OBSERVATION OF SNOW DRIFT FLUX AT MIZUHO STATION, EAST ANTARCTICA, 1982

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Abstract: Continuous observation of snow drift flux was carried out at Mizuho Station (70°42'S, 44°20'E, 2230 m above sea level), East Antarctica, in 1982. Snow drift flux at 1 m height was well correlated with wind velocity. The correlation coefficient on a logarithmic plot was between 0.8 and 0.9. The drift flux was proportional to about the 8 power of wind velocity through the year. The power decreased above −20°C. The drift flux increased when precipitation was observed. From the variation of drift flux, precipitation intensity can be estimated.

1. Introduction

As part of the East Queen Maud Land Program, the 23rd Japanese Antarctic Research Expedition (JARE-23) made the observations of drifting snow at Mizuho Station (70°42'S, 44°20'E, 2230 m above sea level), East Antarctica. In this region, the drifting snow occurs by katabatic winds throughout the year and plays an important role in snow redistribution on the ice sheet (YAMADA and WATANABE, 1978). Mass transportation by drifting snow is enormous; KOBAYASHI (1978) estimated the amount at 10^9 kg/km year at the station, which is not small in comparison with the mass transport by ice sheet flow. Many other investigators examined the drifting snow at Mizuho Station and clarified the characteristics (YAMADA and NARITA, 1975; KOBAYASHI and YAKOYAMA, 1977; INOUE and FUJINO, 1977; NARITA, 1978; KOBAYASHI *et al.*, 1983). However, many problems concerning drifting snow remain, for example, estimation of precipitation, fall velocity of drifting snow particles, dependence of drifting snow on temperature and so on.

The observation of drifting snow by JARE-23 was performed from January 1982 to January 1983; measurements were carried out on snow drift flux, visibility, fall velocity of drifting snow particles, repose angle of particles, erosion ability of drifting snow and so on. In this paper, the measurement system of drifting snow and several findings of snow drift flux are reported.

2. Measurements

Snow drift flux was continuously observed by a slit type collector (surface–0.1 m in height), four cyclone type collectors (0.3, 0.5, 1, 2 m) and four rocket type collectors (4, 8, 15, 30 m). As shown in Fig. 1, snow particles collected by the slit type and cyclone type collectors fell into a chamber 4 m under the snow surface and were weighed once a day. The snow particles stored in the rocket type collectors, which were installed on a 30 m height tower, were weighed at 5 or 10 day intervals.

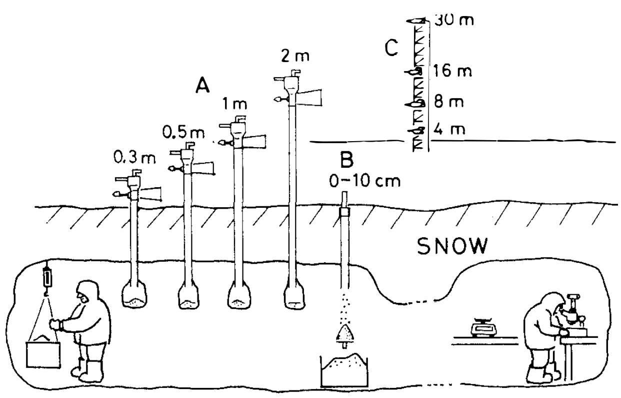

Fig. 1. Illustration of measurement system of snow drift flux. A: cyclone type collectors, B: slit type collector, C: rocket type collectors.

The cyclone type collector was pointed toward the wind direction by vanes (Fig. 2a) and its aerodynamic efficiency is about 0.6 according to a wind-tunnel calibration (KOBAYASHI, 1974). Taking into account the inertia of snow particles, the collection efficiency can be higher than this aerodynamical value. However, the operational efficiency value 0.6 was yet adopted in this observation in order to compare with previous works.

The slit type collector was positioned to measure drift flux close to the snow surface; it has a narrow slit inlet 1 cm in width and about 10 cm in height (Fig. 2b). Its collection efficiency was estimated to be the same as that of the cyclone type, based on extrapolation of the drift flux obtained by the cyclone type. Its inlet must be manually pointed toward the wind direction, but the collector rarely missed the correct wind direction owing to the constancy of the katabatic wind direction.

The rocket type collector was deviced by JARE-22, having a shape similar to that of BUDD *et al.* (1966) (Fig. 2c); its collection efficiency was estimated to be the same as that of the cyclone type, based on calibration by the cyclone type on the 30 m height tower. Examples of drift flux profiles are shown in Fig. 3; discontinuity was not found in the profiles, while the cyclone type was used below 2 m and the rocket type above 4 m.

Wind velocity at 1 m height was evaluated from that at 6.9 m. As routine meteorological work at Mizuho Station, wind velocity was measured at a height of 6.9 m by a windmill type anemometer (TAKAHASHI *et al.*, 1983). According to previous observations of wind velocity profiles on the 30 m height tower (KOBAYASHI *et al.*, 1983), the ratio of wind velocities at the two different height can be obtained; the wind velocity at 1 m height is evaluated from that at 6.9 m less 16%.

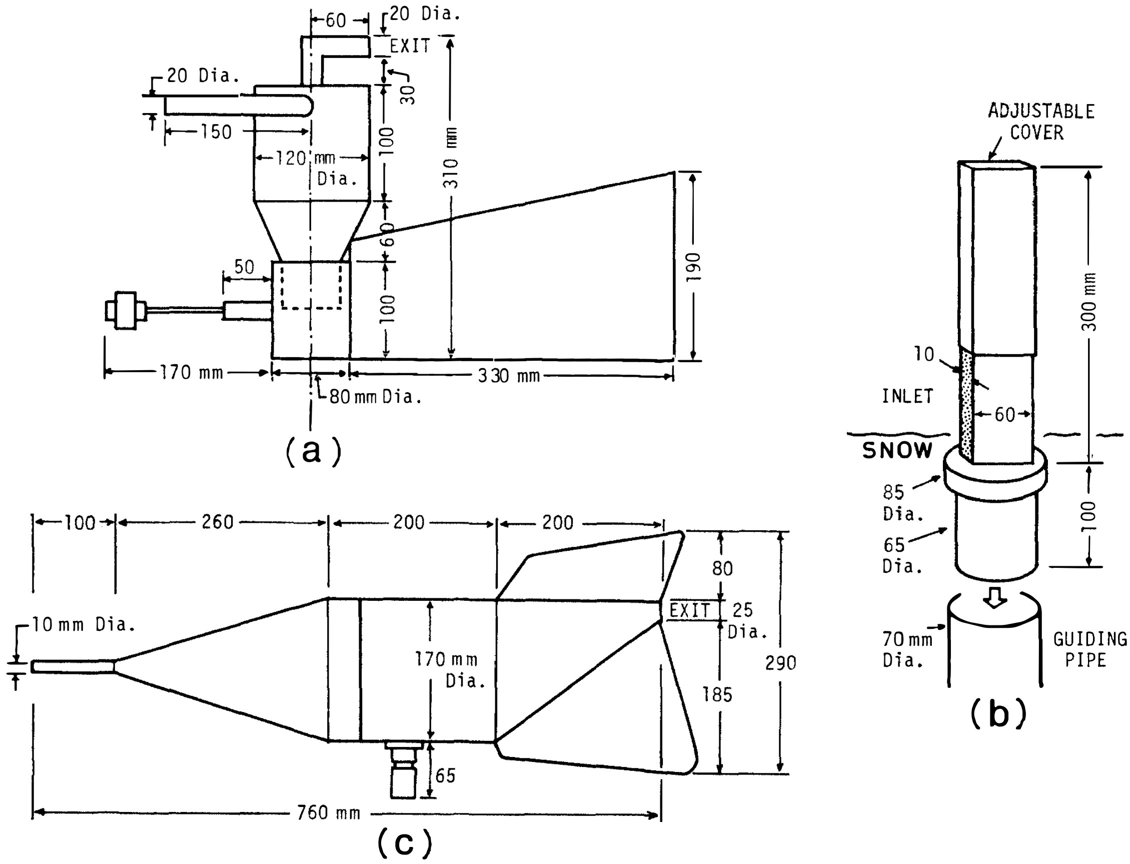

Fig. 2. Snow drift collectors: (a) cyclone type with vanes, (b) slit type and (c) rocket type.

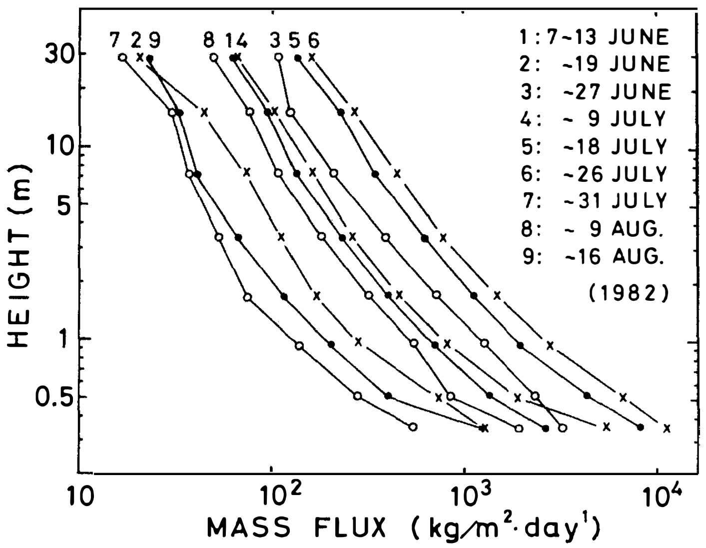

Fig. 3. Snow drift flux profiles at Mizuho Station from June to August 1982.

3. Results and Discussion

3.1. Relation between snow drift flux and wind velocity

The snow drift flux at a height of 1 m adequetly correlated with wind velocity at Mizuho Station (Fig. 4). In Table 1, the correlation between daily snow drift flux

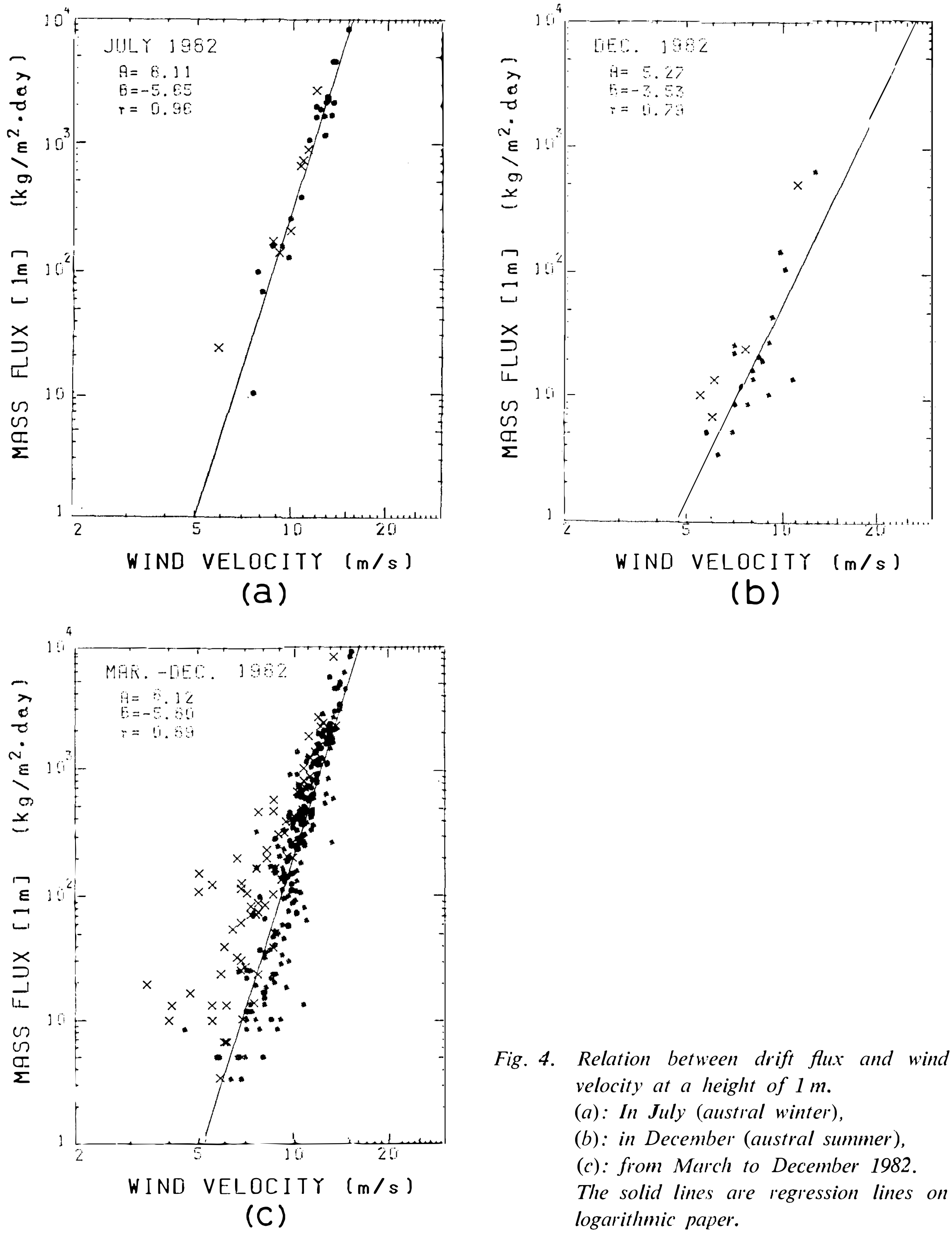

Fig. 4. Relation between drift flux and wind velocity at a height of 1 m.
(a): In July (austral winter),
(b): in December (austral summer),
(c): from March to December 1982.
The solid lines are regression lines on logarithmic paper.

and wind velocity on a logarithmic plot at 1 m height is shown monthly from March to December 1982. Since the drift flux was measured by weighing the store of drifting snow at 13 or 15 h everyday, the wind velocity was obtained by averaging 3-hourly values from 15 h on the previous day to 12 h on the day. The correlation coefficient on a logarithmic plot had a high value of 0.8 or 0.9. Therefore the drift flux can be expressed as a power of wind velocity.

When precipitation of snow or ice prisms was observed (marked by "×" in Fig. 4), the drift flux increased as compared with the case of no precipitation. Data with precipitation were not included in the correlating calculations in Table 1 in order to eliminate the effect of precipitation.

Table 1. Correlation between mass flux $F(kg/m^2 \cdot day)$ and wind velocity $V(m/s)$ at a height of 1 m on a logarithmic plot from March to December 1982.

1982	n	r	A	B	$\bar{V}$ (m/s)	$\bar{F}$ (kg/m²·day)
March	15	0.90	6.2	−3.7	8.7	112
April	19	0.87	6.0	−3.4	10.3	453
May	30	0.87	8.3	−5.9	10.9	563
June	14	0.97	8.1	−5.8	10.3	313
July	23	0.96	8.1	−5.7	11.3	780
August	22	0.90	8.1	−5.7	10.3	307
September	21	0.80	8.2	−6.0	10.5	302
October	23	0.92	9.4	−7.2	10.8	325
November	24	0.88	8.0	−6.0	9.4	63
December	19	0.79	5.3	−3.5	8.3	20
Mar.–Dec.	210	0.89	8.1	−5.8	10.1	234

n: Data number, r: correlation coefficient, A: regression coefficient, B: intercept on logarithmic paper given by the regression equation

$$\log F = A \log V + B \quad (F = 10^B \cdot V^A),$$

$\bar{V}$ and $\bar{F}$: monthly average of V and F respectively.

The drift flux at 1 m height was proportional to the 8.1 power of wind velocity throughout the observation period, as shown in the regression coefficient in Table 1. This power is larger than that for the drift transport rate to wind velocity, which was reported as 3 or 4 in other districts (*e.g.*, KOBAYASHI, 1972). This difference can be derived from the difference in mechanism of particle transport between saltation and suspension; the former prevails in the drift transport rate whereas the latter predominates in the drift flux at 1 m height. However, the details of the mechanism underlying the difference between the two powers has not yet been clarified.

The power of drift flux showed a seasonal change as shown in Table 1; it was above 8 between May and November and below 8 in other months. This change can be derived from the temperature change as shown in the next section.

The threshold wind velocity for occurrence of drifting snow was 6 or 7 m/s, as shown in Fig. 4. Considering the drift flux less than 5 kg/m²·day to be negligible, drifting snow without precipitation occurred at wind velocities more than 6 or 7 m/s, whereas drifting snow with precipitation occurred at smaller wind velocity.

3.2. *Dependence of drift flux on temperature*

The seasonal change of the power by which drift flux is related to wind velocity, mentioned in the previous section, is explained by the temperature dependence of the power. In Table 2, the correlation of drift flux and wind velocity on a logarithmic plot is shown as a function of temperature. The power showed a maximum value of 8.0 at temperatures ranging between −40 and −50°C and decreased at relatively high temperatures; in particular it was 4.4 above −20°C.

Table 2. Correlation between mass flux F(kg/m²·day) and wind velocity V(m/s) on a logarithmic plot as a function of temperature.

Temperarure T(°C)	n	r	A	B	$\bar{V}$ (m/s)	$\bar{F}$ (kg/m²·day)
−10> T>−20°C	21	0.73	4.4	−2.7	8.3	21
T>−30°C	39	0.87	7.3	−5.1	9.2	85
T>−40°C	53	0.89	7.8	−5.4	10.8	480
T>−50°C	88	0.92	8.0	−5.6	10.6	389
T>−60°C	14	0.84	7.5	−5.2	9.7	143
Total	215	0.89	8.0	−5.8	10.1	218

Notation same as Table 1.

Though we cannot fully explain this temperature dependence, the following argument suggests an explanation. The fall velocity of snow particles depends on particle shape. The shape of drifting snow particles showed a temperature dependence related to the observed repose angle in Mizuho Station, which had a maximum value between −30 and −40°C (TAKAHASHI *et al.*, 1984). Therefore, the fall velocity would depend on temperature and can cause a temperature dependence of drift flux, because the drift density is expressed as a power of the fall velocity (SHIOTANI and ARAI, 1953) and the drift flux is the product of drift density and wind velocity.

Many other factors affecting the temperature dependence can be considered. On the saltation of snow particles, NARITA (1978) reported on increase of saltation path length with decreasing temperature; this would cause an increase of drift flux at low temperatures. On the particles' rising from the snow surface, SCHMIDT (1980) examined the temperature dependence of the threshold wind velocity for transport of snow. On wind velocity, air turbulence would depend on temperature; seasonal variations of roughness parameter, friction velocity and atmospheric stability at Mizuho Station were reported by KOBAYASHI (1979) nad KOBAYASHI *et al.* (1983).

3.3. *Increase of drift flux due to precipitation*

In a region on a katabatic wind slope, as at Mizuho Station, estimation of precipitation is difficult owing to the strong drifting snow; accumulation obtained by the snow stake method does not represent precipitation accurately, because of the redistribution of drifting snow. We attempt another way of estimation.

The drift flux increased in the case of precipitation, as mentioned in Section 3.1. The effect of precipitation appeared as a vertical profile variation of snow drift flux. On logarithmic paper, the drift flux profile lost linearity when precipitation was observed; its variation was large at a large height, as shown in the profile of June 23, 1982 in Fig. 5. The reason is explained as follows. While ordinary drifting snow is nearly inversely proportional to height, the increase of horizontal mass flux due to precipitation is proportional to wind velocity and increases with height. Therefore the ratio of the increase to the ordinary drift flux, which is the variation due to precipitation on logarithmic paper, is large at large height.

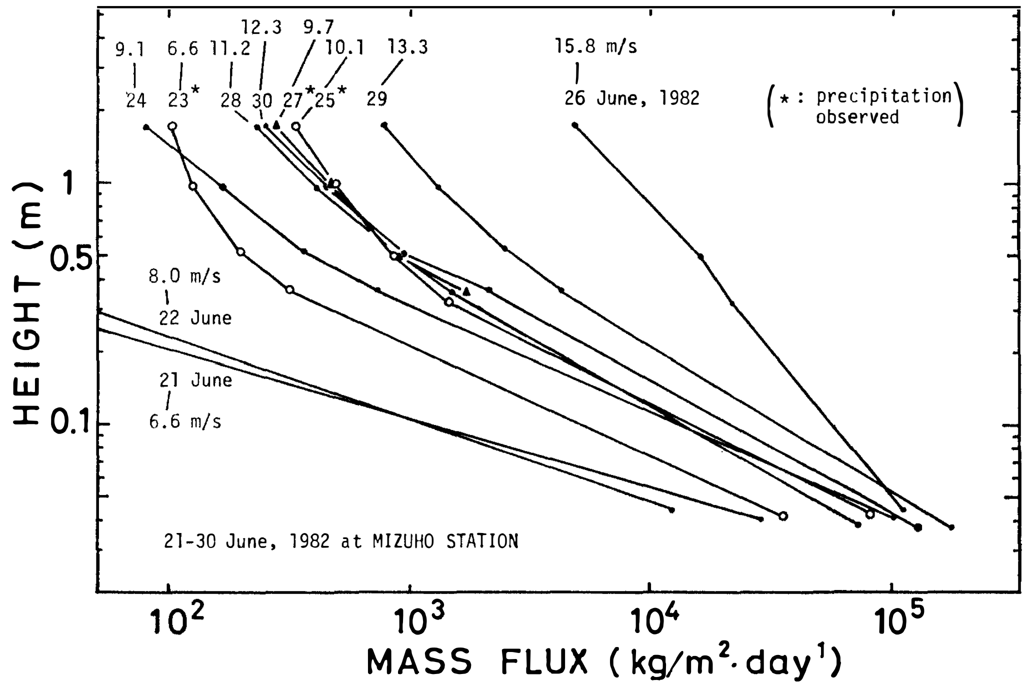

Fig. 5. Vertical profiles of daily drift flux from June 21 to 30, 1982. "" is attached to the date when precipitation was observed.*

The precipitation intensity can be estimated by the increase of horizontal mass flux as follows. The total drift flux $F(z)$ at a height of z is the sum of a drift flux without precipitation $F_d(z)$ and the horizontal mass flux due to precipitation $F_p(z)$;

$$F(z)=F_d(z)+F_p(z). \qquad (1)$$

Here $F_p(z)$ is a product of precipitation density N and wind velocity $V(z)$;

$$F_p(z)=NV(z), \qquad (2)$$

where N can be considered as an increase of drift density due to precipitation independent of height. Since precipitation intensity P is the product of precipitation density N and snow fall velocity W, P is obtained from eq. (2);

$$P=NW=F_p(z)W/V(z). \qquad (3)$$

For example, on June 23 in Fig. 5, V (1 m) was 5.5 m/s and F_p (1 m) was 80 kg/m^2·day which was obtained to a first approximation as the variation from a line extrapolated from the lower part, below 40 cm, where the ratio of the increase was small.

Assuming that W is 0.5 m/s according to the measurement of fall velocity at Mizuho Station (TAKAHASHI *et al*., 1984), the precipitation intensity P of this day was evaluated as 7 kg/m^2·day or 0.7 cm water equivalent in a day.

Another way of evaluating F_p (1 m) is to find the variation from the empirical relation between drift flux and wind velocity without precipitation, as shown in Fig. 4. This way is more convenient when the relation is given. However, even these estimates are not accurate in strong drifting snow. For example, the estimation of the precipitation of 10 kg/m^2·day is valid below a wind velocity of 12 or 13 m/s, supposing that the ratio $F_p(z)/F_d(z)$ of 10% can be measured.

4. Concluding Remarks

Continuous observation of snow drift flux was carried out at Mizuho Station, East Antarctica, in 1982. The snow drift flux at 1 m height closely correlated with wind velocity; the correlation coefficient on a logarithmic plot was 0.8 or 0.9. The threshold wind velocity at a height of 1 m for occurrence of drifting snow was 6 or 7 m/s.

The drift flux was proportional to about the 8th power of wind velocity throughout the year below -20°C; however, it decreased at high temperature, above -20°C. This temperature dependence can be explained by the temperature dependence of snow particle shape, which influences the fall velocity of particles and therefore the drift density.

When precipitation was observed, the drift flux increased and the drift flux profile lost linearity on logarithmic paper. From the increase of drift flux, the intensity of precipitation can be estimated.

Acknowledgments

The authors wish to express their sincere thanks to Mr. H. NISHIMURA, the Institute of Low Temperature Science, Hokkaido University, who provided them with rocket type drift snow collectors. Many thanks are due to the members of wintering party of JARE-23 for their kind support in the observation of drifting snow at Mizuho Station. Thanks are also due to Dr. S. KOBAYASHI and Dr. N. MAENO, the Institute of Low Temperature Science, Hokkaido University, for their valuable comments.

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(Received May 17, 1984; Revised manuscript received September 12, 1984)