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Physical Aspects of the Wind-Snow Interaction in Blowing Snow

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Abstract

New findings and concept concerned with the interaction of the wind and snow surface are presented. When a wind velocity on a snow surface becomes strong enough a phenomenon of blowing snow begins. The threshold velocity for blowing snow depends variously on many parameters such as temperature and textures of snow; The onset of blowing snow is essentially statistical in a sense that a nucleus of fluidization must be formed before the snow is fluidized or blowing snow appears.

The wind profile in blowing snow was found to be modified considerably by movements of snow particles; increase in the wind velocity was found near the snow surface where the motion of particles was maximum. This increase was attributed to the downward transfer of momentum in the horizontal direction due to saltating snow particles. Net downward force was found to act on saltating snow particles, which suggests the downward transport of momentum in the vertical direction.

1. Introduction

The interaction between wind and snow includes an extensive variety of physical processes such as wind structure over the snow surface,[1),2)] heat exchange at the atmosphere-snow interface[3),4)], and complex reliefs on snow surfaces[5)] (e.g., ripples, sastrugi, barchan, and various drift features near fences or buildings). Most of these works seem to have treated the snow surface as the one which is impermeable to air flow and acts as an energy sink because of its high reflectivity, and the movements of snow particles caused by the wind action have only been regarded as the cause for mass transport in the wind direction.

Recently Kobayashi and Ishida[6)] have suggested that the wind turbulence has a close relation with the motion of snow particles on the surface, some-

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times resulting in the formation of wavy patterns on the snow surface. There is a possibility that the interaction of wind and snow particles is more important than considered so far and may be the essential mechanism to determine the physical circumstances of the boundary layer including the atmosphere and the upper snow cover.

The present paper was intended to make clear the physical relations between the particle motion and air flow. The subjects discussed are the onset of fluidization and blowing snow, wind profile in blowing snow, dynamical behaviors of snow particles in the saltation layer, and momentum and energy transfer in fluidized snow. The paper deals with the subjects only briefly, so that more details will be found in each article.

2. Nucleation of fluidized snow and onset of blowing snow[7),8)]

When a velocity of wind blowing over a smooth snow surface exceeds some critical value, snow particles tend to be dislodged from the surface and set in motion, causing blowing snow. However, it is well known that actually the threshold wind velocity for blowing snow cannot be determined uniquely when the temperature and bulk density of the surface snow are known. This uncertainty seems to be caused by a fact that the onset of blowing snow is essentially a statistical phenomenon which is closely related with complex physical properties of snow including particle sizes, shapes, bonds, homogeneity, etc. This situation is similar to that of the formation of a crystal nucleus in a supersaturated solution, which will be explained below in more detail.

Fig. 1 shows the pressure loss of an upward air flow passing through a bed of snow particles in a cylindrical tube. At a higher temperature (–14.1°C, Fig. 1-a), the pressure loss increased with increasing air velocity (A→B), but the snow aggregate was not fluidized at considerably higher velocities. However, when a faint mechanical shock was artificially given to the tube, the aggregate instantly disintegrated into individual particles, initiating fluidization (C→D). On the other hand, at a lower temperature (–30.6°C, Fig. 1-b), the fluidized state of snow appeared spontaneously (C→D) when the air velocity reached a critical value, u_{mf}, that is the minimum fluidization velocity.

The difference in the manner of generation of fluidized snow at the two different temperatures is considered to be related with the increasing adhesive force between snow particles at higher temperatures. However, since nuclei of fluidization are required to be formed within snow before the whole snow is fluidized[7),8)] the above results can be understood as to correspond to

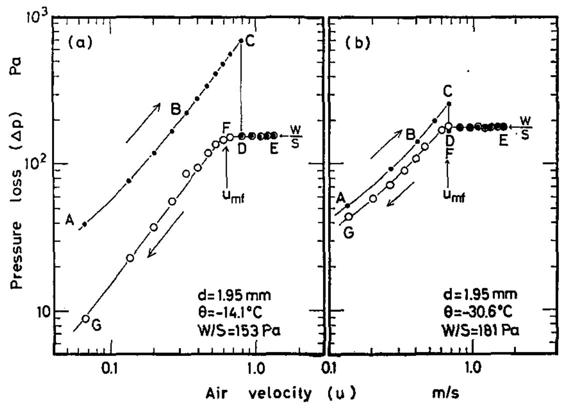

Fig. 1 Pressure loss (Δp) of air flow passing through a snow bed of average particle diameter 1.95 mm. Arrows indicate the order of experiments, and the solid and open circles give the pressure loss measured in increasing and decreasing air velocity, respectively. Details of the fluidization experiment are reported by Maeno and Nishimura.[7),8)]

heterogeneous and homogeneous nucleations in the crystal growth; when the kinetic energy is supplied to the snow aggregate from the air flow, the energy level of the snow increases, and some mechanically weak bonds in the snow tend to be broken locally. When the evergy level reaches some critical value, a nucleus of fluidization, which probably corresponds to a small area within snow broken mechanically, will be formed heterogeneously (e.g., by external mechanical shocks) or homogeneously (that is spontaneously), and grow in size to initiate fluidization of the whole snow.

The pressure loss at the point C in Fig. 1 corresponds to the critical excess energy to form a nucleus of fluidization in the snow. The threshold wind velocity of blowing snow is considered to be in principle the one at which the snow is energized to this level.

Fig. 2 gives the threshold wind velocity, u_{inc}, at which blowing snow begins to appear in the process of increasing wind velocity, and u_{dec} at which blowing snow ceases in the process of decreasing wind velocity. The experiment was conducted in a cold horizontal wind tunnel with a working cross-section of 0.5 m×0.5 m and length of 8.0 m, the whole bottom of which was covered with uniform snow (bulk density about 300 kg/m^3) of 3 cm thickness. In the figure the experiment 2 was made just after the experiment 1; it is considered that most of smaller snow particles with weaker bonds should have been blown away in the first experiment.

Though the number of data is not large, it is clear that u_{inc} is always larger than u_{dec}, and that the difference between the two velocities becomes larger as the temperature is higher or the snow structure is stronger.

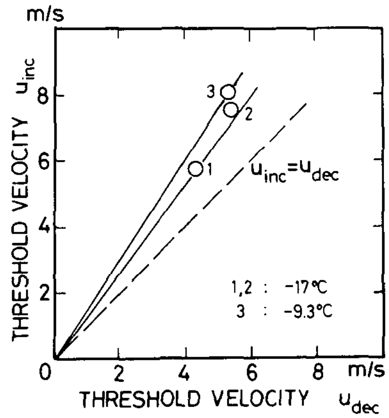

Fig. 2. Threshold wind velocity for the initiation or cease of blowing snow measured in increasing (u_{inc}) and decreasing (u_{dec}) wind velocity in a cold horizontal wind tunnel. The average wind velocity was measured at 25 cm above the snow surface.

The above two kinds of experiments to initiate fluidization in snow by vertical or horizontal air flow show that the onset velocity for fluidization or blowing snow becomes larger as the temperature is raised or the strength of snow is increased. The result is in harmony with the observation by Oura *et al.*[9] at Syowa Station in Antarctica that the threshold wind velocity of blowing snow is almost constant at temperatures below about –7°C, but increases with rising temperature above –7°C.

The above results also suggest that the onset of blowing snow is essentially statistical in nature; as was shown recently[10,11] the bulk density and temperature are not enough to determine the strength of snow. Onset mechanics of blowing snow should utilize more statistical parameters such as introduced in snow mechanics associated with avalanches. More elaborate experimental and theoretical works are required to make clear the formation mechanism of nuclei of fluidization within snow.

3. Wind profile in blowing snow[12]

The vertical profile of fully turbulent flow in a boundary layer with neutral stability is given by

$$\frac{u}{u_*} = \frac{1}{k} \ln\left(\frac{y}{y_0}\right), \tag{1}$$

where u is the mean horizontal wind velocity at height y above the surface, y_0 is the roughness parameter, u_* is the friction velocity ($u_* = \sqrt{\tau/\rho}$, where τ is the shear stress and ρ is the density of air), and k is the von Karman's constant which is usually put to be 0.4. Eq. (1) has been demonstrated by many researchers[1),9),13)] to hold on snow covers in fields and laboratories, but it should be noted that some new treatments[14),15)] have recently been given on the wind structure on a permeable surface, which take account of so-called slip velocity and internal flow within the surface snow.

On the other hand, the wind profile in blowing snow has never been investigated systematically and not understood well, mainly because the velocity measurements in blowing snow and its theoretical treatments are considerably difficult. On the basis of the experimental results, Bagnold[16),17)] concluded that the wind profile in the presence of saltation of sand particles can be expressed by

$$\frac{u}{u_*} = \frac{1}{k} \ln\left(\frac{y}{y_t}\right) + \frac{u_t}{u_*}, \tag{2}$$

where u_t is a constant velocity at height y_t which is the height of a 'focus' of height-velocity lines. Though Chepil[18)] reported that the soil movement modifies the wind velocity in much the same way as does drifting sand, described by Eq. (2), it should be mentioned that Owen[19)] proposed the following equation,

$$\frac{u}{u_*} = 2.5 \ln\left(\frac{2gy}{u_*^2}\right) + D, \tag{3}$$

where g is the acceleration of gravity, D is a constant and $u_*^2/2g$ is the thickness of the saltation layer. This relation was naturally derived from a hypothesis that the saltation layer behaves, so far as the flow outside it is concerned, as an aerodynamic roughness whose height is proportional to the thickness of the layer. Eq. (3) holds except when $2gy/u_*^2$ is appreciably less than unity, corresponding to the interior of the saltation layer. Eq. (3) can also explain the 'focus' obtained by Bagnold.

The wind profile in the saltation layer does not seem to have been measured accurately especially in blowing snow. Fig. 3 shows the vertical profile of the mean horizontal wind velocity measured on a snow cover of 3 cm in thickness and 0.5 m $\times$ 7.0 m in area in a cold wind tunnel at –9.6°C. The wind velocity was measured with a hot-wire anemometer and recorded with an X-Y recorder through a low-pass filter of 0.89 Hz. The probe used was a platimum film coated with a thin quartz. Fig. 3 shows that a turbulent boundary layer of about 10 cm thickness is steadily formed on the snow surface.

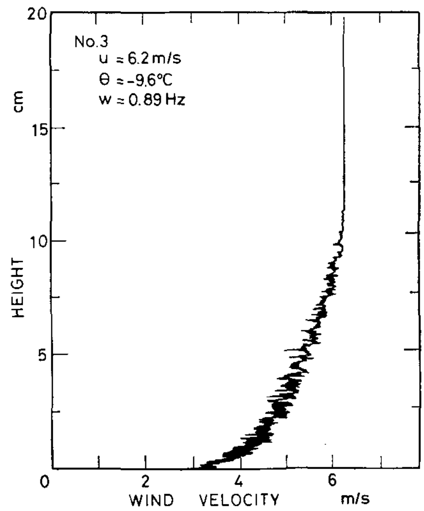

Fig. 3 Horizontal wind velocity (u) measured on a snow cover of 3 cm in thickness and 0.5 m$\times$7.0 m in area in a cold wind tunnel at –9.6°C.

Modifications of the wind profile by the snow particle motion are shown in Fig. 4, in which the blowing snow was generated by supplying seed snow particles far windward to trigger the onset of saltation of snow particles. In Fig. 4-a, the wind velocity in the boundary layer decreased considerably as shown by the dashed curves. But the decrease cannot be simply attributed to the interaction between the wind and saltating particles, because in this case the supply rate of seed snow particles was relatively large so that the seed particles could reach directly the point of the wind measurement. Neverthe-less the cross-over of the two curves near the snow surface should be noted.

Fig. 4-b is the case in which the supply rate of seed snow particles was much smaller. In this case the wind profile can be regarded to have been modified solely by the intraction with the snow particle motion. Now the increase in the velocity just near the snow surface is clear. The plot of ln y against u gives a deviation from a straight-line relationship below a height of about 1 cm. The height corresponds to the thickness of the saltation layer: it was confirmed by the measurement of particle concentrations to be explained in the section 5 that the vertical profile of the particle concentration was exponential and most particles were concentrated within a few centimeters above the snow surface. The possible mechanism of the increase in the wind velocity in the saltation layer will be discussed in the next section.

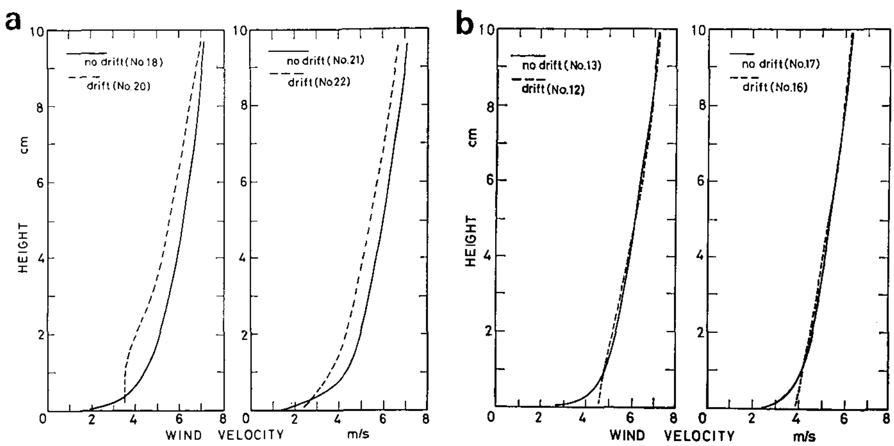

Fig. 4 Horizontal wind velocity measured when seed snow particles were supplied far windward to trigger the initiation of blowing snow. Dashed curves refer to the velocity profiles in the blowing snow. The supply rate of particles in (b) was much smaller than in (a).

4. Motions of snow particles in blowing snow[20]

Photographic investigations of motions of snow particles in blowing snow were first made by Oura *et al.*,[9),21)] who have found that in low drifting snow most snow particles are transported by saltation and that no evidence of creep or suspension of snow particles could be observed. We constructed a simple device[20)] which can take a photograph of particle trajectories with time marks. Some examples are given in Fig. 5. The three modes of transport for blown snow,[5)] i.e., creep, saltation and suspension, can be recognized in the photographs. This technique is very advantageous because we can calculate velocities and accelerations of individual particles in addition to the trajectories.

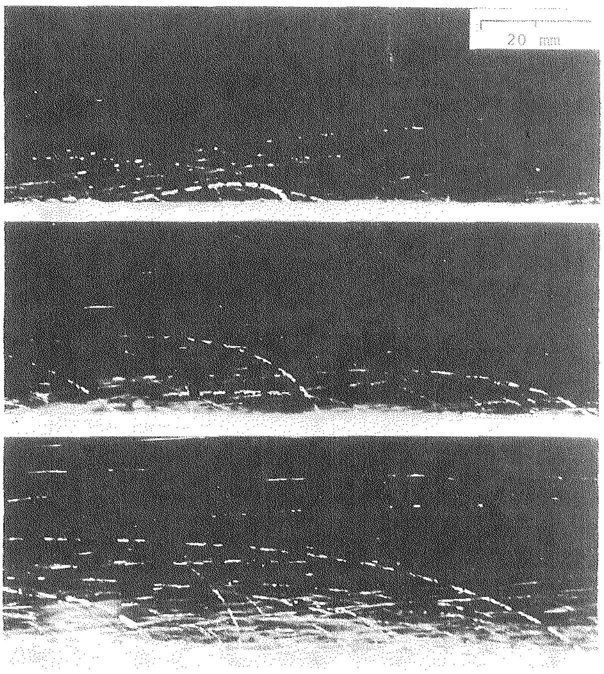

Fig. 5 Photographs showing the motion of snow particles in the saltation layer. The time interval of marks on each trajectory is 3.6 ms. More details of the experiment are given in Araoka and Maeno.[20]

Horizontal (u) and vertical (v) components of velocities of snow particles were calculated from the time marks on trajectories and plotted against the height in Figs. 6 and 7. Open and solid circles refer respectively to ascending and descending particles, and the curve in Fig. 6 is the horizontal mean wind velocity measured with a Pitot tube.

The following inference can be drawn from Fig. 6: the horizontal velocities of ascending particles are accelerated and approach the horizontal wind velocity until the relative velocities become zero. On the other hand, the velocities of descending particles are larger than the horizontal wind velocity and are therefore mostly decelerated.

The result suggests that the effect of snow particle motion on the wind

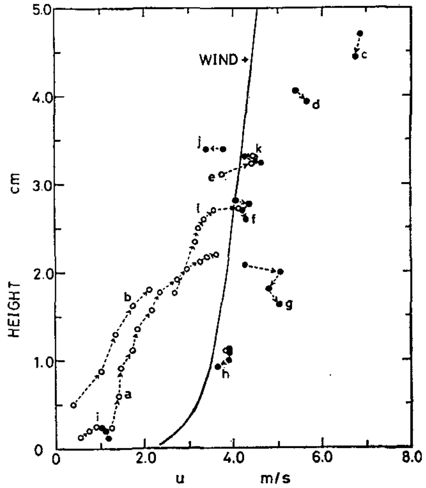

Fig. 6. Horizontal velocity of snow particles plotted against the height. Open and solid circles refer to the ascending and descending particles respectively. The solid line is the profile of mean wind velocity measured with a Pitot tube.

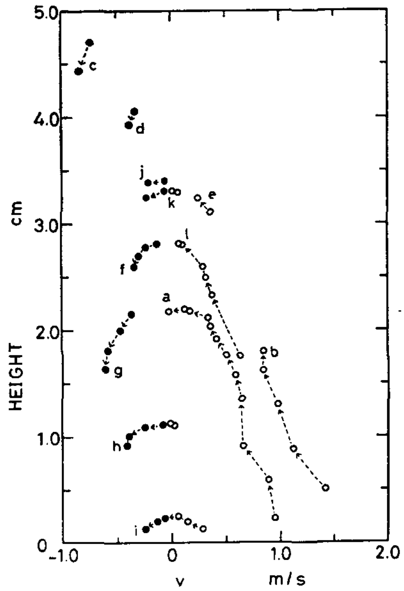

Fig. 7. Vertical velocity of snow particles plotted against the height. Open and solid circles refer to the ascending and descending particles respectively.

velocity profile in the saltation layer is to make it uniform, that is to decrease the shear near the surface, by slowing down the upper parts of wind which move faster and by speeding up the lower parts which move more slowly. The effect can also be interpreted as that the net momentum in the horizontal direction was transported downward by the saltating particles. This result is in good agreement with the increase in the wind velocity in the saltation layer (Fig. 4) discussed in the preceding section. The drag acting on snow particles was calculated from the relative velocities and accelerations at each height and was found to be of Stokes type[20].

Vertical velocities of ascending snow particles are decelerated, but those of descending ones are accelerated (Fig. 7). The result is reasonable if only the sense of gravity is concerned, but not if quantitative aspects are taken into account. Fig. 8 gives the calculated acceleration ($a_y>0$) and deceleration ($a_y<0$) for some particles at each height. If the air in the saltation layer is assumed to show no vertical motion, the acceleration of descending particles should be smaller than g ($=9.8$ m/s^2). However, most observed accelerations are larger (Fig. 8), suggesting that some downward force is acting on the snow particles in the saltation layer. On the other hand, the absolute values of deceleration of ascending particles are much larger than g. This also suggests the exsistence of some force acting downward.

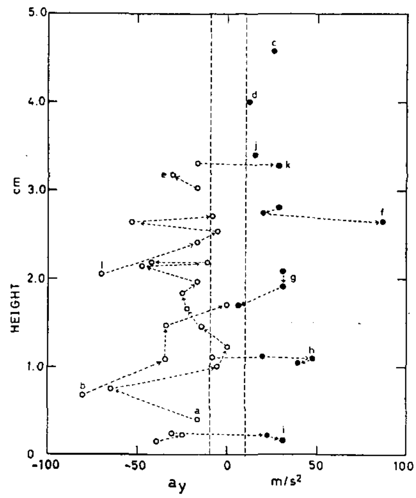

Fig. 8. Acceleration ($a_y>0$) and deceleration ($a_y<0$) of vertical velocities of snow particles plotted against the height. Dashed straight lines refer to the acceleration of gravity.

The above results can be explained if we assume that net momentum in the vertical direction is transported downward by saltating snow particles. This process is probable because the vertical velocities of descending particles are larger than those of ascending ones. However it is not clear whether the downward force is generated by the probable downward bulk motion of air or by the possible pressure gradient.

The result is consistent with that obtained by Bagnold,[16] who found an empirical equation (Eq. (4) in his paper) showing that the maximum height of rise of a sand particle is always smaller than $v^2/2g$, where v is the vertical velocity of the particle. On the other hand, White and Schulz[22] have shown recently that the velocities of ascending glass spheres are accelerated even more largely than g because of a Magnus effect due to spinning of the particles in the wind shear.

Our experimental results suggest that the mechanism to eject and accelerate snow particles is involved within a layer of a few millimeters above or partly including the snow surface, where the wind shear is the largest: incident particles, which have been accelerated extensively, collide with the snow surface with small impact angles, and cause new particles to lift off with large angles nearly equal to 90 degrees. Details of the process will be published elsewhere.[20]

5. Momentum and energy transport in fluidized snow[8),23),24)]

Maeno *et al.*[8),23),24)] have verified experimentally the existence of pseudo-viscosity in fluidized snow, which may enable men or vehicles to 'swim' in an avalanche. The viscosity is considered to be caused by the enhanced turbulence and collisions of snow particles as well as the molecular viscosity of air. In the case of blowing snow, the viscosity of the fluidized snow appears to give rise to the modification of the horizontal wind velocity as shown in Fig. 4 and the downward force acting on snow particles as shown in Figs. 7 and 8.

The effective transport of energy (heat) in the fluidized snow was also demonstrated by Maeno *et al.*,[8),23),24)] who have shown that the heat transfer efficiency is increased by a factor of three or four by the fluidization of snow, and that the considerable portion of the increase is attributable to the collision or approach of snow particles.

Fig. 9 gives the heat transfer coefficient (h) of a brass sphere (6.0 mm in diameter), measured at various heights above a snow surface in a cold wind tunnel at −9.3°C. The solid curve (u) is the profile of mean wind velocity measured with a Pitot tube. It is seen that the value of h (open circles) is larger at higher levels, corresponding to the vertical wind velocity. On the other hand, when a faint blowing snow was generated by supplying seed snow particles far windward, the vertical profile of h was varied considerably as shown by solid circles.

In blowing snow the heat transfer is much effective at lower levels. This result implies that saltating snow particles play an important role in the heat transfer in the fluidized snow layer near the surface. In Fig. 9 is shown the drift density (ρ), that is the mass of snow particles contained in a unit volume of air. The drift density was estimated from a picture as shown in Fig. 10, which was taken with a single shot of stroboscopic light lasting 25 μs. It is clearly shown that the heat transfer coefficient increases with increasing drift density.

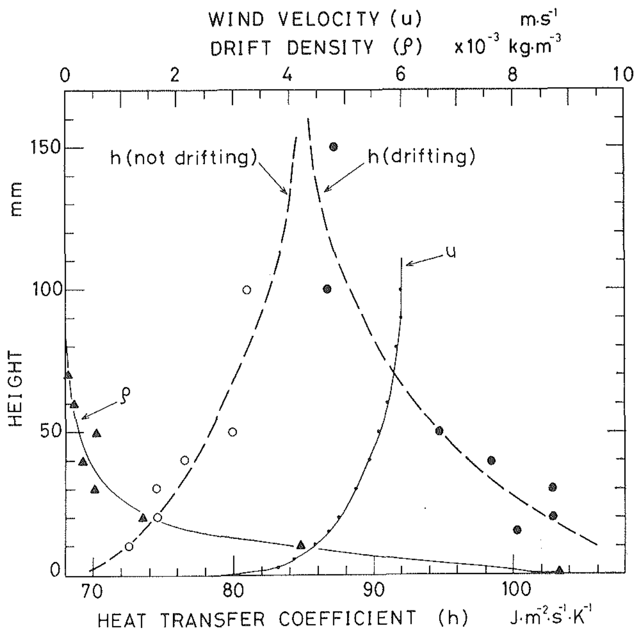

Fig. 9. Heat transfer coefficient (h), drift density (ρ) and mean wind velocity (u) plotted against the height. The measurement was made in a cold wind tunnel at –9.3°C.

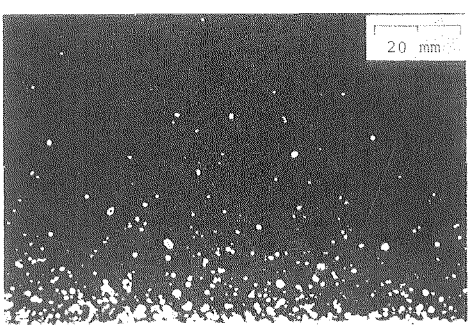

Fig. 10. Photograph of snow particles in blowing snow taken under a single shot of screened strobo light lasting 25 μs.

Fig. 11 gives the heat transfer coefficient measured in natural blowing snow at Haboro in Hokkaido, north part of Japan. Values of heat transfer coefficient (h) measured at various heights are plotted against the mean wind velocity (u) at each height. The parameter is the relative strength of drifting snow, which is specified in the figure caption. The increase in h with increasing wind velocity and drift density is consistent with the result obtained in the fludization experiment[8),23),24)] and in the wind tunnel.[23)] More detailed analyses will be published elsewhere.[25)]

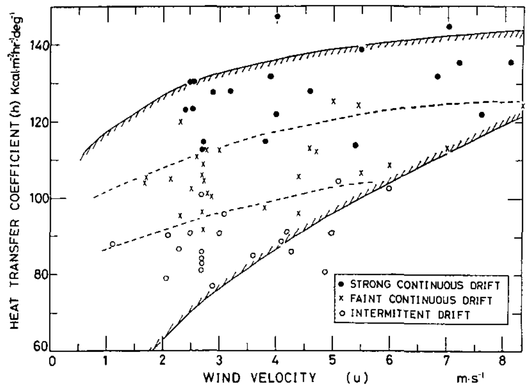

Fig. 11. Heat transfer coefficient (h) against the mean wind velocity (u) in natural blowing snow at Haboro (mean air temperature, $-7.0°C$). The blowing snow was classified into three classes: strong continuous drift (●), faint continuous drift (×), and intermittent drift (o). Dashed lines show rough boundaries.

6. Concluding remarks

Various physical aspects of fluidized snow were discussed in special reference to blowing snow. The discussion has led to a conclusion that the interaction between wind and snow is much complicated but is considerably important in understanding properly the wind structure above the snow cover and the energy and momentum transfer at the air-snow interface. Vertical transfer of momentum through the motion of snow particles is significant in determining the wind structure in the boundary layer on the snow surface. Effective heat transfer is also an important property of fluidized snow or blowing snow, which should be taken into account in estimating the energy balance.

The equation of the energy budget at the air-snow interface where melting does not occur, has often been considered in the form

$$Q_R = Q_T + Q_L + Q_C \,. \tag{4}$$

This equation means that the net radiation flux (Q_R) of short and long wave radiation is balanced with the sum of fluxes of sensible heat due to the wind turbulence (Q_T), latent heat due to the evaporation or condensation of water vapor (Q_L), and conduction heat through the snow cover (Q_C). However, our work suggests that the heat transfer due to the movements of snow particles should be included explicitly in the energy balance equation because blowing snow is known to occur quite frequently in snow fields, especially on the ice sheets in polar regions. It should be emphasized that the vertical heat transfer by snow particle motions is much effective in the boundary layer where strong temperature inversions are generated, and that the presence of fluidized (blowing) snow may modify the radiation properties of the air near the surface.

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