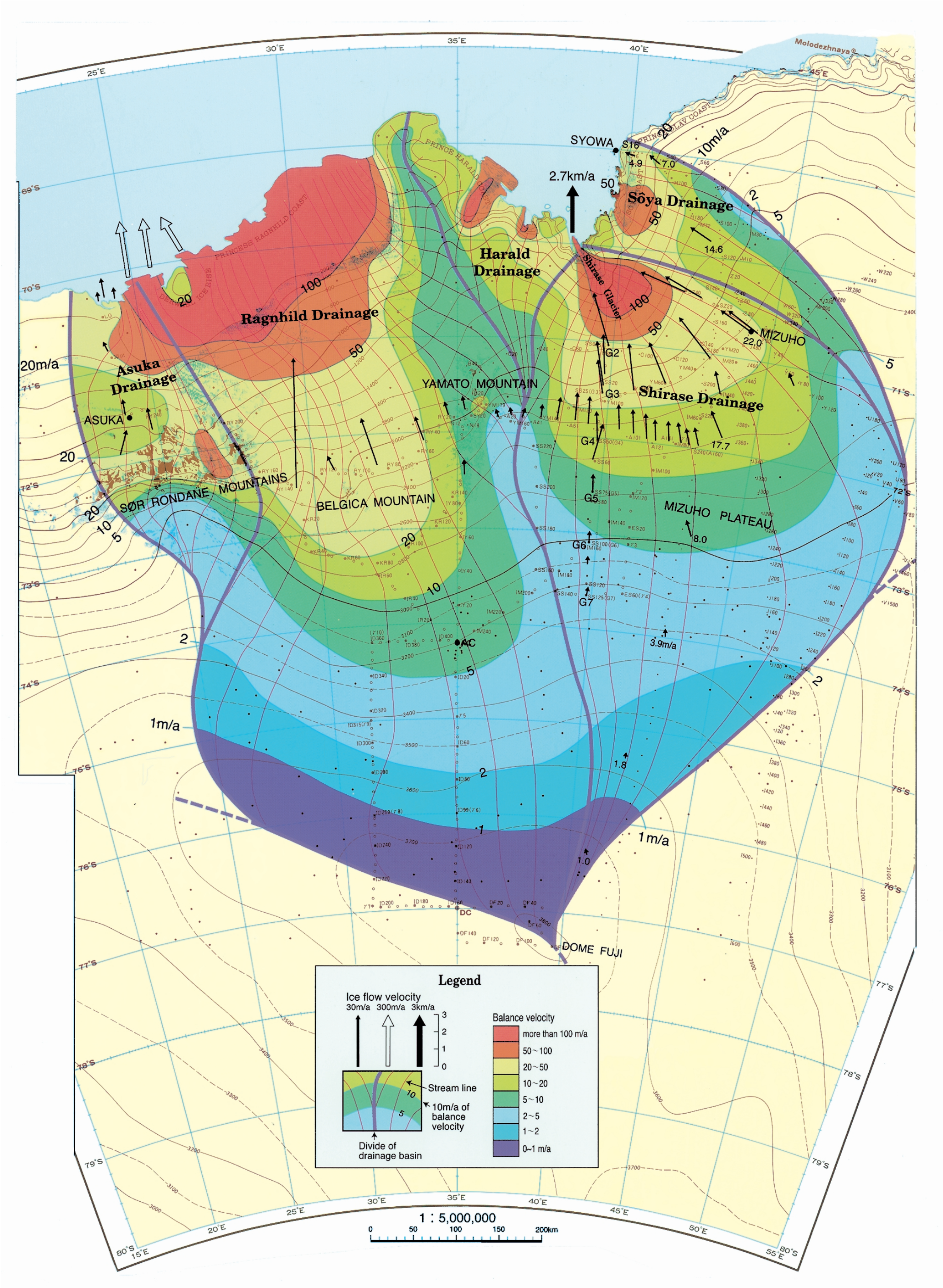

Fig. 3. Ice sheet flow velocity and distribution of balance velocity.
Observed surface velocities are shown by arrows. Calculated balance velocities are presented by isolines and classified by color (modified from Takahashi et al., 1997).

obtained from the difference between two geodetic surveys (the first survey in 1982–1984, the second survey in 1986–1987) using a NNSS (Navy Navigation Satellite System) satellite doppler positioning system.

3.3. GPS survey

In 1992 to 1995, in the Deep Drilling Project on Dome Fuji, surface flow velocities were measured by differential geodetic positioning using a GPS along a traverse route from S16 to Dome Fuji Station at a distance of about 1000 km. On this route, ice radar observations were carried out to determine ice thickness and to analyze internal layering (Takahashi *et al.,* 1997).

4. Observed ice flow velocity

In Fig. 3, the observed surface flow velocities of the ice sheet are plotted. Generally the surface flow velocity was larger in the coastal region and smaller inland. The flow velocity at the mouth of Shirase Glacier is very large, 2.4–2.7 km/a (Nakawo *et al.,* 1978; Fujii, 1981), while the velocities at the coast around the Derwael Ice Rise on the western side of the Ragnhild Drainage are around 300 m/a. The flow of Shirase Glacier is typical for an ice stream, while the ice flow around the Roi Baudouin ice shelf is a typical ice sheet flow where large icebergs are produced from the ice shelf. These two typical flows are common in Antarctica. On the ice sheet, the surface flow velocity was around 20 m/a in the middle part of Shirase drainage basin at about 2000 m in elevation, and less than 10 m/a in the inland region above 3000 m.

5. Thinning of the ice sheet

The thickness change of the ice sheet was observed from two elevation surveys in 1969/70 and 1973/74 along Route A at 72°S latitude (Naruse, 1978a, 1979). Elevations of the ice sheet surface and bedrock surface are shown in Fig. 4a. In Fig. 4b the estimated balance velocities, as mentioned later, are compared with the observed surface flow velocities. Whereas the estimation of the balance velocity has some error, they show similar profiles; they have a velocity peak in the distance of 100 km from a reference point in the Yamato Mountains.

As shown in Fig. 4c, a large thinning rate, about 0.7 m/a, was observed in the main stream of Shirase Glacier between 39°E and 43°E, where the ice sheet thickness was about 2000 m. Possible causes of this thinning include decrease of accumulation rate and increase of flow velocity by basal sliding (Mae and Naruse, 1978).

Along Route SS (about 39.5°E in latitude), thickness change rates were also obtained by the NNSS (JMR) satellite doppler positioning system between 1982 and 1987 as shown in Fig. 5. Although the values of thinning rates were dispersed due to the low accuracy of JMR elevation measurements, large ice thinning is suggested in the lower part of Shirase Glacier, between 150 km and 300 km from the glacier mouth (2000 m to 2600 m in elevation) (Nishio *et al.,* 1989).

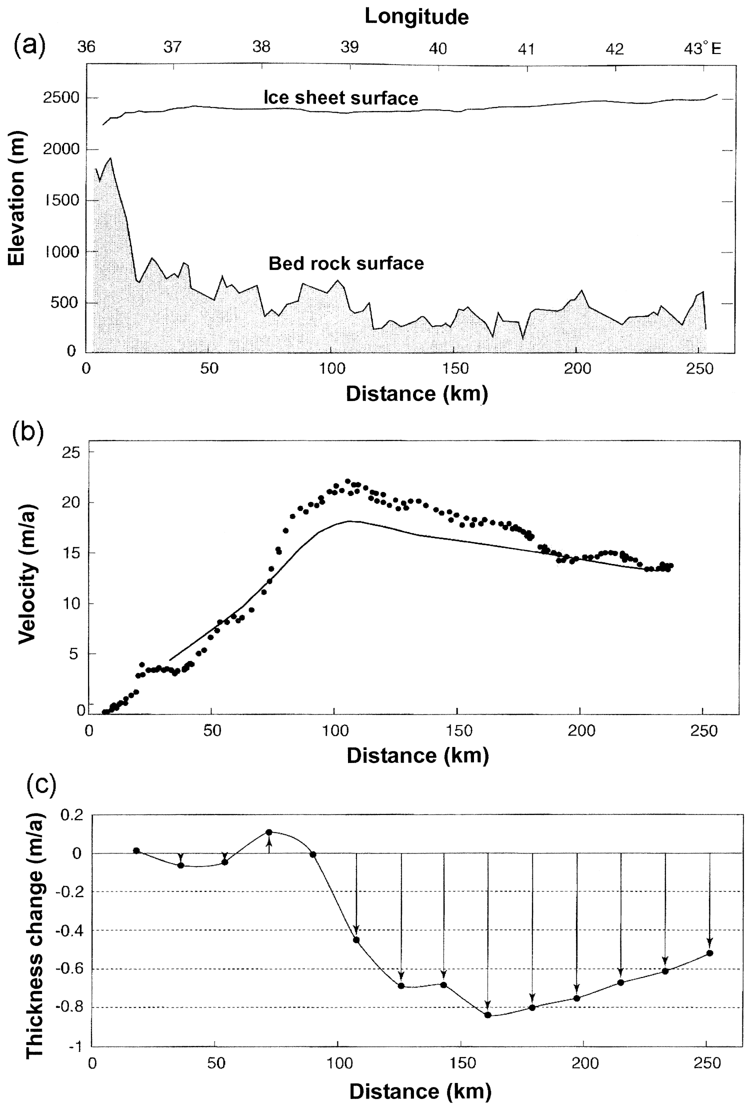

Fig. 4. Ice flow along Route A (about 72°S).
Positioning and ice flow measurement were made by surveys of a triangulation chain 250 km in length from the Yamato Mountains (Naruse, 1978b).
(a) Surface profiles of ice sheet and bedrock.
(b) Ice flow velocity.
Solid circles show observed horizontal velocities and a thin line shows balance velocities across Route A.
(c) Thickness change rate of ice sheet.
A positive value indicates thickening, and a negative value thinning.

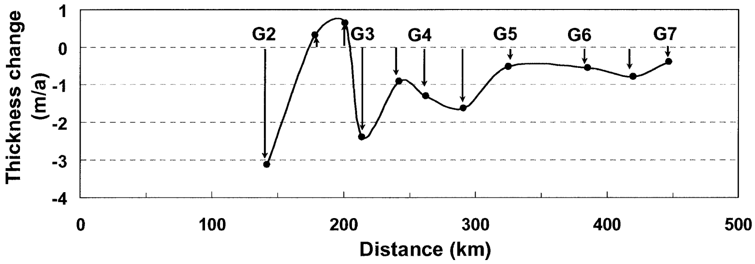

Fig. 5. Thickness change rates of the ice sheet along Route SS. Positive values indicate thickening and negative values thinning. The data were derived from the NNSS (JMR) measurements (Nishio et al., 1988). The distance from the mouth of Shirase Glacier is the same as in Fig. 4.

6. Balance velocity of ice sheet flow

6.1. *Balance velocity*

Assuming a steady ice flow and steady topography, the mass flux through any vertical section perpendicular to the flow direction must balance upstream accumulation. The flow velocity averaged over the ice thickness is calculated from the surface mass balance (accumulation rate) and ice thickness, which is on the assumption of a steady flow and is called the balance velocity (Budd *et al.,* 1971; Paterson, 1994). The balance velocity V_b is obtained from:

$$V_b = \frac{1}{HW} \int b \cdot w(x) \mathrm{d}x, \quad (1)$$

where b is surface mass balance, x is distance from the reference point to an ice flow source, $w(x)$ is width between two adjacent flow lines at x, H is the ice sheet thickness, W is the width between two flow lines at the reference point, and b is integrated in the area between the two flow lines from the reference point up to a flow source.

During the East Queen Maud Land Project (1982–1986), elevation maps of the ice sheet surface and bedrock were made over the five drainage basins in East Dronning Maud Land, and ice thickness H can be obtained from the difference of the two elevations. Considering the flow perpendicular to the contour lines, flow lines are obtained from the ice sheet surface elevation map. The distribution of surface mass balance b is shown in Fig. 2. Using these data sets, the balance velocity V_b was calculated by eq. (1) along flow lines. The results are shown in Fig. 3, where the balance velocities are presented as isolines of velocity and classified by color.

6.2. *Comparison with observed surface velocity*

The ice deformation obeys the well-known Glen's flow-law

$$\dot{\varepsilon} = A\tau^n, \quad (2)$$

where $\dot{\varepsilon}$ is the effective strain rate, A is a temperature dependent coefficient, τ is the effective stress and the exponent n is 3 for ice. When ice flow is only due to the internal deformation, without basal sliding, the ice flow velocity $V(z)$ of an isothermal ice sheet can be approximated as follows:

$$V(z)=V_0\left\{1-\left(1-\frac{z}{H}\right)^{n+1}\right\}, \tag{3}$$

and

$$V_0=\frac{2A}{n+1}(\rho g\theta)^n H^{n+1}, \tag{4}$$

where V_0 is the surface velocity, z is height from the bed, H is the thickness of the ice sheet, ρ is the density of ice, g is gravity acceleration and θ is the slope of the ice sheet surface. As the balance velocity V_b is obtained by integrating $V(z)$ over z,

$$V_b=\frac{n+1}{n+2}V_0. \tag{5}$$

Since the exponent n is almost 3 for ice, the balance velocity V_b is 80% of the surface velocity V_0 for an isothermal ice sheet. For a non-isothermal ice flow, such as a polar ice sheet, ice temperature changes with depth; it is usually colder near the surface and warmer near the bottom. As the coefficient A in eq. (2) exponentially increases with temperature, the deformation of ice in the bottom part is comparatively large and therefore V_b is closer to V_0. For a temperature profile such as is seen at Byrd Station (Gow, 1968) or Ice Stream B (Engelhardt *et al.,* 1990), V_b is $\geqq$90% of V_0. When there is basal sliding at the bottom, V_b is closer to V_0. The estimated balance velocity V_b is compared with the observed surface velocity V_0 as follows.

(a) Route SS along a streamline (about 40°E)

Along Route SS almost aligned to the longitude of 39.5°E, glaciological grid points G2 to G7 and other auxiliary points were established in 1982–84, and resurveyed after 4 to 5 years, where positions and elevations were measured by the NNSS (JMR-4A) satellite doppler positioning system (Nishio *et al.,* 1989).

In Fig. 6c the estimated balance velocities are compared with observed surface velocities, and the ice sheet surface and bedrock profiles are shown in Fig. 6a, where the distance is taken from the mouth of Shirase Glacier. The balance velocities agree well with the observed surface velocities. At almost all points the balance velocities are 10 to 20% less than the surface velocity, which corresponds to the tendency of the balance velocity to be smaller than the surface velocity in the state of no-basal sliding as described before. The balance velocity at the mouth of Shirase Glacier was about 3 km/a, which also agrees with the observed values 2.4–2.7 km/a.

(b) Route from S16 to Dome Fuji Station

Along the traverse route from S16 to Dome Fuji Station, glaciological observations of surface mass balance, flow velocities and ice-radar measurements have been done since 1992.

In Fig. 7 the balance velocities are compared with the surface velocities observed by differential geodetic survey using GPS (Global Positioning System) from 1992 to 1995. The balance velocities seem to be in agreement with the observed surface velocities at most points except in the coastal region around S16. In this coastal region, the

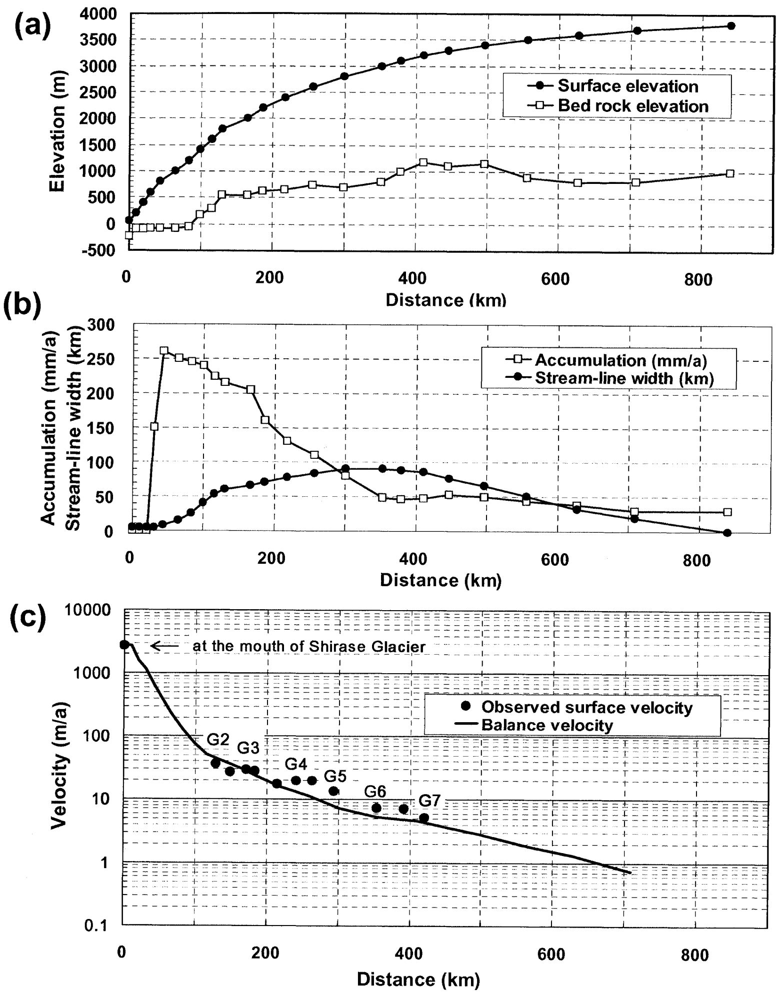

Fig. 6. Balance Velocity and factors for its estimation along a stream line near Route SS (along 40°E).
Distance is from the mouth of Shirase Glacier to Dome Fuji area.
(a) Profiles of ice sheet surface and bedrock topography.
(b) Surface mass balance (accumulation) and stream line width.
Stream line width is a distance between two adjacent stream lines for the integration of the surface mass balance.
(c) Balance velocity and observed surface velocity.
A solid line shows estimated balance velocity along a stream line near Route SS. Solid circles show horizontal component of the surface velocity observed by NNSS (Navy Navigation Satellite System) (Nishio et al., 1988).

topographies of the ice sheet surface and bedrock are complicated and it is difficult to perform a calculation with good accuracy because of the roughly estimated flow lines.

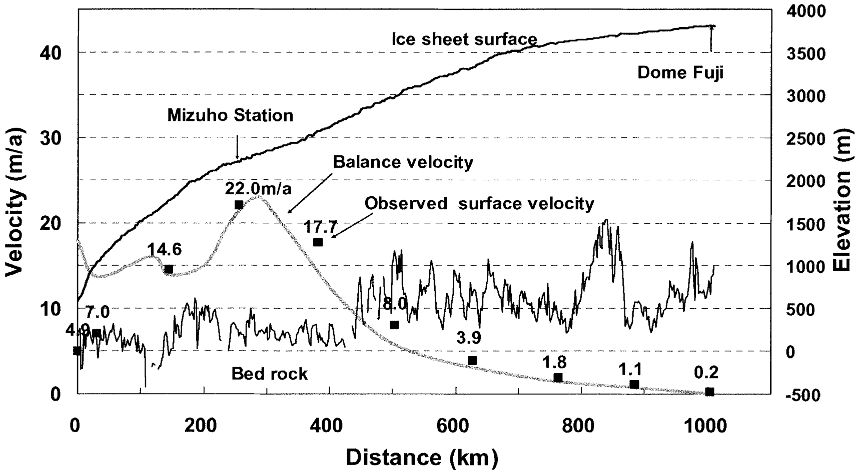

Fig. 7. Surface velocity and balance velocity along the route from S16 to Dome Fuji. Squares show horizontal components of the surface velocities observed by GPS (Global positioning system). The thick line is the balance velocity from Fig. 3 and the thin lines show profiles of ice sheet surface and bed rock topography (modified from Takahashi et al., 1997).

6.3. *Basal sliding area*

For an ice flow caused by ice deformation, the following relation is derived from eqs. (4) and (5)

$$\frac{V_b}{H}=\frac{2A}{n+2}\tau_d{}^n, \text{ where } \tau_d=\rho g H\theta. \tag{6}$$

Equation (6) means that the balance velocity V_b scaled with H has an exponential relation to the basal shear stress $\rho g H\theta$ with the power of n which is 3 for ice. When the flow velocity by basal sliding V_s is added to the deformation flow, V_b/H would become larger than the exponent relation of eq. (6) as follows:

$$\frac{V_b}{H}=\frac{2A}{n+2}(\rho g H\theta)^n+\frac{V_s}{H}. \tag{7}$$

In Fig. 8 the relation between the scaled balance velocity V_b/H and the basal shear stress along a flow line close to Route SS (about 40°E in longitude) is shown. In the inland region (at an elevation of >1400 m) the relation fits a line with the power of $n=3$, which means that the ice flow in this area is almost due to the deformation flow in Glen's flow-law. In the lower area, below 1200 m, the ice flow shows large velocity due to basal sliding, and in the area below 1000 m the velocity of basal sliding should be more than 10 times the velocity due to ice deformation.

6.4. *Accuracy of balance velocity*

Roughly examining errors of eq. (1) for balance velocity accuracy, the surface

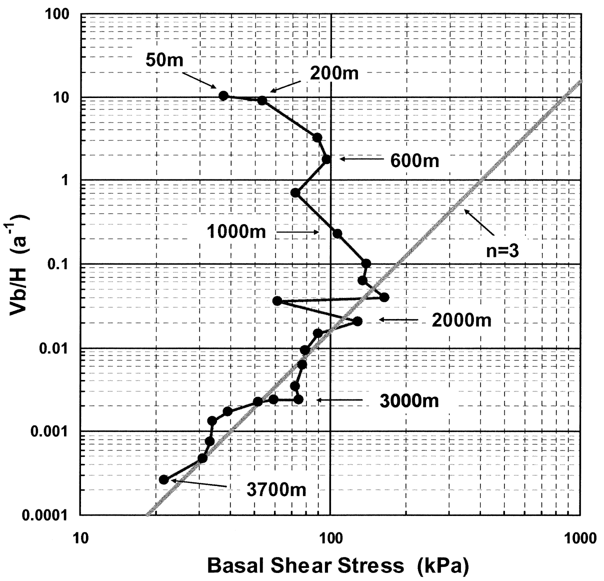

Fig. 8. Scaled balance velocity V_b/H and basal shear stress along a stream line near Route SS. Solid circles are the balance velocity V_b scaled by ice sheet thickness H along a stream line near Route SS. Numbers attached to the circles indicate elevation. A thick line corresponds to a flow with Glen's flow-law exponent $n=3$ without basal sliding.

mass balance b appears to have an error of about 10%, the integral area between the two flow lines 5–10%, and the ice sheet thickness H 5–10%. The errors of thickness H come from surface elevation and bedrock elevation, in which the error of surface elevation is small, 5–10 m, but bedrock elevation is large, 100–200 m, due to the lack of data. Simply adding the errors, the error of the balance velocity can be estimated as 20–30%. Though the error is estimated to be large, the calculated balance velocities have good agreement with the observed surface velocities in the three areas shown in Fig. 5b, Fig. 6c and Fig. 7. This relatively good correlation appears to be due to the fact that the observed velocities are situated along the main flow, where ice thickness has relatively small error due to the thick ice sheet; flow lines have good accuracy because of the simple topography; and the ice thickness was usually observed. Considering that the errors of the integral area and ice thickness would be smaller in an area in the main flow and with simple topography, the error of the balance velocity becomes smaller, about 10–20%. The poor correlation with velocity on S16 is due to the fact that the estimated flow lines for balance velocity were about 50–100 km apart in the middle of the ice flow and they did not respond to the small relief of the ice sheet.

The comparison of mass input and output in/from an ice sheet is important in the Shirase drainage basin for ice sheet instability; several papers have discussed this topic (*e.g.* Yamada and Watanabe, 1978; Fujii, 1981; Takahashi *et al.,* 1994), but the errors of estimation are too large to discuss exactly. In this area more observations, especially of bedrock topography and surface mass balance, are necessary.

7. Concluding remarks

The surface mass balance obtained mainly by the snow stake method was more than 250 mm/a in the coastal region, less than 50 mm/a in the inland region higher than 3500 m in altitude, and about 100 mm/a on average in five drainage basins in East Dronning Maud Land.

The ice flow velocity was 2.4–2.7 km/a at the mouth of Shirase Glacier, which is a typical ice stream, and around 300 m at the terminus of Ragnhild drainage basin, which is a typical sheet flow. On the ice sheet, the surface flow velocity was around 20 m/a in the middle part of Shirase drainage basin at about 2000 m in elevation, and less than 10 m/a in the inland region above 3000 m.

Assuming a steady ice flow, the balance velocity is obtained by integrating the surface mass balance from a reference point to the flow source between adjacent stream lines. The obtained balance velocities corresponded well to the observed surface flow velocities along several routes on the Shirase Glacier. According to the relation between scaled balance velocity and basal shear stress, basal sliding should be large in the part of Shirase Glacier below 1200 m in surface elevation.

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