[Journal of the Faculty of Science, Hokkaido University, Ser. VII (Geophysics), Vol. VI, No. 1, 1979]

Vertical Profile and Horizontal Increase of Drift Snow Transport*

Masao TAKEUCHI**

(Received Oct. 1, 1979)

Abstract

There are a number of published empirical formulae for drift snow transport as a function of wind velocity. Comparing these formulae at the same wind velocity, however, results in considerable disagreement. It is hypothesized that the disparity arises from snow conditions and the various stages of development of drifting snow.

The horizontal distribution of drift flux was measured with snow traps along a transect parallel with the wind, beginning at an upwind boundary that served as the starting point of drifting snow. Results indicate that drift snow transport cannot be defined uniquely unless the drifting snow attains equilibrium (i.e., the snow profile is saturated).

Saltation of snow particles is thought to prevail near the snow surface. However, the vertical flux profile of saltating snow has never been measured. Vertical profiles of drift flux from the snow surface to a height of 30 cm were measured at nine levels, using snow traps composed of nine streamers (compartments). It appears that the saltation flux prevails up to a height of 7–9 cm above the surface, and the suspension flux gradually takes over as the drifting snow develops.

1. Introduction

Studies of drift snow transport have been undertaken by many investigators concerned with the practical problems of drifts and the mass budget of ice caps. There are a number of published empirical formulae relating drift snow transport to wind velocity, derived from actual measurements. Comparing these formulae at the same wind velocity, however, results in considerable disagreement even if differences in the design of snow traps and other measuring conditions are taken into account. These disagreements

* Paper presented for the Symposium on Snow in Motion, 12–17 August 1979, Fort Collins, Colorado. Sponsored by USDA Forest Service and International Glacio logical Society.

** Civil Engineering Research Institute, Hokkaido Development Bureau, Sapporo, Japan 062.

can be considered to arise from snow conditions and various stages of development of drifting snow.

Drift snow transport can be considered to increase from the starting point (some upwind boundary) until saturated drift snow transport has been reached. A knowledge of how snow drift transport increases with distance from a boundary is important in the design of snow protection installations. To determine how the drift snow transport increases, we measured the horizontal distribution leeward from a boundary where drifting snow started.

The vertical distribution of suspended drift snow density formulated by Shiotani[1] was based on turbulent diffusion theory and was consistent with numerous field tests. The vertical distribution of saltating drift snow density, however, has never been determined. We therefore undertook studies to determine vertical profiles of snow concentration in the layer where saltation is thought to prevail, and the changes in these profiles as the drift developed downwind from a boundary.

2. Expressions for drift snow transports

In the study of blowing sand, Bagnold[2] introduced equation (1) for the mass flux of sand in saltation based on the assumption that the shear stress working on the sand surface is equal to the difference in momentum between the sand rising from the surface and the sand dropping onto the surface:

$$Q_{sal} = C \frac{\rho}{g} U_*^3 \tag{1}$$

where Q_{sal} is the mass flux of sand, C is a proportionality constant, ρ is air density, g is gravitational acceleration, and U_* is the shear velocity.

As the saltating snow is similar in transport mechanism to saltating sand, equation (1) holds for drifting snow as well. The vertical profile of suspended drift snow is given by the expression published by Shiotani and Arai.

$$\frac{n}{n_0} = \left(\frac{Z_0}{Z}\right)^{w/kU_*} \tag{2}$$

where n is drift density. Z is height above the snow surface, w is particle fall velocity, k is von Karman's constant, and the subscript (o) denotes conditions at height z_0.

Drift snow transport is given by the integral of the product of drift density and the wind velocity U

$$Q = \int_{Z_0}^{Z} n(Z)\, U(Z)\, dZ \tag{3}$$

While a portion of the total drift is also transported by snow creep, most of the snow drift transport takes place by saltation and suspension. In the above theoretical equations, C in equation (1) and n_0 and w in equation (2) change with snow conditions, temperature, and shape and size of the drift particels. For this reason, the direct measurements of wind velocity alone do not allow use of the theoretical equation. For this reason, snow drift transport has been expressed by empirical formulae based on actual measurements (Table 1). As illustrated in Figure 1, there are extremely large differences in the drift transport for the same velocity. These differences are presumed to arise from the different snow conditions and stages of drift development rather than from the measurement methods employed by researchers.

Table 1. Empirical formulae for snow transport, Q ($g \cdot m^{-1} \cdot s^{-1}$), as a function of wind speed at 1 m height, U_{100}(m/s). Snow transport is the total up to the 2 m height unless otherwise noted.

No.	Investigator	Formula
1	Khrgian (Dyunin[3])	$Q=-5.8+0.267U_{100}+0.123(U_{100})^2$
2	Inanov (Dyunin[3])	$Q=0.0295(U_{100})^3$
3	Mel'nik (Dyunin[3])	$Q=0.092(U_{100})^3$
4	Dyunin[3]	$Q=0.0334(1-(4/U_{100}))(U_{100})^3$
5	Komarov[4]	$Q=0.011(U_{100})^{3.5}-0.67$
6	Budd and others[5]	
	a. (Q for 300 m height)[a]	$\log Q=1.18+0.1080U_{100}$
	b. (Q for 2 m height)[b]	$\log Q=1.22+0.0859U_{100}$
7	Kobayashi and others[6]	$Q=0.03(U_{100})^3$

[a]Derived from the equation for Q over 300 m, $\log Q=1.18+0.0887\ U_{1000}$, assuming $z_0=0.246$ mm and $U_*=0.0377\ U_{1000}$.
[b]From regression analysis of data in Budd and others[5], Appendix E.

3. Increase in drift snow transport with drift fetch

Drift transport increases with distance from an upwind boundary until it reaches a maximum (or saturated) value that is determined by the wind velocity. The stages of a developing snow drift transport can theoretically include thousands of variations even for the same wind velocity and snow conditions. We therefore measured how the drift transport increased with distance downwind from its starting point.

The measurements were carried out near the mouth of the Ishikari River in

Hokkaido. The 300 m wide ice-free river provided an effective boundary across which snow transport was negligible. The Sea of Japan is about 500 m upwind of the river, and the intervening area contains some houses and scattered trees about 3 m in height. Downwind of the river, the land is a nearly level reed swamp, completely snow-covered at the time of this experiment.

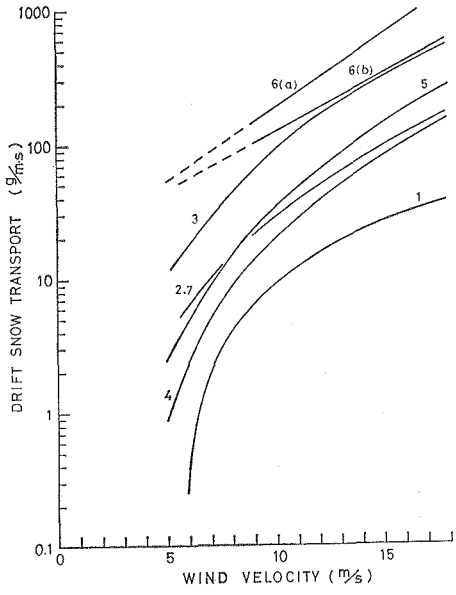

Fig. 1 Empirical formulae for drift snow transport as a function of wind velocity, as given in Table 1.

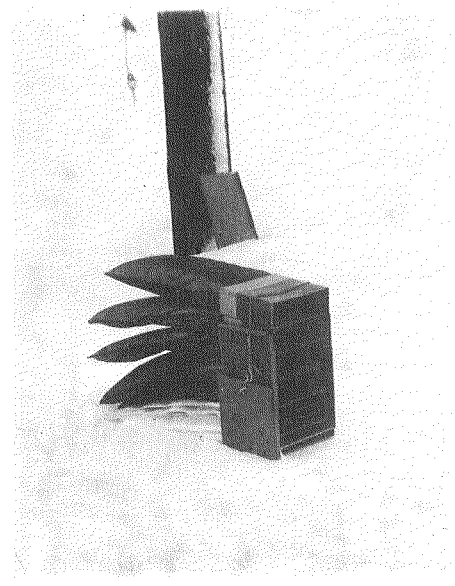

Fig. 2 The snow trap used in the study.

Three to four snow traps were placed at distances of 20 to 700 m downwind from the riverside and were measured simultaneously. The snow trap consisted of a stack of open-ended boxes (streamers), each 15 cm wide and with a bag attached on one end (Figure 2). Inlet height of the five lowest sections was 2 cm, and 5 cm for the upper four compartments, allowing measurement of the vertical profile of drift flux up to 30 cm above the snow surface. As shown in Figure 3, the aerodynamic efficiency of the streamers, as determined in wind tunnel tests, decreases as the height from the surface, because of overlap among bags. To allow correction for aerodynamic efficiency, wind profiles were measured during each run using 5 anemometers installed

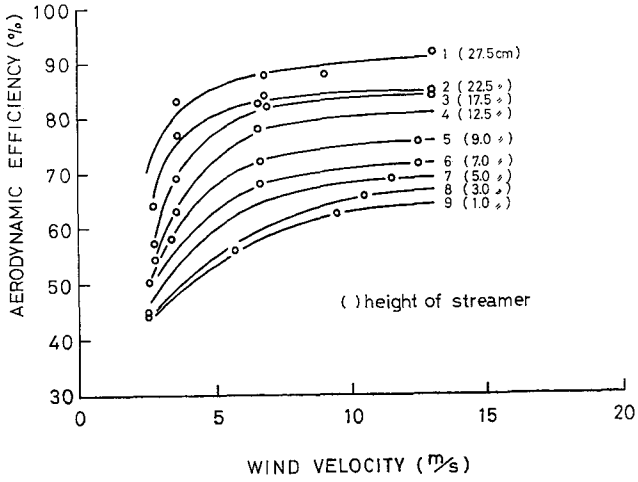

Fig. 3 Aerodynamic efficiency curves of the snow traps, as determined in wind tunnel tests. Wind velocities are the free-flow values at heights corresponding to the center of the intake opening.

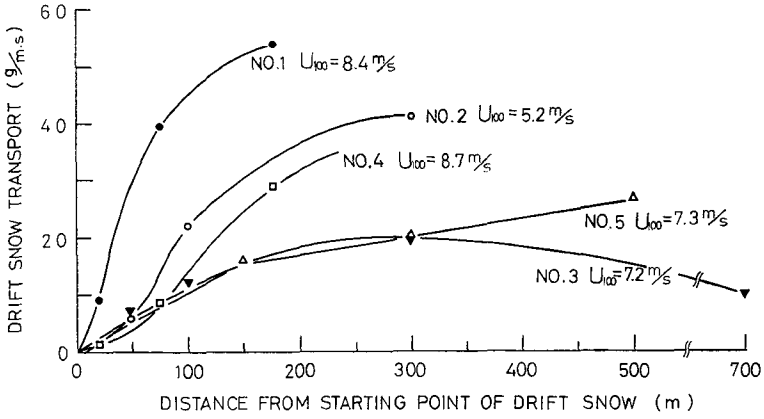

Fig. 4 Horizontal distribution of drift snow transport as measured with snow traps.

at heights up to 6 m above the snow surface.

Figure 4 shows the horizontal distribution of drift transport measured in this study. Drift transport increases sharply from the boundary up to 150–200 m downwind, and continues to increase even at distances over 300 m. The drift transport increases until the saturated state is achieved, which depends on the prevailing wind velocity and snow conditions. In the area of

increasing drift transport, the snow surface may be eroded, and falling snow does not accumulate on the surface. When the saturated drift transport is exceeded, a part of the drift settles out and accumulates. Thus, it is possible to find erosion areas as well as accumulation areas by measuring the snow depth. The distance from the boundary to an accumulation area can be considered the average distance over which the drift transport increases.

Figure 5 shows the horizontal distribution of the snow depth measured with a snow sampler from the riverside to a point 700 m downwind at the same locations where the drift transport was measured. This data illustrates that the snow depth increases from the river bank to 200 m, remains practically constant from 200 m to 350 m, and increases again from 350 to 500 m. The area up to 200 m from the river is assumed to be an erosion area where the drift increases, characterized by eroded snow and drifted falling snow.

In the region 200–350 m from the river, the erosion is limited and the drift is simply increased by falling snow. Beyond 350 m is the accumulation area where the drift transport is saturated. Bearing in mind that the drift increases beyond the saturated value only during heavy snowfall in Hokkaido, a longer fetch distance is considered necessary to achieve a saturated drift transport in the abence of snowfall.

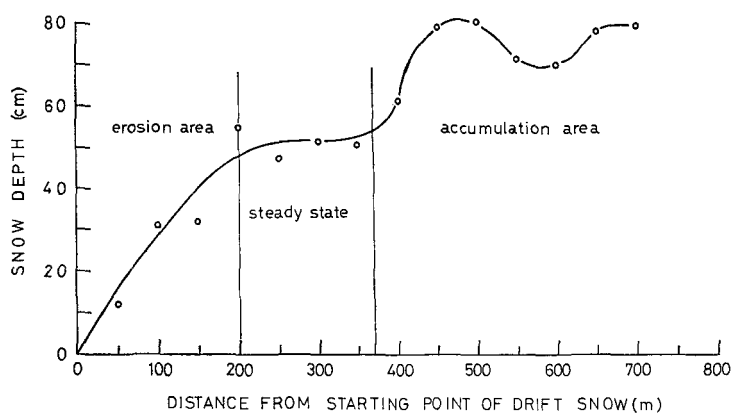

Fig. 5 Horizontal distribution of snow depth downwind from boundary.

4. Snow quality and saturated drift transport

The constant in equation (1) as well as n_0, Z_0 and w in equation (2) change with snow conditions. If we assume sufficiently developed drifting snow in a

saturated state, for a given wind velocity these variables depend on the resistance the snow surface offers to drifting snow particles. The shear stress of the wind ($\tau_0=\rho U_*^2$) acting on the snow surface can be considered as also acting on the number of snow particles, N, impacting on the surface. If the coefficient of friction between the snow surface and the drifting snow particles is given by tan ϕ, the equation of motion for one snow particle becomes

$$\frac{\tau_0}{N}=C'\frac{\pi d^3}{6}(\sigma-\rho)g\tan\phi \tag{4}$$

where C' is a proportionality constant, d is spherical diameter of the particle, σ is particle density, ρ is air density, and ϕ is the angle of friction. This relation holds basically only for surface creep but can include saltation. When snow particles collide with the surface during surface creep and saltation, they continue their motion if the kinetic energy of the drifting particles at the time of the impact is larger than the friction force at the snow surface, but stop if the kinetic energy is smaller than the friction. We ignore here the adhesion of snow particles. As the surface creep and saltation increase, the drift flux due to suspension also increases.

Assuming that the friction between the drifting particles and the surface influences the drift flux, we measured the coefficient of friction. For this determination, we cut off a 20 cm×20 cm piece of the snow surface and supported it at different angles from the horizontal. Snow particles were then dropped from a height of 5 cm above the inclined snow surface, and the angle where the snow particles started to roll on the surface was measured as an angle of friction. These observations were carried out at the same place where the drift was measured.

The coefficient of friction between the snow surface and particles is shown in Figure 6 as a function of snow quality and temperature. The snow type is classified into fresh snow having a particle diameter 0.5–3 mm, settled dry snow 0.1–1 mm in diameter, and old firm snow 2–4 mm in diameter. As the figure shows, the coefficient of friction decreases in the following order: fresh dry snow, settled dry snow, and old firm snow. As the surface resistance to the drifting snow increases, the saturated drift flux decreases and deposition results. Where a surface of old firm snow adjoins that of a fresh dry snow, for example, the drift is swept over the old firm snow and deposition in the form of small dunes occurs on the surface of the fresh dry snow.

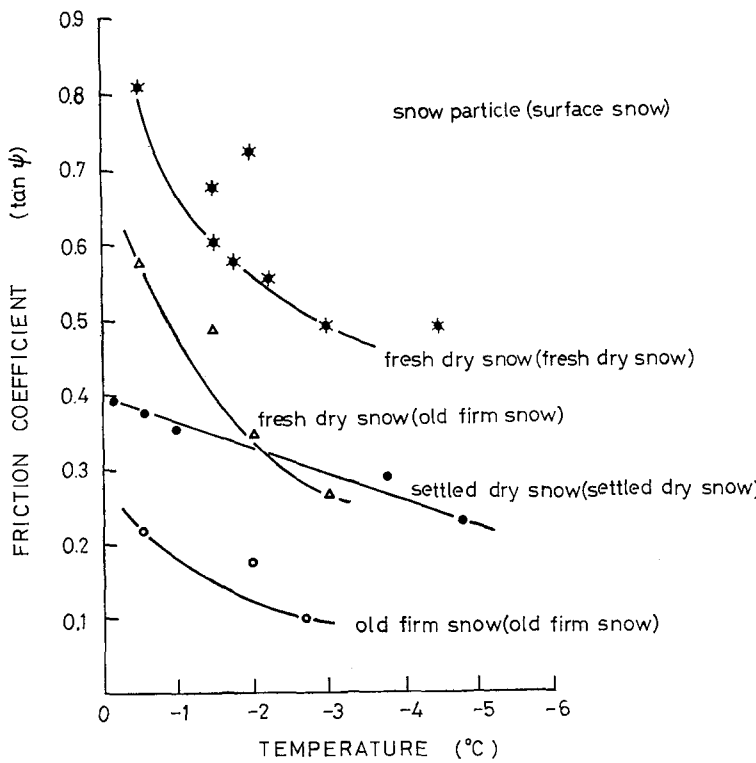

Fig. 6 The coefficient of friction of snow surface and particles as a function of snow quality and temperature.

Because the drift flux can not be uniquely defined until the drifting snow fully develops, we define the amount of drift in terms of the saturated drift transport. This definition allows us to formulate a reliable empirical equation for the drift snow transport. Because the saturated drift flux changes with snow conditions, it is necessary to observe the quality of the snow surface whenever the saturated drift is recorded. After the saturated drift has been reached, a part of the drift accumulates on the surface. Thus, the saturated drift transprot is defined as that associated with deposition. These measurements were carried out at a location 350 m downwind from the river boundary.

Table 2 shows the drift snow transport measured directly up to 30 cm over the surface (Q_0^{30}), and that calculated from the vertical profile of the drift flux up to 2 m over the surface (Q_0^{200}). Also shown are wind velocity at the 1 m height (U_{100}), snow temperature and snow quality. The values of (Q_0^{200})

Table 2. Saturated drift snow transport, wind velocity at 1 m height (U_{100}) temperature and surface snow as measured in this study. Q_o^{30} is the drift transport measured between the surface and a height of 30 cm. Q_o^{200} is the drift transport between the surface and a height of 200 cm, estimated by extrapolating the vertical profile of the mass flux.

Drift snow transport ($g \cdot m^{-1} \cdot s^{-1}$)		Wind velocity, U_{100}	Temperature (°C)	Surface snow
Q_o^{30}	Q_o^{200}			
55.8	58.0	777	−3.7	Old firm snow
36.7	36.8	955	−2.9	Settled dry snow
8.5	8.8	690	−2.0	Settled dry snow
14.0	16.8	950	−5.0	Fresh dry snow
161.7	164.0	1135	−1.3	Old firm snow
90.2	94.0	1210	−1.3	Settled dry snow
86.2	94.0	1230	−1.6	Settled dry snow

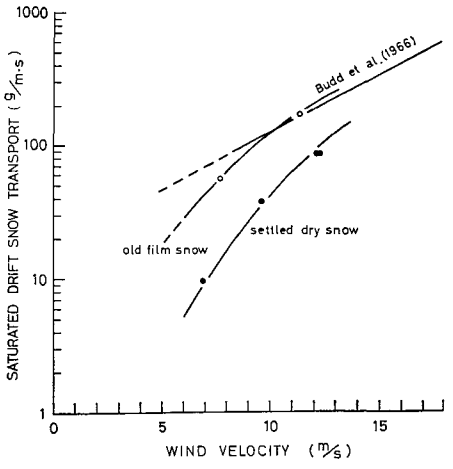

Fig. 7 Saturated drift snow transport between the surface and a height of 200 cm. Wind velocity is that at 1 m height.

are plotted in Figure 7 along with empirical equations derived from data published by Budd and others[5], which provide the largest estimates for drift transport of all other investigators. The drift transport values obtained by Budd and others[5] at Byrd Station are similar to our values. The results of our measurements are given by the empirical formulae

$$Q_0^{200} = 0.2\,(U_{100})^{2.7} \qquad \text{(old firm snow)} \tag{5}$$

$$Q_0^{200} = 2.9\cdot 10^{-3}(U_{100})^{4.16} \qquad \text{(settled dry snow)} \tag{6}$$

where Q is in $g\cdot m^{-1}\cdot s^{-1}$ and U is in m/s.

5. Vertical profile of drift snow transport

Vertical profiles of drift transport measured by Budd and others[5], Shiotani[1], and other investigators primarily represented suspension due to turbulent diffusion. We therefore measured the vertical profile of the drift flux near the surface where saltation is considered dominant.

Kawamura[7] assumed that saltating particles leave the surface vertically, and that the vertical velocity component of particle, leaving the randomly oriented surface obeys Maxwell's law of statistical distribution. He therefore proposed that the vertical profile of saltating particle flux could be given by

$$q(Z) = 2N_0\left(\frac{\alpha U}{g}\right)\exp\left(-\frac{Z}{\pi h_0}\right) \tag{7}$$

where N_0 is the number of particles leaving the surface per unit area in unit time, α is a constant equal to $3\pi\mu d$ (μ is the viscosity coefficient of air and d is particle diameter), h_0 is the average maximum trajectory height of particles, and U is wind velocity.

The vertical profile of the drift flux due to suspension is obtained by multiplying the density of the drift (n) from equation (2) by the wind velocity. The logarithmic wind profile is usually written in the form

$$U(Z) = \frac{U_*}{k}\ln\left(\frac{Z}{Z_0}\right) \tag{8}$$

From equations (2) and (8)

$$q(Z) = n(Z)\,U(Z) = \frac{U_*}{k}\ln\left(\frac{Z}{Z_0}\right)n_0\left(\frac{Z_0}{Z}\right)^{w/kU_*} \tag{9}$$

Equation (7) indicates that the logarithm of the drift flux, $\ln q(Z)$, is linearly related to the height Z, whereas equation (9) means that the logarithm of the drift flux, $\ln q(Z)$, and the logarithm of height, $\ln Z$, are linearly related except for points very near Z_0.

In Figure 8, the measured vertical profiles of the drift flux are plotted with logarithmic scales. Symbols I and IV denote old firm snow, II and III

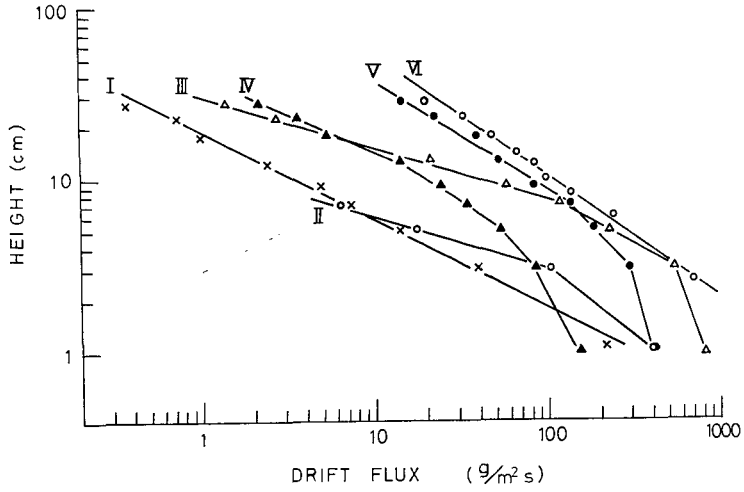

Fig. 8 Logarithmic profile of drift flux.

are for settled dry snow, and IV and V represent fresh dry snow. As shown in the figure, for old firm snow the data plot linearly until close to the surface, but curvature is much more pronounced near the surface of settled dry snow and fresh dry snow. When these latter data are plotted with a logarithmic scale for drift flux and a linear scale for height (Figure 9), the curves near the surface become straight lines having a bend at 7–9 cm. Line II corresponds to drifting snow due to saltation alone, and shows that the vertical profile of saltation agrees with the prediction of Kawamura[7].

Line III has a bend at a height of 22 cm in addition to the one at 8 cm. Close examination of Figure 8 reveales that line III is straight above a height of 8 cm. Thus, the bend indicated at 22 cm on the semilogarithmic plot is an artifact. The same holds true for line IV. The portion of lines IV and V with heights below the bend at 7–9 cm height should correspond to the top of the layer where saltation is predominant.

We also investigated the change in vertical distribution of the drift flux during development of drifting snow. Simultaneous measurements were carried out at locations 20, 75, and 175 m from the riverbank where the drift originates.

Figure 10 shows the results for a surface of old firm snow. At the 75 m location, the drift flux continues to increase, while the vertical profile of drift flux close to the surface is already a straight line when plotted on logarithmic

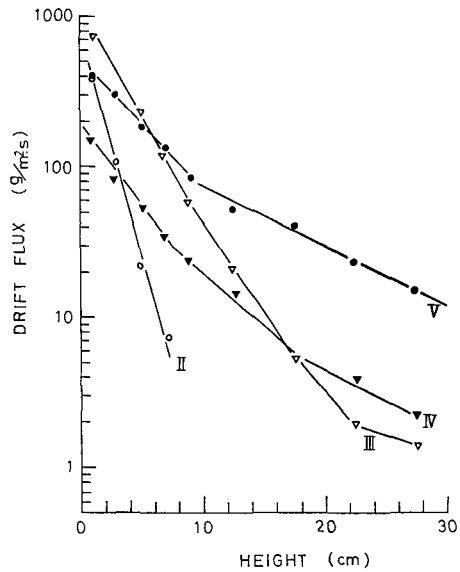

Fig. 9 Semi-logarithmic profile of drift flux.

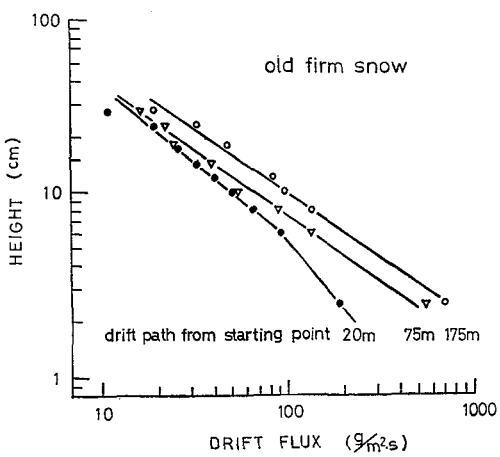

Fig. 10 Vertical profiles of drift flux simultaneously observed over an old firm snow surface at three distances from the boundary.

paper. Figure 11 corresponds to a surface of settled dry snow, and shows that during the development of drifting snow, the linear relationship on a logarithmic scale extends to very near the surface. A drifting snow is considered to start with creep and saltation, developing into suspension.

The development is influenced by the resistance offered by the snow surface. For example, it is likely that the turbulent diffusion reaches the surface relatively quickly when the surface has little resistance as is the case for old firm snow. As the drifting snow develops, the vertical profile of drift flux apparently becomes that of suspension drift down to near the snow surface. This is because the amount of saltation drift is negligible compared to the suspension drift when the drifting snow is fully developed.

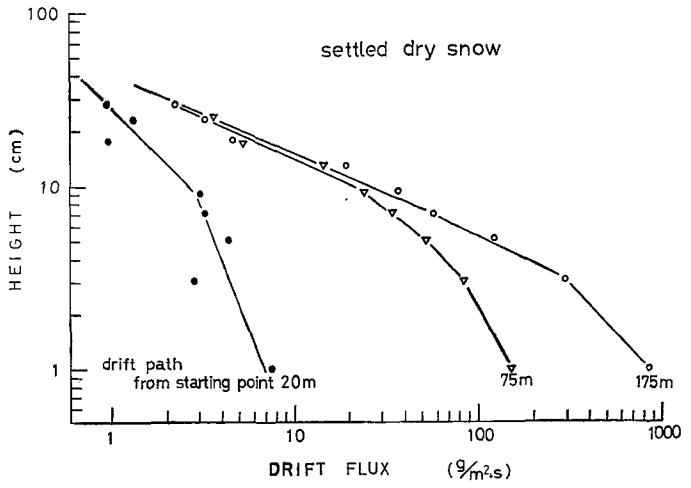

Fig. 11 Vertical profiles of drift flux simultaneously observed over a settled dry snow surface at three distances from the boundary.

6. Conclusions

A number of very different empirical equations for drift snow transport have been published. The studies reported here suggest that the drift snow transport cannot be defined uniquely unless it has developed into the saturated drift, as determined by measuring the drift flux simultaneously at several locations downwind from a boundary. Our results also show that the drift transport depends on quality of the snow as it affects particle motion. These results help explain the large differences among the published empirical formulae for drift transport.

Measurements of the horizontal distribution of the drift flux downwind from a boundary show that the drift flux increases rapidly in the region up to 150–200 m from the start of he drifting snow, and that it continues to

increase to more than 300 m. Based on measurements of snow depth at various distances downwind from a boundary, it is estimated that the drift flux reaches saturation about 350 m downwind from the starting point.

Measurements of the vertical profile of the drift flux up to a 30 cm height suggested that, for a surface of settled dry snow or fresh dry snow, the saltation flux prevails up to a height of 7–9 cm, and the suspension flux gradually takes over as the drifting snow developes.

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