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A preliminary estimation of drifting snow convergence along a flow line of Shirase Glacier, East Antarctica

Shuhei TAKAHASHI
Kitami Institute of Technology, Koen-cho 165, Kitami 090 Japan

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Abstract

The drifting snow convergence is obtained by estimating the snow drift transport rate caused by katabatic winds on the ice sheet. The estimated convergence showed a large positive value in the coastal region and a negative value at about 300 km distance from the coast, whereas it is negligible in the inland region further than 400 km. The large amount of net accumulation in the coastal region can be roughly explained by the drifting snow convergence in addition to the precipitation and the sublimation.

1. Introduction

On the slope of the ice sheet in Antarctica, katabatic winds are formed by the gravitational force of cold air masses, and drifting snow is generated throughout a year. This drifting snow redistributes an enormous mass of snow along the wind stream line. On the convex surface of the inland part of the ice sheet, the wind speed increases to leeward and snow particles are exported from the surface by the divergence of drifting snow. This exported snow mass settles down on the concave surface where the wind speed decreases. This redistribution is believed to have an effect on the distribution of net accumulation.

It has been reported that local mass balance is related closely to the surface topography of inland ice sheets (Schytt, 1955 ; Swithinbank, 1959 ; Black and Budd, 1964 ; Gow *et al.*, 1972). Whillans (1975) discussed mass movement due to the drifting snow in Marie Byrd Land, Antarctica. Takahashi (1988) estimated the divergence of drifting snow on Mizuho Plateau, East Antarctica, and attributed the cause of the formation of the bare ice field to the divergence.

In this paper, the drifting snow convergence is estimated one-dimensionally on an ice flow line of Shirase Glacier located in East Queen Maud Land, East Antarctica (Fig. 1), and compared with net accumulation.

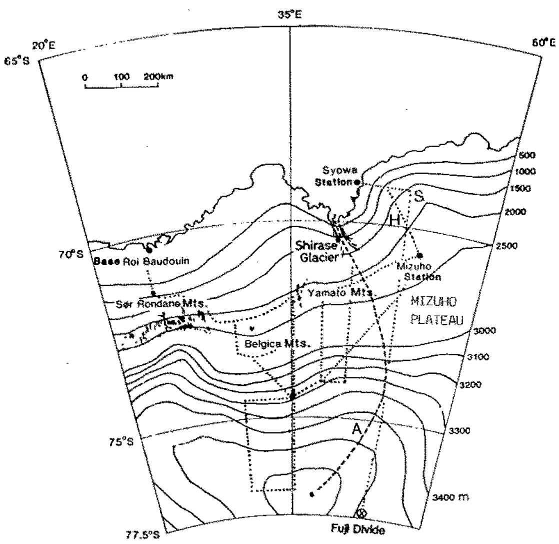

Fig. 1. Contour map of East Queen Maud Land, East Antarctica. Dotted lines are the main routes of oversnow traverses by the Japanese Antarctic Research Expedition. Dashed line A represents the flow line of Shirase Glacier used for calculation of drifting snow convergence.

2. Drifting snow convergence

2. 1. *The equations of katabatic winds*

Katabatic winds depend on surface slope and inversion intensity (*e.g.* Ball, 1960 ; Adachi, 1983 ; Schwerdtfeger, 1984). In the two-layer model, the wind speed of the lower layer is determined by balancing of the Coriolis force, the friction force and the gradient force of the lower layer. Aligning the x-axis with the direction of the maximum slope and neglecting the geostrophic winds, the equations of this model are given by :

$$0=-(k/H)\ V\ u+f\ v+g(\Delta\theta/\theta)\sin A \quad \text{(1a)}$$

$$0=-(k/H)\ V\ v-f\ u \quad \text{(1b)}$$

where k is the friction coefficient, H is the inversion layer thickness, V is the absolute wind speed, u and v are x- and y- wind components, f is the Coriolis parameter, g is the gravity acceleration, A is the surface slope, $\Delta\theta$ is the inversion intensity (the potential temperature difference of the two layers) and θ is the mean potential temperature of the lower layer. The solutions of these equations are

$$V=(F_g H/k)\cos B \quad \text{(2a)}$$

$$\cos B=-f^2H/2F_gk+\sqrt{(f^2H/2F_gk)^2+1} \quad \text{(2b)}$$

where B is the deviation angle of the wind from the fall line and F_g (the gradient force) $= g(\Delta\theta/\theta)\sin A$.

At Mizuho Station, the annual mean wind speed was 11.1 m s^{-1} and B was about 45°, $A=4\times10^{-3}$ and $f=-1.387\times10^{-4}$ s^{-1}. Therefore, F_g is given by 2.18×10^{-3} m s^{-2} and k/H is 1.25×10^{-5} m^{-1} by solving eqs. (2a) and (2b). Since the annual mean temperature in 1982 was -33.3°C and k is assumed to be 5×10^{-3}, $\Delta\theta$ is 13.3 K and H is 400 m. From these values, the relation between wind velocity V and inclination A

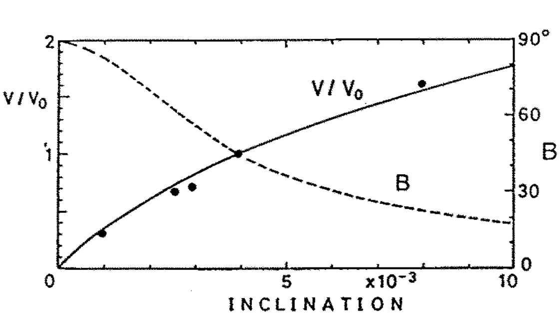

Fig. 2. Ratio of wind speed to that of Mizuho Station, V/V_0 and the deviation angle of wind direction B as a function of surface inclination obtained from eqs. (2a) and (2b). Solid circles are the ratios quoted from Inoue *et al.* (1983).

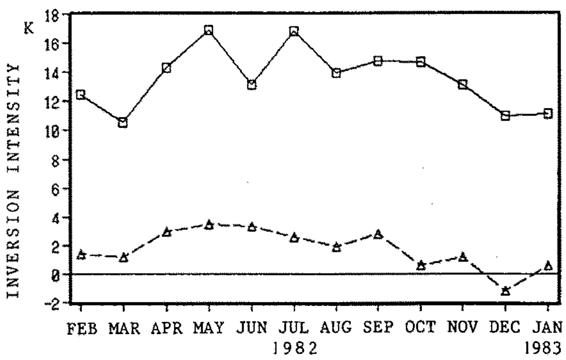

Fig. 3. Monthly inversion intensity from February, 1982 to January, 1983. Triangle represents the inversion intensity at Syowa Station obtained from the aerological data. Square represents the intensity at Mizuho Station obtained from the wind speed and temperature by eqs. (2a) and (2b).

with the same inversion intensity was obtained and shown in Fig. 2, where V is normalized by the wind speed at Mizuho Station V_0. The wind speed data obtained at various slopes on Mizuho Plateau by Inoue *et al.* (1983) agreed with this relation as shown in Fig. 2.

However, there are several problems in the coastal region. One of the important problems is the locality of inversion intensity. If the $\Delta\theta$ of 13.3 K, obtained at Mizuho Station, is constant everywhere, the annual mean wind speed is estimated to be about 25 m s^{-1} at the coast, which is too large and apart from the reality. Therefore, the inversion intensity should be more small at the coast. At Syowa Station, located on Prince Orav Coast, the monthly inversion intensity in 1982 was obtained from the difference between the surface potential temperature and the potential temperature extrapolated from the aerological data at the level of 850 mb and 700 mb, whereas the intensity at Mizuho Station was obtained from the wind speed and temperature by eqs. (2a) and (2b) with the given value of A, f, k and H. As shown in Fig. 3, the seasonal variations of inversion intensity at the two Stations were similar, but its annual means were very different : 1.7 K at Syowa Station and 13.6 K at Mizuho Station. Therefore, assuming a linearity to altitude, the annual mean of inversion intensity at the altitude of h (m) is given from the values at the two Stations as

$$\Delta\theta= 1.7+0.00534\ h. \quad \text{(3)}$$

From these eqs. of (2a),(2b) and (3), the annual mean of katabatic wind speed at any altitude can be estimated from its surface inclination and altitude.

2. 2. The equation of drifting snow convergence

At Mizuho Station, Takahashi (1985a) obtained a snow drift transport rate from March 1982 to January 1983 by integrating snow drift flux from the surface to a height of 30 m. The relation between the drift transport rate q (kg m^{-1}day^{-1}) and wind speed V (m s^{-1}) at 1 m height was as follows,

$$q=6.2\times10^{-2}\ V^{5.17}. \tag{4}$$

The annual transport rate at Mizuho Station Q_0 was estimated as 3×10^6 kg m^{-1}a^{-1}. From this value and eq.(4), the annual transport rate Q (kg m^{-1}a^{-1}) at a certain point with wind speed V can be given by

$$Q=Q_0(V/V_0)^m \tag{5}$$

where m is 5.17 and V_0 is the average wind speed at Mizuho Station, 11.1 m s^{-1}.

If the altitude of the ice sheet topography varies only with x-axis, the drifting snow convergence can be given by the difference in the x-components of the transport rate at two different points, 1 and 2, along x-axis :

$$\text{conv } Q=(Q_1\cos B_1-Q_2\cos B_2)/\Delta L$$
$$=Q_0\{(V_1/V_0)^m\cos B_1-(V_2/V_0)^m\cos B_2\}/\Delta L, \tag{6}$$

where ΔL is the horizontal distance between the two points, and B_1 and B_2 are the deviation angles at the two points.

2. 3. Drifting snow convergence along a flow line of Shirase Glacier

Nishio *et al.* (1984) examined the ice sheet profile along the ice flow line of Shirase Glacier shown in Fig. 1. On this profile in the length of 800 km, the drifting snow convergence is calculated by eq. (6) in a grid distance of 25 km. In Fig. 4, the calculated convergence is shown together with the calculated wind speed and deviation angle. In the Figure, the unit of convergence is converted from [kg m^{-3}a^{-1}] into [mm a^{-1}] : the water equivalent. The convergence is about 250 mm a^{-1} in the coastal region at 25 km distance from the coast, and about -150 mm a^{-1} with some deviation in the middle part of the ice sheet at about 300 km distance, around the altitude of 2000 m, whereas it is negligible on the inland part of the ice sheet further than 400 km, above the altitude of 3000 m. This result indicates that drifting snow is redistributed from the middle part of the ice sheet to the coastal region.

Along the same flow line, Nishio *et al.* (1984) obtained a theoretical profile of the ice sheet by Weertman's equation (Weertman, 1961). On this theoretical profile, the drifting snow convergence is calculated, as shown in Fig. 5. Compared with that of the real profile shown in Fig. 4, the variation of the convergence along the flow line is considerably large : a large value of 500 mm a^{-1} was seen at the coast, and a large negative value of about -250 mm a^{-1} was around the altitude of 2000 m.

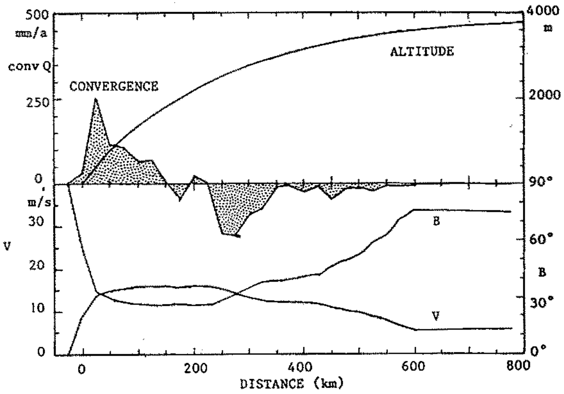

Fig. 4. Drifting snow convergence on a profile of the ice sheet along the flow line shown in Fig. 1. The wind speed V, the deviation angle B and ice sheet profile for the calculation are shown together. The ice sheet profile is quoted from Nishio *et al.* (1984).

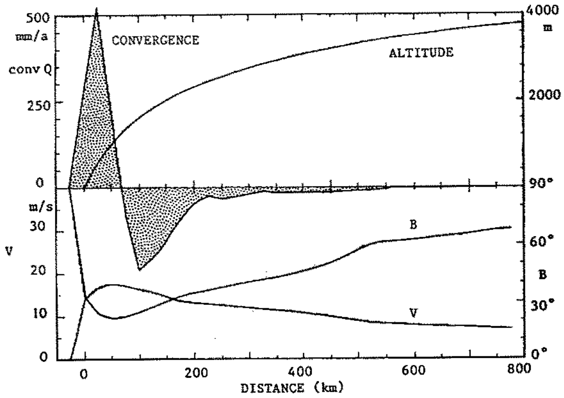

Fig. 5. Drifting snow convergence on the theoretical profile of the ice sheet calculated with Weertman's equation (Nishio *et al.*, 1984) along the same flow line in Fig. 1, with wind speed V, the deviation angle B and the ice sheet profile.

3. Discussion

3. 1. Net accumulation on Mizuho Plateau

At Mizuho Station (2230 m above sea level, 250

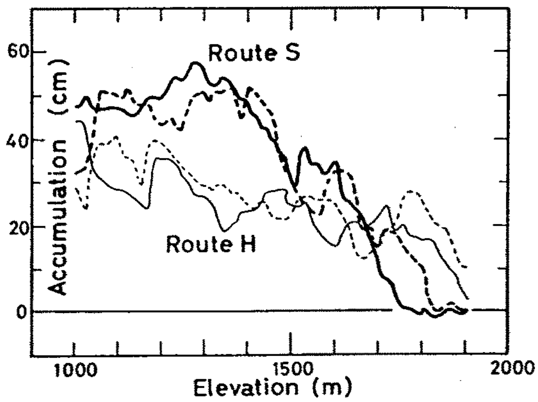

Fig. 6. Annual net accumulation by the snow stake method along Routes S and H in 1973 (solid lines) and 1974 (broken lines), quoted from Yamada *et al.* (1978).

km away from the coast), the net accumulation of about 70 mm a^{-1} (Narita and Maeno, 1979) is much smaller than the sum of the annual precipitation of about 200 mm a^{-1} (Kobayashi *et al.*, 1985 ; Takahashi, 1985b) and the sublimation of -50 mm a^{-1} (Fujii and Kusunoki, 1982 ; Takahashi, 1988). Takahashi (1988) explained this difference by the negative convergence (divergence) of drifting snow. In Fig. 4, the similar negative convergence is seen around the altitude of Mizuho Station, although the flow line is different from that of Mizuho Station.

Yamada *et al.* (1978) obtained the net accumulation on Mizuho Plateau by the snow stake method along two Route S and H (Fig. 6). They suggested that the significant difference of net accumulation between the two Route is caused by the difference of katabatic wind speed due to the surface topography of the ice sheet. This suggestion means the difference of drifting snow convergence due to the topography.

Assuming the snow density of 600 kg m^{-3}, a general view of Fig. 6 indicates that the net accumulations of the two Routes have a large value of 200 mm a^{-1} to 300 mm a^{-1} at the altitude between 1000 m and 1500 m, and decreases with altitude to a small value of 0 to 60 mm a^{-1} at the altitude of 2000 m. Since the precipitation and the sublimation are not expected to vary so much with the altitude change from 1000 m to 2000m, the variation of net accumulation in this coastal region is explained mainly by the drifting snow convergence shown in Fig. 4, which has a peak of about 250 mm.

3. 2. Drifting snow convergence on a theoretical profile of the ice sheet

In the theoretical profile given by Weertman's equation (Weertman, 1961), the drifting snow convergence showed a considerably large positive value at the coast and large negative value at 100 km distance from the coast (Fig.5). The large positive convergence is caused by the decreasing of katabatic wind speed owing to the inversion intensity decreasing, whereas the negative convergence is derived from the convex topography of the profile. This systematic dependency of drifting snow convergence on the topographic profile tends to make the profile smooth, and another equilibrium profile should be formed after a long period of time. Hence, the model of the equilibrium profiles of the ice sheet should take into account the effect of drifting snow convergence.

3. 3. Problems in the katabatic wind equations

There are several problems in the katabatic wind equations of eqs. (1a) and (1b) as will be described below.

(a) Inertia term

Parish and Bromwich (1986) assessed the inertia term of $u\partial u/\partial x$, which was neglected on the left-hand side of eq. (1a) by the assumption of homogeneous flow. By their assessment, the ratio of the inertia term to the gravitational force exceeds 0.25 at the distance of less than 75 km from the coast, which has an altitude lower than 1000 m, where the solution for homogeneous flow is not valid. However, since the negative convergence (divergence) in the middle part of the ice sheet is not affected by this term, the integration of convergence along the coastal flow line is unchanged, and therefore, the convergence is dispersed only in a wider costal region by the term.

(b) Thickness of inversion layer

The inversion layer thickness H was assumed to be constant in the katabatic wind equations, but is expected to change locally. One possible assumption is that the thickness is inverse to wind speed : $H = H_0 \ V_0/V$, where H_0 and V_0 are the thickness and wind speed at a reference point respectively. This assumption satisfies the continuity equation of the lower layer. By this assumption, the friction force is proportional to the cubic of wind speed, and the variation of wind speed with inclination is expected to be small compared with the solution of eqs. (2a) and (2b). This small variation of wind speed makes the magnitude of convergence small.

(c) Inversion intensity

The positive convergence of drifting snow in the coastal region is caused by the decreasing of katabatic wind due to the weakening of inversion intensity. Therefore the variation of inversion intensity is important for the calculation of the convergence. In this paper, the inversion intensity at an altitude is given from the intensities at two Stations by the assumption of a linearity to altitude. However, the intensity is controlled by many other factors: wind speed, surface condition, thickness of the inversion layer and especially the radiative condition. It is necessary to simulate the inversion intensity by a heat budget model, although the procedure is complicated. Otherwise, the inversion intensity should be observed extensively at various points on the ice sheet.

3. 4. Unsaturation of drifting snow

To obtain the drifting snow convergence by eq. (6), the drifting snow was assumed to be saturated. If the net accumulation is positive, the surface is generally covered with loose snow particles and these particles can be always supplied from the surface to the air, then the drifting snow is saturated. However, on a bare ice surface or a glazed surface, the net accumulation is zero or negative by sublimation and divergence of drifting snow, and no loose snow particles are available from the surface except for the occasion of precipitation. In this state, the drifting snow is unsaturated and the equations are not valid. For the convergence taking account of the unsaturation, another possible method is to calculate the snow transport rate along a trajectory of katabatic wind by checking whether the surface can supply snow particles or not.

4. Concluding remarks

Along the ice flow line on Shirase Glacier, East Antarctica, the katabatic wind speed was calculated from the surface inclination and altitude by assuming that inversion density varies linearly with altitude. From this wind speed, the drifting snow convergence was estimated one-dimensionally on the flow line.

The estimated convergence showed a large positive value of about 250 mm a^{-1} near the coast and a negative value of about -150 mm a^{-1} at about 300 km from the coast, whereas it is negligible in the inland region further than 400 km. The local variation of net accumulation in the coastal region can be roughly explained by the positive drifting snow convergence in addition to the precipitation and the sublimation.

The drifting snow convergence depends on a curvature of topography : it is negative in convex topography and positive in concave one. This systematic variation of the convergence on a profile should have an effect on it in a long period of time. This effect of the drifting snow convergence should be taken into account when a theoretical model of the equilibrium profile of an ice sheet is made.

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