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Effect of wind on surface hoar growth on snow

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Abstract. Field observations of surface hoar formation were carried out with the measurements of water vapor condensation rate, snow surface temperature, air temperature, humidity, wind speed, and net radiation. Large surface hoar crystals were observed to form under the breezy wind, 1 to 2 m s^{-1}, at 0.1 m high. The condensation rate increased linearly with the product of the vapor pressure gradient and the wind speed. The bulk transfer coefficient of water vapor C_e was roughly constant when the surface hoar crystals were small, whereas it showed some increase as the hoar crystals grew to several millimeters in height. A possible cause of this increase in C_e is that the developed surface hoar crystals modify the aerodynamic roughness and consequently increase the turbulent transfer of water vapor.

1. Introduction

Surface hoar crystals are formed by the deposition of water vapor onto the snow surface during the clear night. Surface hoar crystals have long been of interest to avalanche researchers, because these crystals may form a typical sliding layer for a slab avalanche release. When surface hoar crystals are buried by a subsequent snowfall, the surface hoar layer often becomes a weak layer because of their lack of intercrystalline bonding and weak attachment to the original snow surface (Figure 1). *Lang et al.* [1984] reported that although a surface hoar layer would become visually undetectable, shear strength remained too low to measure for extended periods of time. *Föhn* [1992] observed that weak layers consisted of aged surface hoar more frequently than the faceted snow particle and the depth hoar. Thus to establish the exact avalanche forecasting system, it is essential to inquire into the details of suitable meteorological condition for the surface hoar growth.

However, no previous quantitative study of the relation between the surface hoar growth and meteorological conditions has been conducted. *Lang et al.* [1984] obtained only profiles of temperature near the snow surface. *Colbeck* [1988] concluded theoretically that surface hoar could not grow at the observed rates if molecular diffusion was the only mechanism of vapor transfer to the snow surface; then some wind was necessary for turbulent transfer to an interfacial sublayer, though perceptible winds had been reported to preclude surface hoar growth. It is already known that the dewfall growth mechanism which seems very similar to the surface hoar can occur only if there are both the turbulence and the stable condition which create the vapor inversion [*Oke,* 1987]. *Oke* [1987] noted that the critical wind speed depends on the roughness of the surface; hence the snow surface roughness may have considerable effect on the critical wind speed in surface hoar formation. In this study, field observations were carried out to clarify the formation rate of surface hoar crystals quantitatively, particularly considering the wind effect.

2. Observation Site and Methods

Observations were carried out on selected humid nights for nearly 2 months from January to March 1994. The site is located at a mountain ridge near the avalanche research station (240 m above sea level) of the Institute of Low Temperature Science, at the Teshio Experimental Forest of Hokkaido University in Toikanbetsu, northern Hokkaido, Japan. The station (45°N, 142°E) is close to the Japan Sea (about 20 km in distance), and the southwest is the predominant wind direction in winter.

Observation methods and instruments used are as follows:

1. Vapor condensation rates were obtained by the weight change of 20-mm-thick snow on an aluminum plate of 0.5 × 0.6 m^2. The measurement device and method are shown in Figure 2. The snow plate was set on a shallow pit which is adjusted to the depth of the plate beforehand. The measurements were carried out every 30 min with an electric balance during nighttime, and the hoarfrost under the plate was scraped carefully by a plastic card.
2. Temperatures were measured with copper-constantan thermocouples at the snow surface, at heights of 5 mm, 10 mm, 20 mm, 0.1 m, and 1 m above the surface, and at depths of 10 mm, 20 mm, 50 mm, 0.1 m, and 0.15 m below the surface. Since the temperatures above the surface might be underestimated owing to radiational cooling on the thermocouple itself at night, we corrected the data with the following equation [*Kondo,* 1982]:

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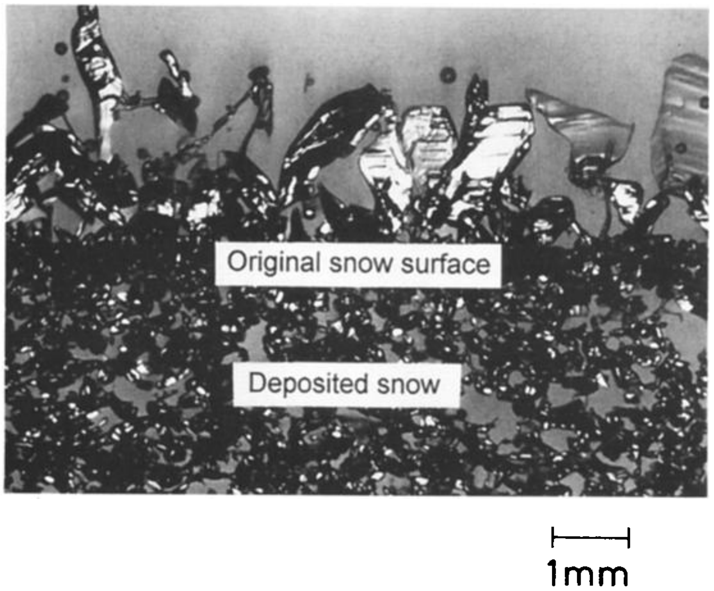

Figure 1. Vertical thin sections of layers of surface hoar crystals. (Left) Surface hoar crystals, about 3 mm high, grew on the original snow surface before snowfall. (Right) After subsequent snowfall, the surface hoar layer became relatively low density and weak. (Photo by T. Fukuzawa.)

$$\Delta T = \frac{0.41 Q_n L}{C_1 + C_2 L^{0.5} (u^2 + C_3 L \Delta T)^{0.25}}, \qquad (1)$$

where ΔT is the correcting value of the temperature (in degrees centigrade), Q_n is the net radiation (in watts per square meter), L is the diameter of the thermocouple (in meters), and u is the wind speed (in meters per second). C_1, C_2, and C_3 are correction factors and are 0.02, 1.33, and 0.0278, respectively.

3. Relative humidity was measured at heights of 0.1 m and 1 m with hygrometers of the static electric capacitance type. The sensor (VAISALA) at 1 m high was encased in a radiation shield, and the sensor (5 mm in size and made by KANOMAX Japan, Inc.) at a height of 0.1 m was surrounded by a metal pipe 10 mm in diameter; hence the radiation effect could be neglected. These sensors showed the relative humidity with respect to liquid water. Water vapor pressure was calculated from

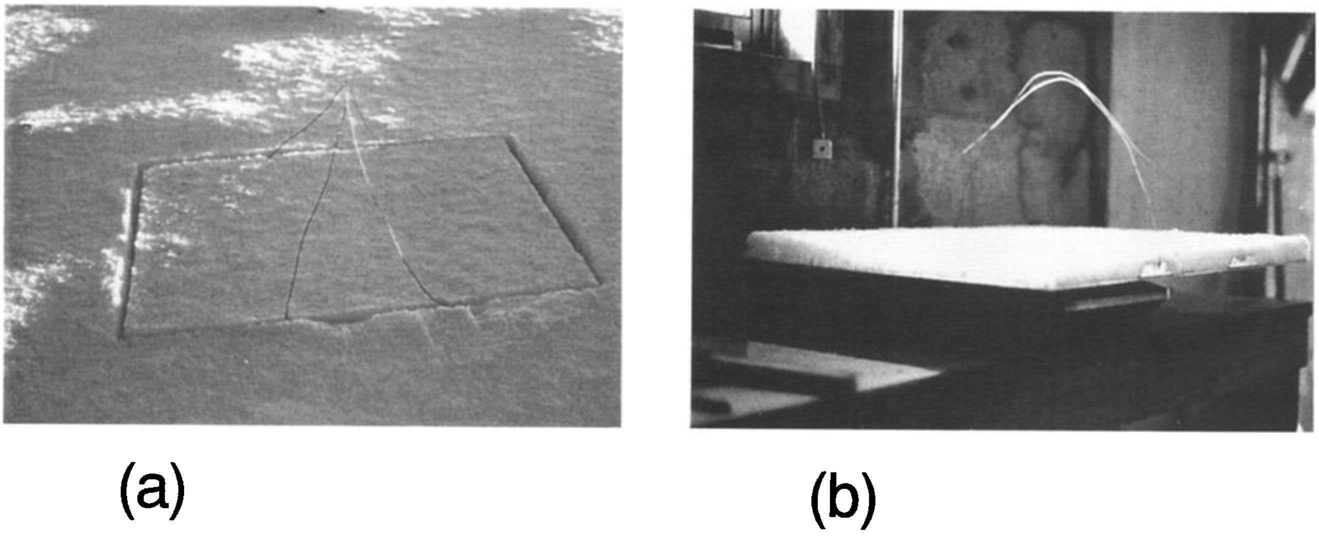

Figure 2. Measuring device for vapor condensation. (a) Aluminum plate with 20-mm snow placed in the shallow pit. (b) Measurement of the weight change with an electric balance.

the humidity and the temperature data. Vapor pressure at the snow surface was obtained assuming that it was saturated at the surface. In this paper, ΔP shows the vapor pressure difference between the height of 0.1 m and the snow surface.

4. Wind speed was measured at heights of 0.1 m and 1 m with three-cup anemometers (Makino Applied Instruments, Inc.). The height of the sensor was about 0.2 m; therefore the bottom of the sensor at 0.1 m high was buried in snow.

5. Net radiation was measured at a height of 1 m with an all-wave net radiometer (EKO Instruments Co., Ltd.).

All the meteorological data were measured every minute and averaged for 30 min.

In the heat balance calculation at the snow surface, the latent heat flux was derived from the vapor condensation rate, and the conductive heat flux below the snow surface was estimated by the temperature gradient, supposing that the thermal conductivity of snow with a density of 200 kg m^{-3} was 0.2 W m^{-1}K^{-1} [*Izumi and Huzioka,* 1975]. The sensible heat flux H (in watts per square meter) was calculated with a bulk method [*Takeuchi and Kondo,* 1981]:

$$H = -C_h c_p \rho u_z (T_z - T_s), \tag{2}$$

where C_h is the bulk transfer coefficient of heat, c_p is the specific heat of air (in J kg^{-1}K^{-1}), ρ is the density of air (in kilograms per cubic meter), u_z is the wind speed (in meters per second) at height z, and T_z and T_s are the temperatures at a height z and the surface, respectively. C_h was calculated from the observation data, according to *Stull* [1988], as follows:

$$C_h = \frac{\kappa^2}{\phi_m \phi_h \ln(z/z_0) \ln(z/z_h)}, \tag{3}$$

where κ is Kármán's constant, ϕ_m and ϕ_h are dimensionless stability functions of momentum and heat, respectively, and z_0 and z_h are aerodynamic roughnesses of momentum and heat, respectively. C_h was given by substituting ϕ_m, ϕ_h, z_0, and z_h into equation (3). Stability functions ϕ_m and ϕ_h are expressed with an index of air stability [*Thom,* 1975] as given by

$$\phi_m = \phi_h = \phi_e = (1 - 5R_{iB})^{-1} \quad 0 \le R_{iB} \le 0.2, \tag{4}$$

where ϕ_e is the dimensionless stability function of water vapor and R_{iB} is the bulk Richardson number, defined by [*Oke,* 1987]

$$R_{iB} = \frac{g(T_z - T_s)z}{T u_z^2}, \tag{5}$$

where g is the gravitational acceleration (in meters per second per second), and T is the mean absolute temperature of the air (in kelvins). Under neutral conditions it is reasonable to assume the following logarithmic profile:

$$u_z = \frac{u_*}{\kappa} \ln \frac{z}{z_0} \tag{6}$$

$$T_z - T_s = \frac{T_*}{\kappa} \ln \frac{z}{z_h}, \tag{7}$$

where u_* and T_* are the friction velocity (in meters per second) and friction temperature (in kelvins), respectively; z_0 and z_h were obtained by substituting the profiles of the temperature and the wind speed in equations (6) and (7).

3. Results

Surface hoar crystals were formed during the clear, humid, and light breezy nights. During the observation period, conspicuous formations of surface hoar were observed only at four nights: January 8 to 9, February 27 to 28, February 28 to March 1, and March 1 to 2, 1994. At other clear nights, large surface hoar did not form because of dry or windy conditions. In general, the surface hoar grew up when the snow surface temperature was 5 °C (or more) lower than the air temperature, the humidity was higher than 90%, and the wind speed was 1 to 2 m s^{-1} at 0.1 m high.

On March 1 to 2, 1994, it was clear all night. The surface hoar crystals were observed to grow up to 5 mm high. Time variations of the temperatures above and below the snow surface, the relative humidity, the vapor pressure, the wind speed, the heat balance components at the snow surface, and the condensation rate of water vapor are shown in Figures 3a-3g, respectively.

The temperatures both above and below the surface had gradually decreased until 0330 LT and then generally increased. The temperatures increased with height (Figure 3a) and depth (Figure 3b) and formed large temperature gradients both above and below the surface. The profile of air temperatures was nearly logarithmic with height except during 0100 to 0330 LT. The temperature difference between 1 m high and the snow surface was about 5 °C before midnight, and it reached 8 °C later. This is mainly because the sensible heat flux which may warm the snow surface decreased with a decrease in wind speed after midnight. After 0400 LT the temperature difference between the air and the snow surface decreased as the wind speed increased, and the snow temperature just below the surface became lower than the surface temperature owing to the small thermal conductivity of snow. Since the humidity was high as shown in Figure 3c, the large temperature difference in the air caused a large ΔP, and water vapor in the air was transferred to the snow surface. The wind speeds were 0.8 – 1.6 m s^{-1} at 0.1 m high and 1.5 – 2.6 m s^{-1} at 1 m high except during 0100 to 0330 LT. The net radiation nearly balanced with the sum of the sensible heat flux, the latent heat flux, and the conductive heat flux below the surface, all of which were obtained independently. The net radiation had a maximum (85 W m^{-2}) at 1900 LT and gradually decreased. The sensible heat flux was largest in the incoming flux, whereas the contribution of the latent heat flux was small but not negligible. Condensation occurred all night without a systematic change, and an average condensation rate was about 3.7×10^{-6} kg m^{-2} s^{-1}.

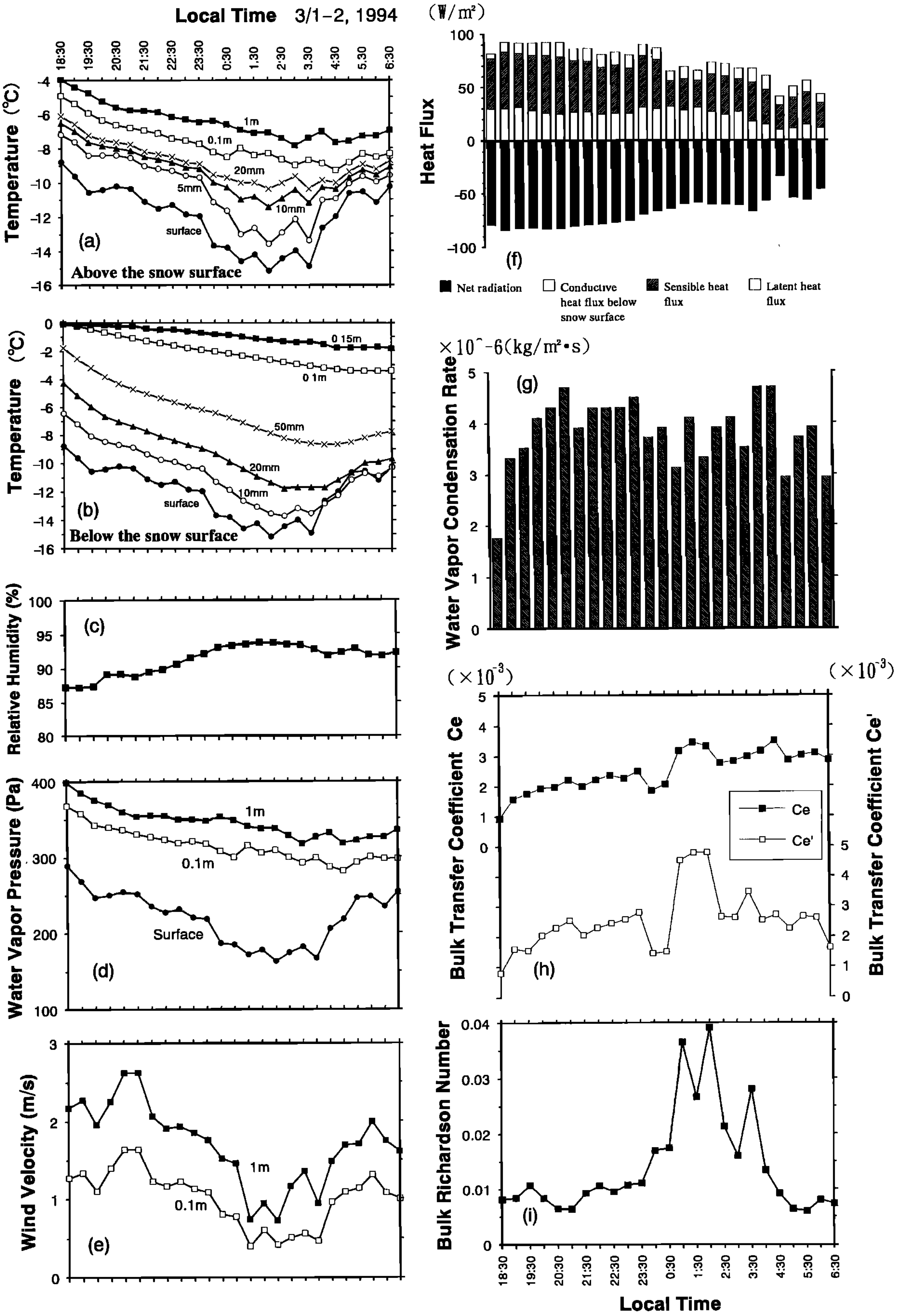

Figure 3. Time variations of meteorological elements from March 1 to 2, 1994. (a) Air temperatures at five different heights above the snow surface and the surface temperature. (b) Snow temperatures at five different depths below the snow surface and the surface temperature. (c) Relative humidity at 1 m high. (d) Vapor pressures at 0 m, 0.1 m, and 1 m above the surface. (e) Wind velocities at 0.1 m and 1 m above the surface. (f) Each heat balance component at the snow surface. (g) Water vapor condensation rate. (h) Bulk transfer coefficient of water vapor. (i) Bulk Richardson number for the layer from 0 to 1 m.

In the following two graphs, all the data of the clear nights from January to March 1994 are plotted, and the values are averaged for 30 min. Figure 4 shows the relation between the condensation rate and wind speed at 0.1 m high. When the wind speed was from 1 to 2 m s^{-1}, condensation rates were relatively high. On the contrary, evaporation occurred in the case of $\Delta P < 0$. Figure 5 shows the condensation rate plotted against the product of ΔP and wind speed (0.1 m high). Since the measurements focused on the hoar formation, the number of data in the evaporation part was not enough. However, it is reasonable to conclude that a good linear relationship ($R = 0.93$) exists in both condensation and evaporation range. At any rate, it must be noted that the condensation rate of surface hoar can be roughly estimated from meteorological elements, which are air temperature, surface temperature, humidity, and wind speed.

4. Discussion

4.1. Wind Effect on Surface Hoar Formation

As we mentioned before, the surface hoar was observed to form under breezy wind. Here we discuss the effect of wind on the surface hoar formation. The transfer coefficient of water vapor K_e is defined as [*Oke,* 1987]

$$\frac{E}{\rho} = -K_e \frac{dq}{dz}, \tag{8}$$

where E is the water vapor flux (in kg m^{-2} s^{-1}), dq/dz is the specific humidity gradient, and height z is taken positive upward. Substituting the values on March 1 to 2, 1994, the range of K_e was calculated to be $2.3 \sim 8.5 \times 10^{-3}$ m^2 s^{-1} between 1 m high and the snow surface, and the mean value was 5.0×10^{-3} m^2 s^{-1}. Since the molecular diffusivity of water vapor in air is 2.22×10^{-5} m^2 s^{-1} at 0 °C [*Fischer,* 1988], it is obvious that the large surface hoar was produced by turbulent transfer of water vapor. Thus the wind is required for surface hoar formation; it brings the vapor from water vapor sources such as sea or river as well.

However, when the wind speed is too high, the snow surface obtains a large amount of sensible heat from the air, so that the vapor pressure gradient decreases and the condensation rate becomes small [*Colbeck,* 1988]. Consequently, it is reasonable to conclude that there would be suitable wind speed for surface hoar crystal growth. In fact, our observation in Figure 4 showed that the large surface hoar formed under moderate wind, 1 to 2 m s^{-1}. As was pointed out in the introduction, this result was similar to the theoretical aspects of dewfall [*Oke,* 1987].

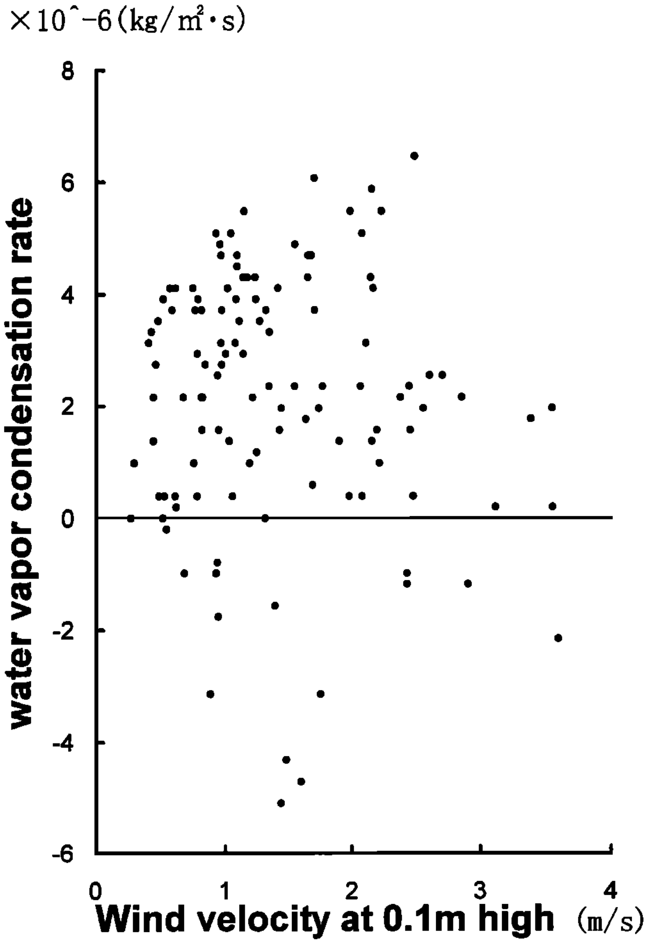

Figure 4. Relation between the condensation rate and wind speed at 0.1 m high, obtained from January to March 1994.

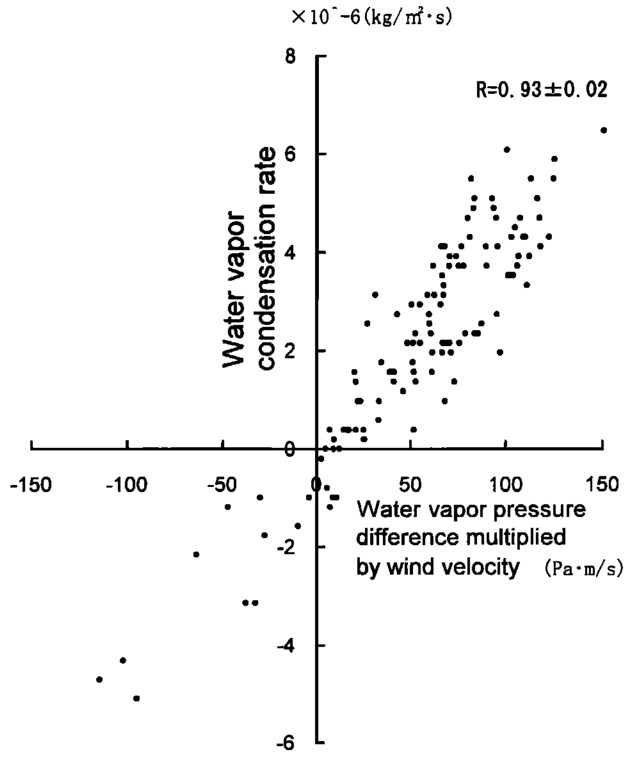

Figure 5. Water vapor condensation rate plotted against the product of ΔP and wind speed (0.1 m high), obtained from January to March 1994.

4.2. Increase in Bulk Transfer Coefficient

Figure 5 indicates that the condensation rate of surface hoar can be estimated with the bulk method [*Takeuchi and Kondo*, 1981]:

$$E = -C_e \rho u_z (q_z - q_s), \tag{9}$$

where C_e is the bulk transfer coefficient of water vapor, and q_z and q_s are the specific humidities at z and the snow surface, respectively. Since $q \simeq 0.622 p_w / p$ [*Sutton*, 1953], where p_w is water vapor pressure and p is atmospheric pressure (in pascals), equation (9) becomes

$$E = -\frac{0.622}{p} C_e \rho u_z \Delta P, \tag{10}$$

where ΔP is the difference of water vapor pressure (in pascals) between height z and the snow surface. Substituting the data of Figure 5 in equation (10), we obtained the mean value of C_e as 2.9×10^{-3} at 1 m high. It agrees approximately with the mean of C_h as 3.3×10^{-3} at 1 m high obtained with equations (3)–(6), although the standard deviation of C_h was not small.

However, in the progress of large surface hoar formation, C_e calculated from equation (10) increased generally with time as shown in Figure 3h. Since C_e is a function of height, air stability, and surface roughness, we first discuss the effect of air stability. Figure 3i shows the time variation of R_{iB} for the layer from 0 to 1 m. R_{iB} increased from 0100 to 0330 LT because of the decrease in wind speed. *Ishikawa and Kodama* [1994] observed in snowmelt seasons that C_h were scattered widely and difficult to specify at wind speed of less than 1.5 m s^{-1} at 1 m high. Thus we can see that the large fluctuation that appeared at midnight in Figure 3h is due to low wind speed and less turbulence. We should notice that except for midnight, R_{iB} was kept constant around 0.01 during the increase in C_e. On the other hand, the bulk transfer coefficient of water vapor can be obtained using the data of two levels in the atmosphere:

$$E = -\frac{0.622}{p} C'_e \rho (u_{1_m} - u_{0.1_m})(P_{1_m} - P_{0.1_m}). \tag{11}$$

Here C'_e depends not on roughnesses but on stability [*Takeuchi and Kondo*, 1981]. In fact, time variation of C'_e in Figure 3h did not show the evident increase except for midnight on March 1–2. Accordingly, we can conclude that the change of air stability was not the case in Figure 3h.

Second, the change of aerodynamic roughness might cause the increase in C_e. Microscopic photographs of surface hoar crystals in Figure 6 indicate their growth with time. If surface hoar crystals grow to several millimeters in height as shown in Figure 6, they would disturb the air flow near the snow surface more or less and increase the turbulence. C_e can be expressed in the same way as (3):

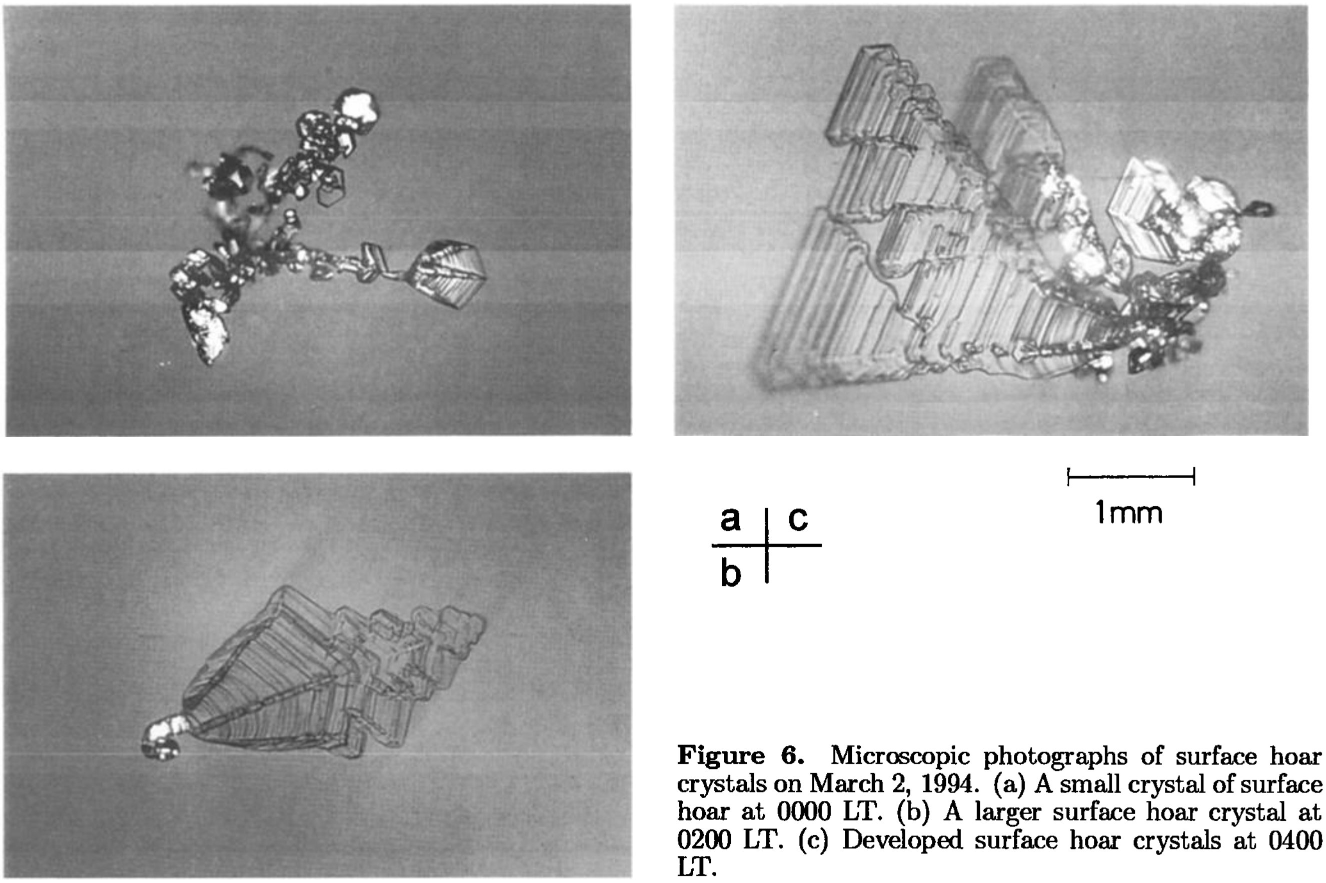

Figure 6. Microscopic photographs of surface hoar crystals on March 2, 1994. (a) A small crystal of surface hoar at 0000 LT. (b) A larger surface hoar crystal at 0200 LT. (c) Developed surface hoar crystals at 0400 LT.

$$C_e = \frac{\kappa^2}{\phi_m \phi_e \ln(z/z_0)\ln(z/z_q)}, \qquad (12)$$

where z_q is aerodynamic roughnesses of water vapor. Although z_0 and z_q in equation (12) do not correspond to the actual height of protrusion, both of them increase as the surface hoar growth. So the surface hoar formation itself leads to the increase in C_e, that is to say, a feedback system exists on the mechanism of surface hoar condensation.

5. Conclusion

The field observations showed a linear relationship between the vapor condensation rate and ΔP multiplied by wind speed. This relation made it possible to estimate the condensation rate of surface hoar from the meteorological data. The optimum wind conditions for the formation of surface hoar was found to be 1 to $2\,\mathrm{m\,s^{-1}}$ at 0.1 m high. When surface hoar crystals grew to several millimeters in size, C_e increased with time. The cause of the increase of C_e was considered to be the increase of surface hoar height, which may induce an increase of aerodynamic roughness. Surface hoar formation involves a feedback mechanism.

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