

## VERTICAL PROFILE AND HORIZONTAL INCREASE OF DRIFT-SNOW TRANSPORT

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**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 up-wind 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.

**RÉSUMÉ.** *Profil vertical et accroissement horizontal pour le transport de la neige par le vent.* Il existe un grand nombre de formules empiriques publiées pour le transport de la neige en fonction de la vitesse du vent. Cependant, en comparant ces formules pour la même vitesse de vent, on trouve de considérables discordances. On émet l'hypothèse que ces différences proviennent de l'état de la neige et des différents stades de développement de la congère.

La distribution horizontale du flux de neige a été mesurée à l'aide de trappes à neige le long d'un alignement parallèle au vent, commençant en un point à l'amont considéré comme le point d'origine du drift. Les résultats montrent qu'on ne peut définir le transport de la neige soufflée de manière unique à moins que le flux de neige n'ait atteint son point d'équilibre (c'est-à-dire que le profil de neige est saturé).

On pense que la saltation des particules de neige est le phénomène qui prévaut près de la surface de la neige. Cependant, les profils verticaux de la neige en saltation n'ont jamais été mesurés. Des profils verticaux de drift depuis la surface de la neige jusqu'à une hauteur de 30 cm ont été mesurés à neuf niveaux grâce à des trappes à neige composées de neuf compartiments. Il apparaît que le flux de saltation prédomine jusqu'à une hauteur de 7 à 9 cm au-dessus de la surface et que le flux en suspension prend progressivement le pas quand le transport de neige se développe.

**ZUSAMMENFASSUNG.** *Vertikalprofil und Horizontalzunahme des Transportes von Driftschnee.* Es gibt eine Anzahl von Veröffentlichungen mit empirischen Formeln für die Schneedrift als Funktion der Windgeschwindigkeit. Vergleicht man jedoch diese Formeln bei gleicher Windgeschwindigkeit, so ergeben sich beträchtliche Abweichungen. Es wird behauptet, dass die Unstimmigkeiten von der Beschaffenheit des Schnees und von den verschiedenen Stadien der Driftentwicklung herrühren.

Die horizontale Verteilung des Driftflusses wurde mit Schneefallen längs eines Profils parallel zum Wind gemessen; der Anfangspunkt lag an einer Grenze in Richtung gegen den Wind, die als Ausgangspunkt der Schneedrift diente. Die Ergebnisse zeigen, dass der Transport von Driftschnee nicht eindeutig zu definieren ist, bevor die Schneedrift einen Gleichgewichtszustand erreicht (d.h. das Schneeprofil gesättigt ist).

Nahe der Schneeoberfläche dürften Sprungbewegungen der Schneepartikel vorherrschen. Das vertikale Flussprofil springenden Schnees wurde jedoch noch nie gemessen. Vertikalprofile des Driftflusses von der Schneeoberfläche bis 30 m Höhe wurden mit Schneefallen, die aus neun Kammern zusammengesetzt waren, in neun Niveaus gemessen. Es scheint, dass der Sprungfluss bis zu einer Höhe von 7–9 cm über der Oberfläche vorherrscht, während der Suspensionsfluss allmählich mit der Entwicklung der Schneedrift einsetzt.

### INTRODUCTION

Studies of drift-snow transport have been undertaken by many investigators concerned with the practical problems of drifts and the mass balance 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 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 up-wind boundary) until saturated drift-snow transport has been reached. A knowledge of how drift-snow 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 (1953) 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 down-wind from a boundary.

#### EXPRESSIONS FOR DRIFT-SNOW TRANSPORT

In his study of blowing sand, Bagnold (1941) introduced an equation 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:

$$Q_{\text{sal}} = \frac{C\rho U_{\star}^3}{g}, \quad (1)$$

where  $Q_{\text{sal}}$  is the mass flux of sand,  $C$  is a proportionality constant,  $\rho$  is air density,  $g$  is gravitational acceleration, and  $U_{\star}$  is the shear velocity.

As saltating snow is similar in transport mechanism to saltating sand, Equation (1) should hold for drifting snow as well. The vertical profile of suspended drift snow is given by the expression published by Shiotani and Arai (1953).

$$\frac{\eta}{\eta_0} = \left(\frac{z_0}{z}\right)^{w/kU_{\star}}, \quad (2)$$

where  $\eta$  is drift density,  $z$  is height above the snow surface,  $w$  is particle fall velocity,  $k$  is von Kármán's constant, and the subscript 0 denotes conditions at height  $z_0$ .

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

$$Q = \int_{z_0}^z n(z) U(z) dz. \quad (3)$$

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  $\eta_0$  and  $w$  in Equation (2) change with snow conditions, temperature, and shape and size of the drift particles. For this reason, direct measurements of wind velocity alone do not allow the use of the theoretical equation. For this reason, snow-drift transport has been expressed by empirical formulae based on actual measurements (Table I). 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 the different researchers.

#### INCREASE IN DRIFT-SNOW TRANSPORT WITH DRIFT FETCH

Drift transport increases with distance from an up-wind 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 down-wind from its starting point.

TABLE I. EMPIRICAL FORMULAE FOR SNOW TRANSPORT  $Q$  ( $\text{g m}^{-1} \text{s}^{-1}$ ) AS A FUNCTION OF WIND SPEED AT 1 m HEIGHT,  $U_{100}$  ( $\text{m s}^{-1}$ ). SNOW TRANSPORT IS THE TOTAL UP TO A HEIGHT OF 2 m UNLESS OTHERWISE NOTED

No.	Investigator	Formula
1	Khrgian (Dyunin, 1954)	$Q = -5.8 + 0.267U_{100} + 0.123(U_{100})^2$
2	Ivanov (Dyunin, 1954)	$Q = 0.0295(U_{100})^3$
3	Mel'nik (Dyunin, 1954)	$Q = 0.092(U_{100})^3$
4	Dyunin (1954)	$Q = 0.0334[1 - (4/U_{100})](U_{100})^3$
5	Komarov (1954)	$Q = 0.011(U_{100})^{3.5} - 0.67$
6	Budd and others (1966)	
	a. ( $Q$ for 300 m height)*	$\log Q = 1.18 + 0.1080U_{100}$
	b. ( $Q$ for 2 m height)†	$\log Q = 1.22 + 0.0859U_{100}$
7	Kobayashi and others (1969)	$Q = 0.03(U_{100})^3$

\* Derived from the equation for  $Q$  over 300 m,  $\log Q = 1.18 + 0.0887U_{1000}$ , assuming  $z_0 = 0.246$  mm and  $U_* = 0.0377U_{1000}$ .

† From regression analysis of data in Budd and others (1966), appendix E.

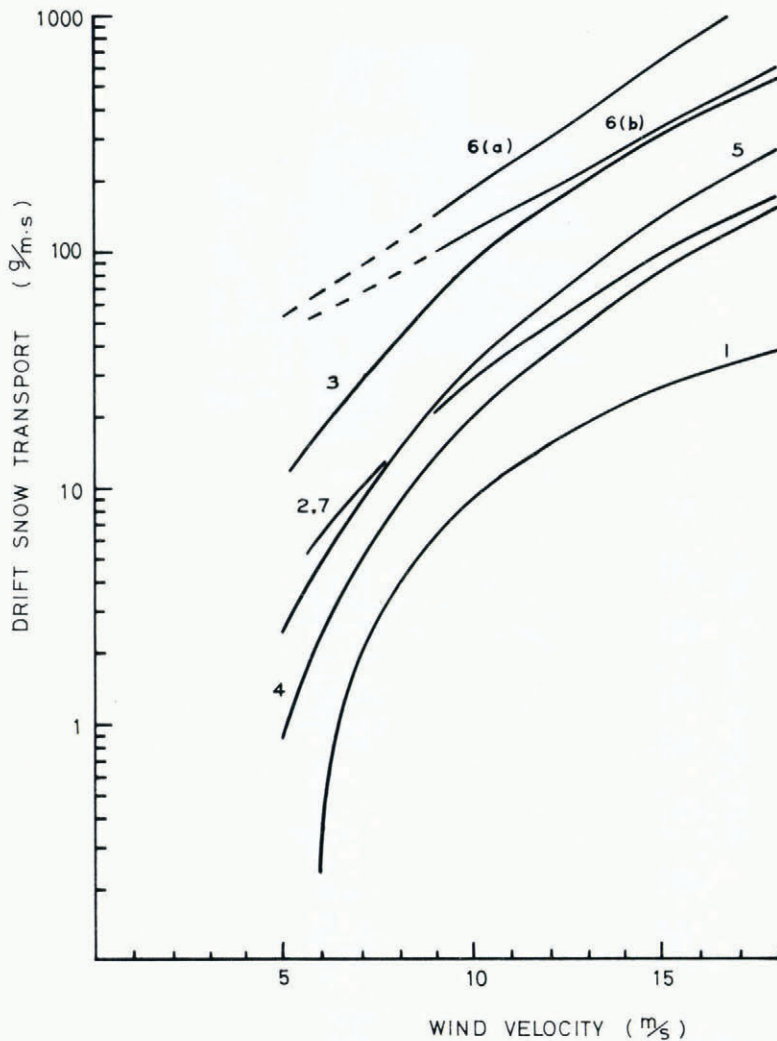


Fig. 1. Empirical formulae for drift-snow transport as a function of wind velocity, as given in Table I.

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 up-wind of the river, and the intervening area contains some houses and scattered trees about 3 m in height. Down-wind of the river, the land is a nearly level reed swamp, completely snow-covered at the time of this experiment.

Three to four snow traps were placed at distances of 20 to 700 m down-wind 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 (Fig. 2). The inlet heights of the five lowest sections were 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 five anemometers installed at heights up to 6 m above the snow surface.

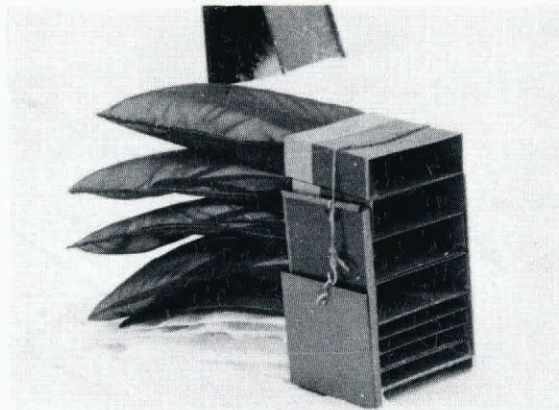


Fig. 2. The snow trap used in the study.

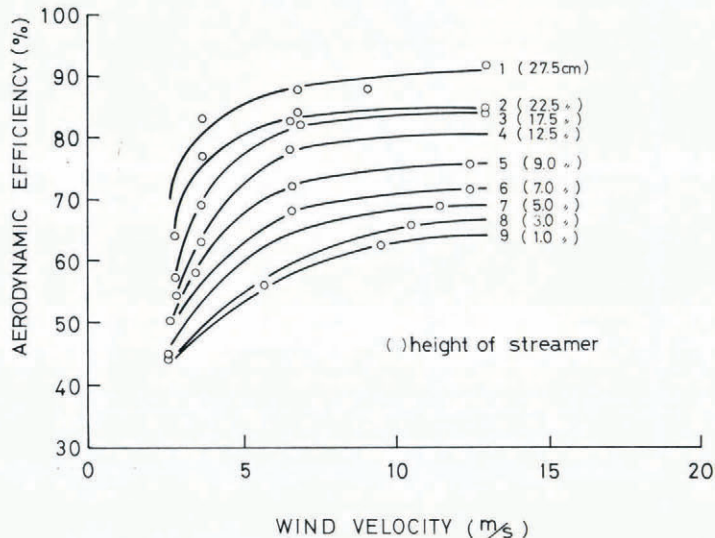


Fig. 3. Aerodynamic efficiency curves of the snow traps, as determined in wind-tunnel tests. Wind velocities are the free-flow values at heights corresponding to the centre of the intake opening.

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 down-wind, 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 down-wind at the same locations at which the drift transport was measured. These data illustrate 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.

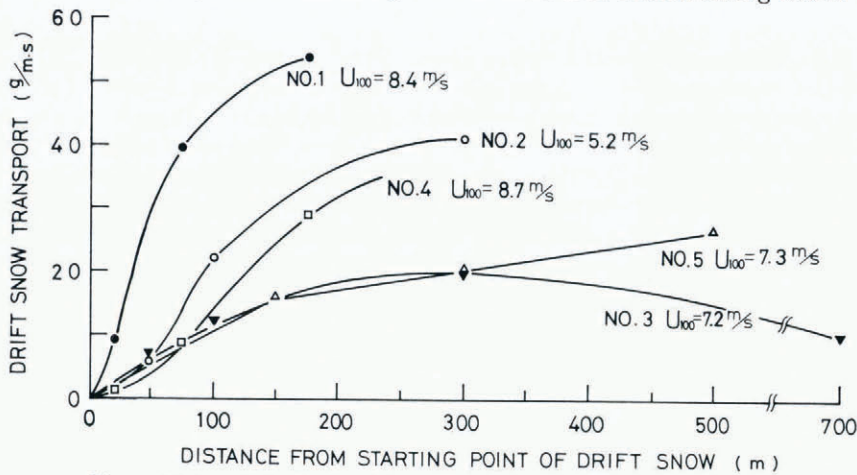


Fig. 4. Horizontal distribution of drift-snow transport as measured with snow traps.

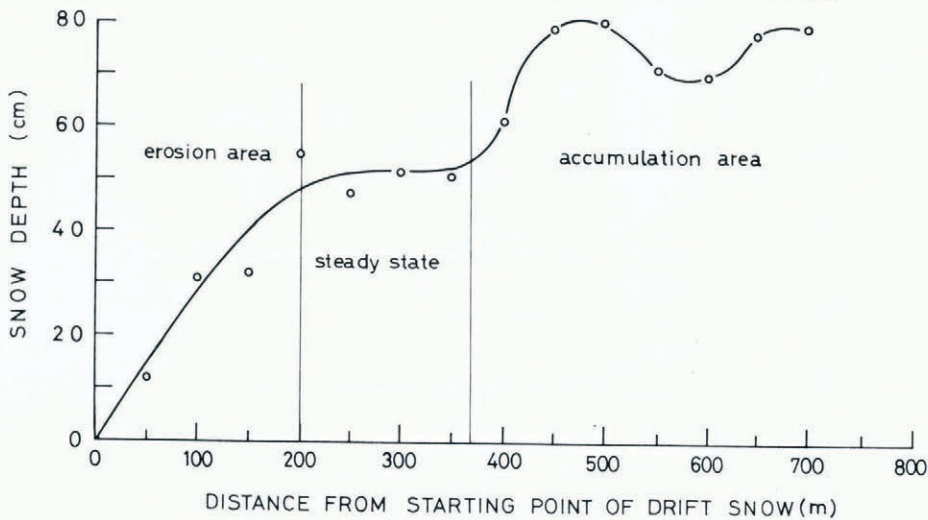


Fig. 5. Horizontal distribution of snow depth downward from boundary.

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 snow-fall in Hokkaido, a longer fetch distance is considered necessary to achieve a saturated drift transport in the absence of snow-fall.

#### 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  $N$  of snow particles impacting on the surface. If the coefficient of friction between the snow surface and the drifting snow particles is given by  $\tan \phi$ , the equation of motion for one snow particle becomes

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

where  $C'$  is a proportionality constant,  $d$  is spherical diameter of the particle,  $\sigma$  is particle density,  $\rho$  is air density, and  $\phi$  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.

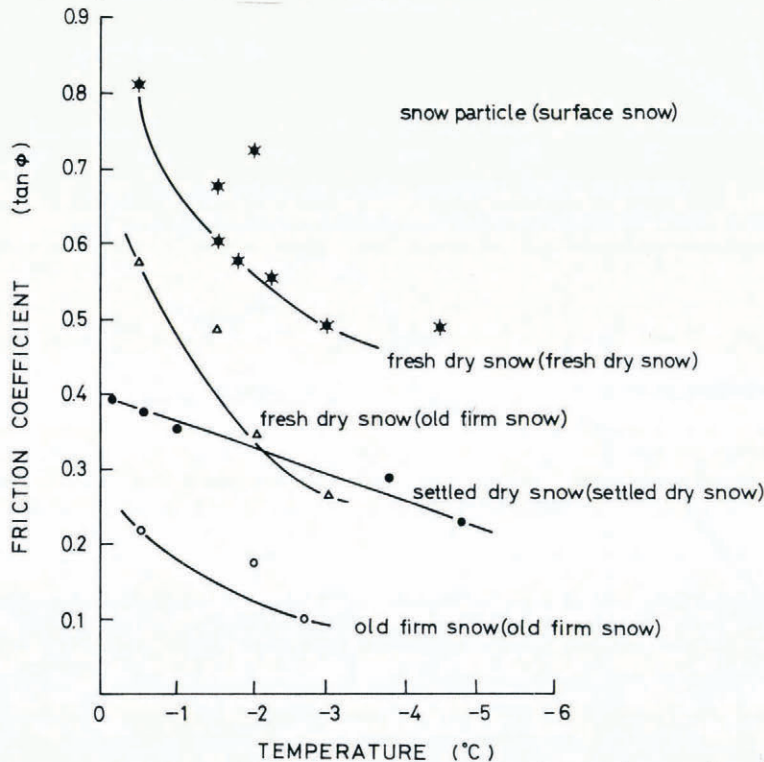


Fig. 6. The coefficient of friction of snow surface and particles as a function of snow quality and temperature.

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 at which 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.

Because the drift flux cannot 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 saturated drift has been reached, a part of the drift accumulates on the surface. Thus, saturated drift transport is defined as that associated with deposition. These measurements were carried out at a location 350 m downwind from the river boundary.

Table II 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  $Q_0^{200}$  are plotted in Figure 7 along with empirical equations derived from data published by Budd and others (1966), which provide the largest estimates for drift-transport of all the investigators. The drift-transport values obtained by Budd and others (1966) at Byrd Station are similar to our values. The results of our measurements are given by the empirical formulae

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

$$Q_0^{200} = 2.9 \times 10^{-3}(U_{100})^{4.16} \quad (\text{settled dry snow}), \quad (6)$$

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

TABLE II. SATURATED DRIFT-SNOW TRANSPORT, WIND VELOCITY AT 1 m HEIGHT ( $U_{100}$ ), TEMPERATURE, AND SURFACE SNOW AS MEASURED IN THIS STUDY.  $Q_0^{30}$  IS THE DRIFT TRANSPORT MEASURED BETWEEN THE SURFACE AND A HEIGHT OF 30 CM.  $Q_0^{200}$  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		Wind velocity	Temperature	Surface snow
$Q_0^{30}$	$Q_0^{200}$	$U_{100}$	$^{\circ}\text{C}$	
$\text{g m}^{-1} \text{s}^{-1}$	$\text{g m}^{-1} \text{s}^{-1}$	$\text{m s}^{-1}$		
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	1 135	-1.3	Old firm snow
90.2	94.0	1 210	-1.3	Settled dry snow
86.2	94.0	1 230	-1.6	Settled dry snow

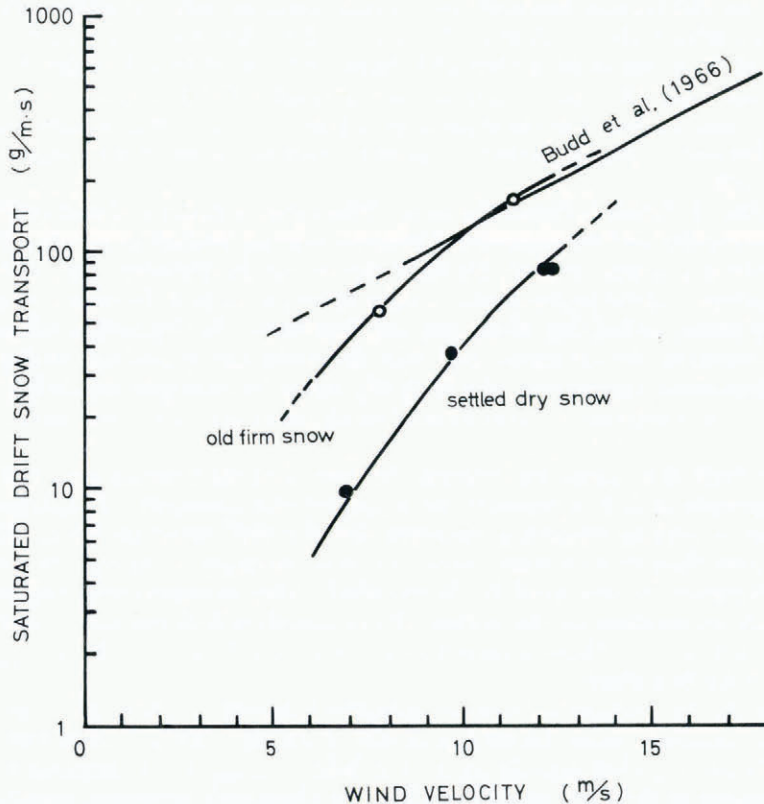


Fig. 7. Saturated drift-snow transport between the surface and height of 200 cm. Wind velocity is that at 1 m height.

#### VERTICAL PROFILE OF DRIFT-SNOW TRANSPORT

Vertical profiles of drift transport measured by Budd and others (1966), Shiotani (1953), 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 (1948) assumed that saltating particles leave the surface vertically, and that the vertical velocity component of a 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

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

where  $N_0$  is the number of particles leaving the surface per unit area in unit time,  $\alpha$  is a constant equal to  $3\pi\mu d$  ( $\mu$  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  $\eta$  of the drift from Equation (2) by the wind velocity. The logarithmic wind profile is usually written in the form

$$U(z) = \frac{U_*}{k} \ln \left( \frac{z}{z_0} \right). \quad (8)$$



From Equations (2) and (8)

$$q(z) = \eta(z) U(z) = \frac{U_*}{k} \ln \left( \frac{z}{z_0} \right) \eta_0 \left( \frac{z_0}{z} \right)^{w/k U_*} \tag{9}$$

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 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 (Fig. 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 (1948).

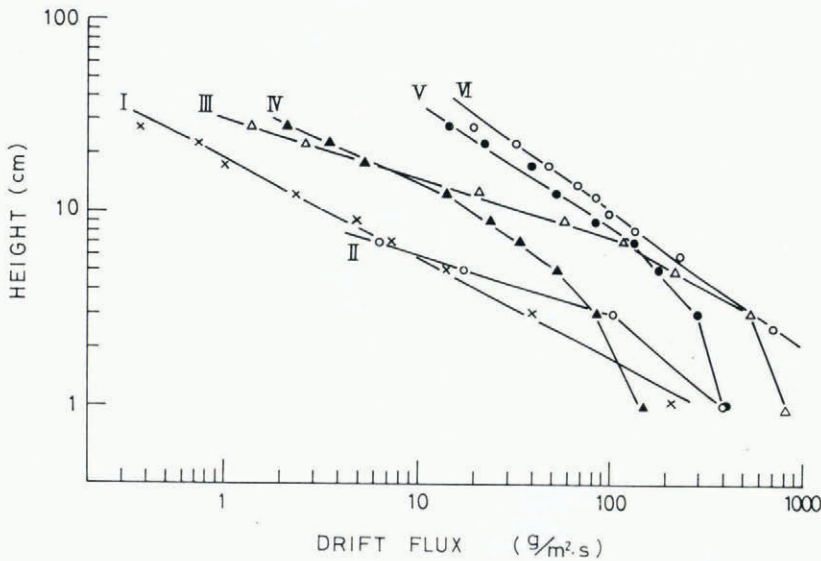


Fig. 8. Logarithmic profile of drift flux.

Line III has a bend at a height of 22 cm in addition to the one at 8 cm. Close examination of Figure 8 reveals that line III is straight above a height of 8 cm. Thus, the bend indicated at 22 cm on the semi-logarithmic 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 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. Drifting snow is considered to

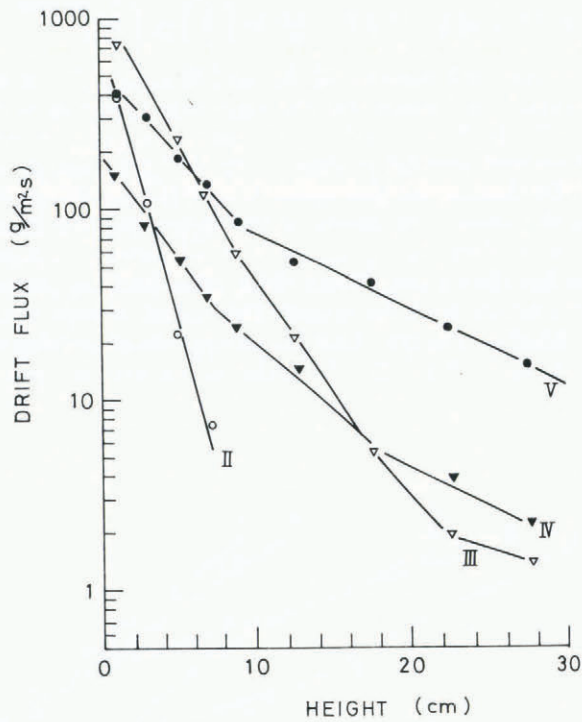


Fig. 9. Semi-logarithmic profile of drift flux.

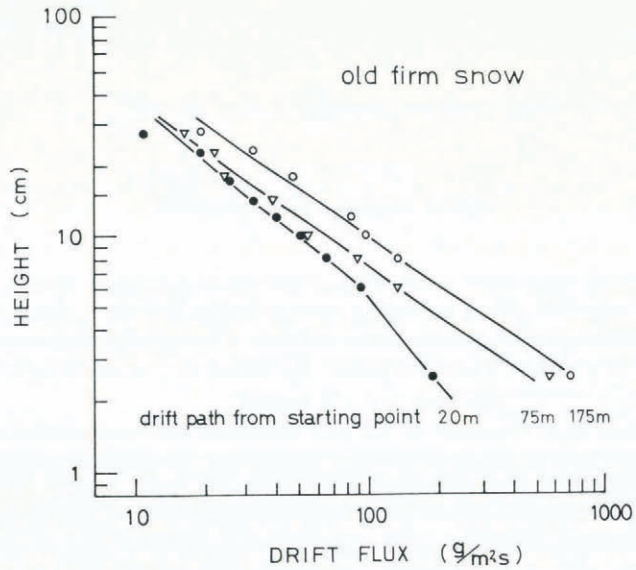


Fig. 10. Vertical profiles of drift flux simultaneously observed over an old firm snow surface at three distances from the boundary.

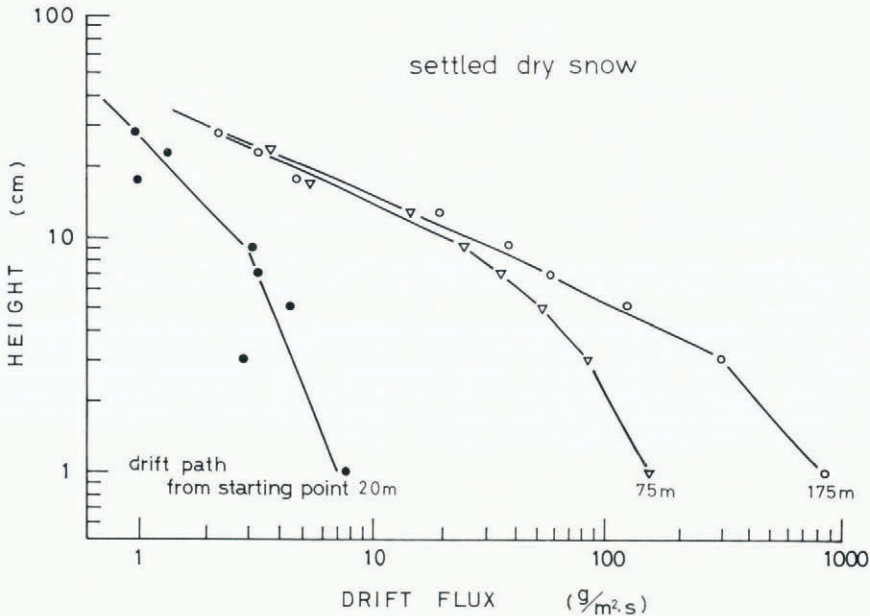


Fig. 11. Vertical profiles of drift flux simultaneously observed over a settled dry snow at three distances from the boundary.

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.

#### CONCLUSIONS

A number of very different empirical equations for drift snow transport have been published. The studies reported here suggest that drift-snow transport cannot be defined uniquely unless it has developed into saturated drift as determined by measuring the drift flux simultaneously at several locations down-wind from a boundary. Our results also show that the drift transport depends on the 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 down-wind from a boundary show that the drift flux increases rapidly in the region up to 150–200 m from the start of the drifting snow, and that it continues to increase to more than 300 m. Based on measurements of snow depth at various distances down-wind from a boundary, it is estimated that the drift flux reaches saturation about 350 m down-wind from the starting point.

Measurements of the vertical profile of the drift flux up to a height of 30 cm 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 develops.

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