Estimates of Threshold Wind Speeds for Snow Transport Using Meteorological Data

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ABSTRACT

The threshold wind speed for snow transport is related to properties of the surface snowpack: snow particle bonding, cohesion, and kinetic friction. These properties are controlled by meteorological factors. A method is proposed that relates the threshold wind speed for the initiation of snow transport to standard surface meteorological observations. A complete dataset on the hourly threshold condition for snow transport as determined from visual observation was developed for 16 stations on the prairies of western Canada over six winters. The threshold wind speeds for wet snow transport are significantly different from those for dry snow transport. The majority of recorded threshold 10-m wind speeds ranged from 7 to 14 m s^{-1} with an average of 9.9 m s^{-1} for wet snow transport, and from 4 to 11 m s^{-1} with an average of 7.7 m s^{-1} for dry snow transport. The observations display a nonlinear but generally positive correlation between threshold wind speed and air temperature. An empirical model between threshold wind speed and air temperature was developed for dry snow conditions. The model, on average, provides a good estimate of the threshold wind speed.

1. Introduction

The threshold wind speed for snow transport is the minimum wind speed initiating or sustaining saltation of snow. It is a critical parameter of numerical models that simulate the wind transport of snow as it

1) governs the length of time over which wind transport will occur,
2) indicates the availability of surface particles for transport, and
3) indexes the atmospheric force required to dislodge surface particles.

The threshold is included in many equations for saltation of snow in air, and is extensively used in physically based equations that describe the snow transport phenomenon (Dyunin 1959; Kind 1981; Schmidt 1986; Pomeroy and Gray 1990). There is increasing interest in incorporating algorithms of snow transport and sublimation in hydrological and meteorological models so that snow cover development and ablation can be faithfully represented. However, though there are proposed relationships between the threshold wind speed and snow cover physical conditions (Kotlyakov 1961; Schmidt 1980, 1981), there are no accepted or known methods for determining the variation of the threshold condition with the meteorological conditions that control snow physical properties. It is the purpose of this paper to briefly review the physical conditions that lead to the establishment of a threshold wind speed for snow transport and then to develop from observations a simple method to calculate this wind speed using standard meteorological data. Such a development will provide approximations of the threshold wind speed to be used in snow transport and sublimation flux calculations in hydrological and climatological models, such as the Prairie Blowing Snow Model (Pomeroy et al. 1993).

2. Transport threshold and snow properties

Transport thresholds are often defined as either a dynamic threshold (particles are moving and impact the surface) or a static threshold (initiation of transport, only fluid stress on the surface). The dynamic threshold is that wind speed at which particles stop moving, while the static threshold is the wind speed at which particles start moving (Bagnold 1941). For many theoretical and practical studies involving movement of sand, the dynamic threshold is preferred because the processes of particle impact and dislodgement are active at the cessation of saltation. The force applied by these particles ejects further particles (Anderson and Haff 1991) and as a result, the dynamic threshold wind speed is roughly

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80% of that at the static threshold for aeolian movement of noncohesive materials (Greeley and Iversen 1985). For snow transport the distinction is less useful. The snow surface is strongly conditioned by particle impact and undergoes densification and crystal rebonding (Schmidt 1981; Gray et al. 1970). As a result, the dynamic threshold wind speed of a short snow transport event may be lower than that for a long snow transport event even if initial snowpack conditions are the same. Schmidt (1980) showed that particle impact forces are necessary to eject surface snow particles and that fluid forces alone are insufficient to initiate snow transport under normal terrestrial wind regimes. Hence, even the initiation of snow transport must be a ''dynamic'' threshold, likely initiated by ''flurries'' of saltating grains in response to a turbulent sweep as postulated by Nickling (1988).

Previous studies have shown that threshold wind speed can be estimated using knowledge of the particle size distribution in the lower suspension layer of drifting snow. Schmidt (1981, 1986) provided an expression based on field measurements to relate threshold wind speed to the wind speed and the mean particle diameter as follows:

$$U_t = U - \left(\frac{d}{100}\right)^4 - 4, \tag{1}$$

where U_t (m s^{-1}) is threshold wind speed, U (m s^{-1}) is the wind speed, and d (μm) is mean diameter, all as measured 25 cm above the surface. Note that for suspended snow d is positively correlated with U (Budd 1966). Schmidt (1981, 1986) demonstrated that this expression could be inaccurate for a general case because it does not consider a more important factor, cohesive bonding force, which depends on time and temperature histories of the snow surface. In addition, the requirement of measuring d precludes use of this expression with standard meteorological data.

Mellor (1965) reported that wind speeds at a height of 10 m in the range 3–8 m s^{-1} are sufficient to dislodge loose, unbonded snow, and wind speeds greater than 30 m s^{-1} are required to dislodge snow particles bonded by the freeze–thaw process. Schmidt (1980) reported that the threshold wind speed for snow transport will increase with time since snow deposition. Pomeroy et. al. (1993) identified typical values for the threshold friction velocity (related to wind speed) ranging from 0.15 to 0.25 m s^{-1} for fresh, loose, dry snow during snowfall and from 0.25 to 1.0 m s^{-1} for older, wind-hardened, dense or wet snow. Observations in the Antarctic (Budd et al. 1966) suggest high threshold wind speeds 14 m s^{-1} for snow transport at extremely cold temperatures. All studies indicate that the snowpack resistance determining the threshold condition for transport is related to snow particle bonding, cohesion, and kinetic properties.

Wind-deposited snow is often densely packed, providing a large number of contact points and hence bonds. Snow particle bonding tends to increase gradually, or sometimes drastically, from day to day as a result of mass and energy fluxes and changes in the state of the snow crystals. One important process that can lead to this increase is metamorphism, which changes the structure of snow over time. Metamorphism can occur under a temperature gradient or under uniform temperature. Water vapor, controlled by the vapor pressure, will diffuse from the warmer part of snowpack to the colder part of snowpack, where the saturation vapor pressure is lower, condensation occurs, and crystals grow. This process occurs due to vertical gradients in snowpack temperature and is termed temperature gradient metamorphism (Sommerfeld and LaChapelle 1970; Colbeck 1987). Snow metamorphism also occurs when the temperature is uniform with snowpack depth or gradients are very small (Sommerfeld and LaChapelle 1970). The vapor pressure varies with the shape of ice crystals; vapor pressure decreases from a convex surface, through a plane surface, to a concave surface. This indicates that at a microscopic scale where temperature is relatively uniform, vapor will move from convex surfaces to concave surface. As this process occurs under conditions of uniform temperature, it is known as equitemperature metamorphism. Equitemperature metamorphism results in the growth of ice bonds by sintering and strong cohesion between crystals.

Yong and Metaxas (1985) reported that at a relatively constant temperature of -13°C, the density of the natural fresh snow (initially about 100 kg m^{-3}) aged for 30 and 50 days increased to 300 and 400 kg m^{-3}, respectively. They referred to this change as age hardening. Gray and Morland (1995) agreed that rapid snow compaction in initial periods after deposition is due to the thermal processes of metamorphism. Perla (1985) found that under either strong or weak temperature gradients, the surface area per unit mass of snow crystals decreased and density increased over time. This results in an increase in the number of intercrystal bonds per unit mass but not necessarily an increase in the strength of individual bonds. The increase in the number of bonds is an important factor governing overall snow strength, however. Metamorphism, whether temperature gradient or equitemperature, generally results in increase of surface snow density, mechanical continuity, and hence strength over time.

The development of intercrystal bonds in a metamorphosing snowpack is closely related to mass and heat transfer in snowpack. Recent physical modeling studies of snowpacks demonstrate that the requirements of theoretical models to predict the mass and heat transfer in field conditions cannot be satisfied (Gray and Morland 1994; Gray et al. 1995). In practice, previous studies have related the growth of intercrystal bonds to temperature (Hosler et al. 1957; Ôura et al. 1967), to time and temperature (Hobbs and Mason 1964), or to wind action (Gray et al. 1970). Hobbs and Mason (1964) reported that the bond growth in snow is directly related to both temperature and time and that bonds grow over time even under constant temperature.

Cohesion is related to viscous forces associated with thin layers of liquid water on snow crystals under warm conditions. Cohesive forces increase dramatically as snow becomes wet. Nakaya and Matsumoto (1954) demonstrated the existence of a liquid water film on the ice surfaces even at the ambient air temperatures below 0°C. Experiments on snow indicate that the cohesion between snow particles increases exponentially with increasing temperature (Schmidt 1980), most likely because of the increase in thickness of the quasi-liquid layer surrounding snow crystals. Hosler et al. (1957) found that cohesion of snow particles at 0°C is eight times that at −15°C. Ôura et al. (1967) show the threshold wind speed increases with increasing temperature above −7°C, presumably as the quasi-liquid layer on the surface of snow particles became substantially thicker. Conklin and Bales (1993) show a quasi-liquid-layer thickness of 3–30 nm at −60°C, increasing to 500–3000 nm at −1°C; the increase in layer thickness with temperature is nonlinear and becomes rapid above −8°C. Evidently, three snow cohesive regimes important to the initiation of snow transport exist:

1) a wet cohesive regime governed by the amount of snowmelt water and important at temperatures near to 0°C,
2) a warm cohesive regime governed by a thicker quasi-liquid layer, and
3) a cold cohesive regime indicated by a thinner quasi-liquid layer.

Elastic and frictional properties of surface snow affect the saltation process and may also have an impact on the threshold conditions for transport. Particle rebound and expulsion from the snow surface are necessary for transport to proceed, hence factors that limit further application of particle momentum to saltation transport will raise the threshold wind speed condition. Mellor (1975) shows that Young's modulus of elasticity for snow is a function of snow temperature, increasing approximately 30% as the temperature drops from −1°C to −40°C. This increase in elasticity increases the force necessary to sufficiently deform particle bonds upon impact to cause failure and hence shatter and expel surface snow particles. Another property of snow with a relationship to temperature is the sliding or kinetic friction of snow, which refers to the ability to slide plates of snow crystals past each other. Langham (1981) presents Norwegian data that show a 60% increase in the sliding friction force as the temperature drops from 0°C to −25°C. As Perla and Glenne (1981) note, this kinetic friction force is not necessarily determined by the quasi-liquid layers surrounding ice crystals but by snow crystals acting as a "solid lubricant" analogous to graphite. At temperature below −25°C or so, snow cohesion due to viscous forces becomes minimal. However, as elasticity and kinetic friction increase with decreasing temperature, these parameters may contribute to a small increase in the shear stress necessary to initiate snow transport at very cold temperatures. It is therefore likely that a fourth, less well understood, snow property regime important to snow transport exists at very cold temperatures: a frictional resistance regime, governed by kinetic friction and elasticity.

As surmised from the snow properties reviewed here, it is extremely difficult to calculate snow surface resistance to wind erosion on a completely physical basis. Field data show that one important parameter, the surface bond strength, is extremely difficult to measure (Martinelli 1983). The meteorological and soil temperature sequences leading to different metamorphic states in the snowpack are notoriously complex as is evident from the experience of avalanche prediction programs. However, snow wetness, kinetic friction and elasticity, and even metamorphism are strongly related to temperature. It can be concluded that the major factors controlling snow bonding, cohesion, and frictional properties are related to snowpack temperature. Hence, threshold wind speed should be correlated with snow temperature. As the upper snow layers in open areas are subject to wind pumping and hence well mixed with respect to the overlying air (Colbeck 1989), snow surface temperature is strongly related to air temperature for exposed snow covers in windy conditions. This provides a foundation for physically based models of threshold conditions for snow transport that can use standard meteorological data.

3. Observations

The Atmospheric Environment Service (AES) of Environment Canada records the hourly occurrence of visual observations of "blowing snow" in its digital archives and "drifting snow" in the remarks column of its daily synoptic data sheets for many locations in its primary network of meteorological stations. The wind speed recorded for the hour at which snow transport is noted to begin can form a basis for a transport threshold dataset. Snow transport is observed through two weather types under AES manual observation guidelines.[1] Blowing snow is defined in the manual observations booklet as snow lifted by the wind such that it obscures visible range at "eye level" to less than 9.7 km. Drifting snow is a visual observation of snow moving along the ground and is relatively more subjective. Presuming a mean particle radius of 60 μm and Mie scattering of light, Pomeroy and Male's (1988) optical model of suspended snow predicts a mass concentration of 0.0181 g m^{-3} at

[1] For clarity the term snow transport will be used to denote either "drifting snow" or "blowing snow," which are defined separately by meteorological services but have no real physical distinction in terms of the snow transport initiation.

FIG. 1. Location of the meteorological stations in the Prairie Provinces of western Canada.

the visual range limit for blowing snow defined by AES. This mass concentration at 2-m height does not correspond to the cessation or initiation of transport, and does not even indicate a unique mass flux profile, because near-ground fluxes vary with surface roughness. The observed occurrence of blowing snow is therefore not necessarily the threshold for transport of snow. The observation of drifting snow provides a better indication of the transport threshold as drifting snow includes saltation fluxes near the ground when suspension of snow is negligible. The subjective nature of the drifting snow definition, rapid changes in wind speed, and the sharp decrease in visible range possible as the wind speed increases during a snow storm means that in practice some thresholds for transport will be based upon the initiation of blowing snow and others based on the initiation of drifting snow. Both criteria are combined in this analysis, which considers the initiation of snow transport as the most appropriate threshold dataset.

AES hourly meteorological observations from 16 stations in the Canadian prairies were used to examine the initiation of snow transport in relation to wind speed, instantaneous air temperature from November 1970 to April 1976. Digital records of the hourly occurrence of blowing snow were combined with digitized notations from the remarks column of daily synoptic records to produce a digital record of the hourly occurrence of snow transport. The stations, in the provinces of Alberta, Saskatchewan, and Manitoba (Fig. 1), represent regional climates from cool and subhumid with lower wind speeds and few midwinter melts in the northeast to warmer and semiarid with higher wind speeds, frequent midwinter melts, and chinooks (foehn-like winds) in the southwest. Table 1 shows the snow year climatic characteristics of the stations by the average hourly temperature and the average snowfall as water equivalent of the snow years, 1970–76. With the coverage of six snow years and 16 stations, the results from AES meteorological data can be confidently applied to the Canadian prairies and other regions with similar climates.

AES meteorological data provide the time over which snow transport occurs, wind speed, temperature, precipitation type (rain, snowfall), snow depth, and the occurrence of snowfall for each snow transport event on an hourly scale. The threshold wind speed for a single transport event must be somewhere between the current wind speed for the event and the previous wind speeds. AES air temperature is the dry-bulb temperature measured a 2-m height and wind speed is measured by a cup anemometer. Most of the anemometers at the 16 locations used were mounted at 10 m above the ground (Table 1) in a flat, open exposure such as at an airport, but some were not. Using the following method, wind speeds other than measured 10 m above the ground are converted to the 10-m wind speed as

$$U_z = \frac{u_*}{k} \ln\left(\frac{z}{z_o}\right), \tag{2}$$

where U_z (m s^{-1}) is wind speed at the height z (m); u_* (m s^{-1}) is friction velocity; k is von Kármán's constant (0.4); and z_o (m) is surface roughness length, which is 0.5–10.0 $\times$ 10^{-4} m (Oke 1978). For this conversion, z_o is assumed as 0.5 $\times$ 10^{-4} m. Given U_z, z, k, and z_o, u_*

TABLE 1. Meteorological stations used to examine the snow transport threshold condition.

Station	Province	Anemometer height (m) 70–71	71–72	72–73	73–74	74–75	75–76	Temperature (°C)	Snowfall (mm)
Calgary	Alberta	20.4	20.4	20.4	18.3	18.3	18.3	−4.7	122
Coronation	Alberta	13.4	13.4	13.4	13.4	13.4	13.4	−8.4	169
Medicine Hat	Alberta	10	10	10	10	10	10	−4.4	113
Peace River	Alberta	10	10	10	10	10	10	−10.3	155
Dauphin	Manitoba	10	10	10	10	10	10	−10.0	141
Portage La Prairie	Manitoba	9.1	9.1	9.1	9.1	10	10	−8.9	143
Winnipeg	Manitoba	10	10	10	10	10	10	−9.7	106
Broadview	Saskatchewan	10	10	10	10	10	10	−9.3	123
Estevan	Saskatchewan	10	10	10	10	10	10	−7.0	107
Moose Jaw	Saskatchewan	10	10	10	10	10	10	−7.1	127
North Battleford	Saskatchewan	11	11	11	11–14	10–14	10	−9.8	111
Prince Albert	Saskatchewan	10	10	10	10	10	10	−11.6	103
Regina	Saskatchewan	10	10	10	10	10	10	−8.9	113
Swift Current	Saskatchewan	10	10	10	10	10	10	−6.7	132
Wynyard	Saskatchewan	10	10	10	10	10	10	−9.9	126
Yorkton	Saskatchewan	10	10	10	10	10	10	−10.6	125

is found from Eq. (2). Then given u_*, $z = 10$ m, k and z_o, U_z is converted to U_{10}, the 10-m wind speed (m s^{-1}). Due to lack of information, the correction could not be made for the differential exposure of anemometers, though in general this is not a significant problem on the Canadian prairies.

4. Observed threshold wind speed variation

The threshold wind speed for snow transport is defined for this paper as the wind speed when the snow particles just start to move. For an hourly database of snow transport events, it is known that an event starts sometime within a specified hour; however, it is not known exactly when the event starts nor what are the wind speed and air temperature at that moment. To approximate the transport threshold condition, the wind speed and temperature of the previous hour and of the hour when transport was observed to begin were averaged and taken as the threshold wind speed and the threshold temperature for the snow transport event.

There is a physical difference between the threshold wind speeds for transport of wet snow and dry snow because of the substantially larger cohesion of wet snow. Wet snow is snow that has either received melting temperatures of 0°C or above, or wet precipitation in the form of freezing rain or freezing drizzle since the last snowfall. Wet snow, defined in this manner, behaves in the *wet cohesive regime* for snow transport. Dry snow may behave in the *warm cohesive regime, cold cohesive regime,* or *cold kinetic resistance regime.* Threshold wind speeds are therefore examined separately for wet snow and dry snow. Figure 2 shows the frequency of observed threshold wind speeds for these two snow conditions. The majority of thresholds were from 7 to 14 m s^{-1} for wet snow, higher than the range of 4 to 11 m s^{-1} for dry snow. On average, thresholds are found to be 9.9 m s^{-1} for wet snow and 7.7 m s^{-1} for dry snow. The absolute frequency of wet snow transport is very low, particularly for wet snow caused by melting events. Almost no wet snow transport events were recorded for snow that had undergone air temperatures greater than 5°C. Transport of dry snow occurred not only more frequently but also with a wide range of air temperatures; from 0° to −40°C or so. The threshold wind speed for wet snow transport varies substantially (Fig. 2) and is not highly related to the air temperature. A single, averaged value of 9.9 m s^{-1} may be appropriate as the estimate of threshold wind speed for wet snow transport.

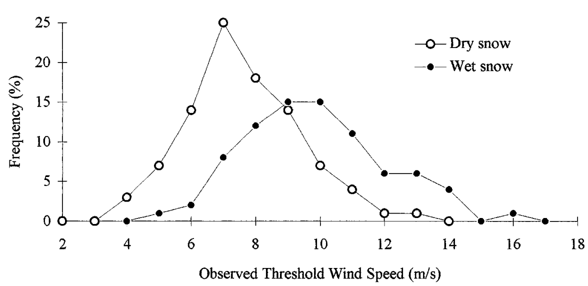

FIG. 2. Relative frequency of observed transport threshold wind speed for dry snow and wet snow for the 16 meteorological stations, 1970–76 snow years.

Dry snow is snow that has not received temperatures of 0°C or above, or wet precipitation since the last snowfall. Figure 3 shows the variation of threshold wind speeds with ambient air temperature for dry snow. Although the data are very scattered, there is the following pattern:

1) from roughly −25°C to 0°C, an increase in threshold wind speed with ambient air temperature is evident;
2) the rate at which the threshold wind speed increases becomes higher as ambient air temperature becomes warmer; and
3) at temperatures below roughly −25°C, the threshold wind speed tends to rise slightly as the temperature drops.

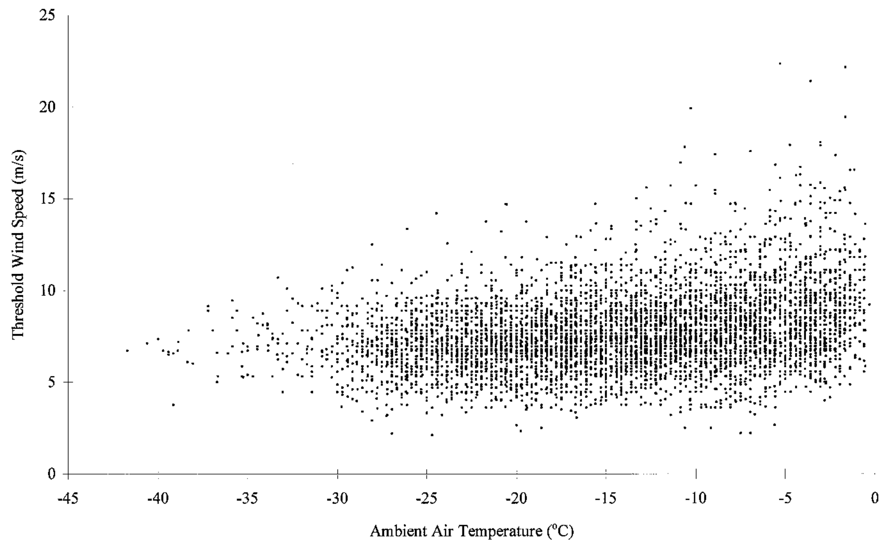

FIG. 3. Variation of individual transport threshold wind speeds with ambient air temperature for the 16 meteorological stations, 1970–76 snow years.

This pattern is illustrated for the relationship between the mean threshold wind speed and ambient air temperature in Fig. 4. Confidence limits were calculated according to observed data so that 95% of the data fall between the upper and lower limits as shown in Fig. 4. Note that the fluctuation of mean threshold wind speed is greater when the temperature is below $-30°C$; this is due to the smaller number of observations at these temperatures.

Figures 3 and 4 display a nonlinear relationship between mean threshold wind speed and ambient air temperature. Three temperature-associated resistance regimes for dry snow transport are observed, with the following relationships between ambient air temperature and threshold wind speed, tentatively:

1) warm cohesive regime ($T > -10°C$) with a higher positive slope,
2) cold cohesive regime ($-25°C < T < -10°C$) with a lower positive slope,
3) cold kinetic resistance regime ($T < -25°C$) with a negative slope. Here, T is ambient air temperature.

The third regime suggests that elastic and kinetic frictional forces, which increase with decreasing temperature, can dominate cohesion and resistance to movement

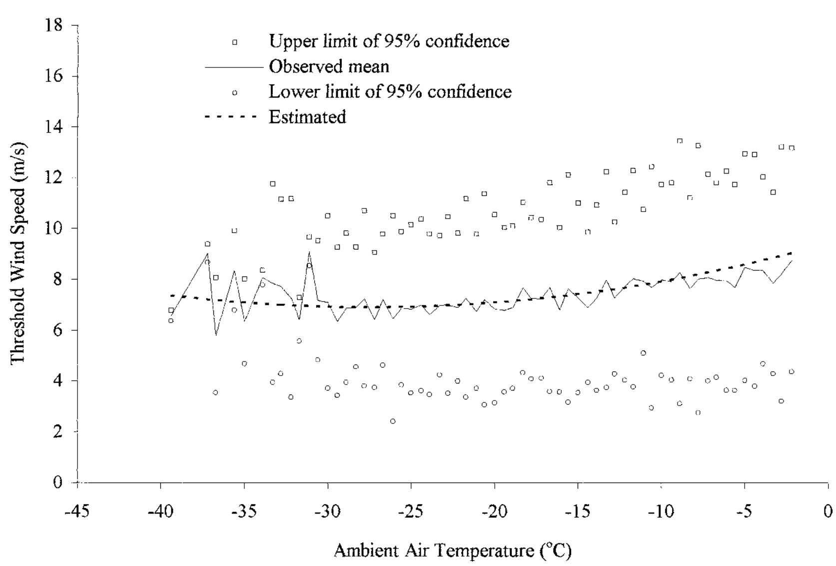

FIG. 4. Estimated threshold wind speed and observed mean threshold wind speed: variation with ambient air temperature.

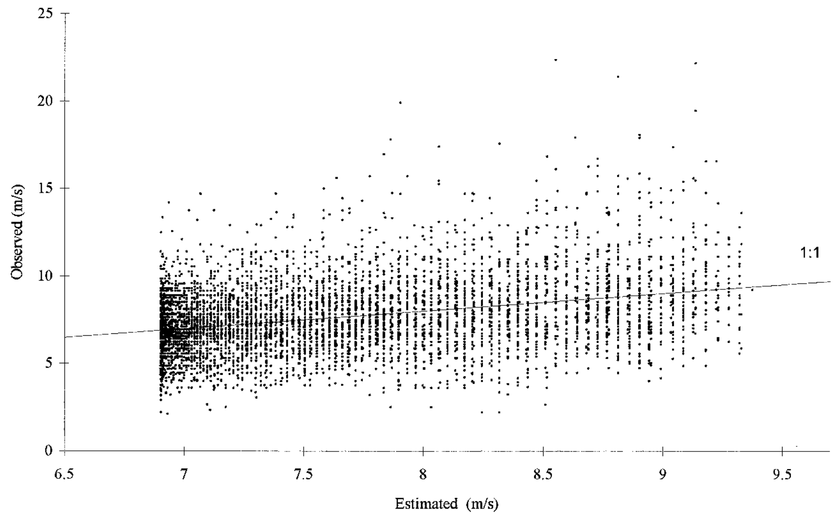

FIG. 5. Comparison between the estimated and observed individual threshold wind speeds.

at temperatures below −25°C where little or no quasi-liquid layer exists.

The exact temperature boundaries for these three regimes are unknown due to the gradual change of threshold wind speed with temperature and the above criteria are merely guidelines. The physical phenomena that underlie these regimes were discussed earlier in this paper. The measurements shown in Figs. 3 and 4 support the existence of the three transport threshold regimes. Hence, a common threshold wind speed for dry snow transport in various meteorological conditions does not exist, and a nonlinear variation in transport threshold wind speed with air temperature can be discerned. Overall, the threshold wind speed exhibits a large variability for the same threshold air temperature. This is expected to be due to the following:

1) properties of both falling and surface snow particles (e.g., particle size, shape, and density) vary with factors other than the current air temperature;
2) wind speeds vary around the recorded hourly value and gustiness is not well represented in meteorological records;
3) ambient air temperature at 2-m height above the ground is an approximation of the snow surface temperature;
4) snow surface roughness length varies from station to station and over time; and
5) the visual determination of ''drifting'' snow is qualitative and observers could have inconsistent criteria to identify it.

5. Snow transport threshold model

The physics of surface snow resistance in setting a threshold condition for snow transport are not sufficiently well understood to permit the development of a formal physical model. However, the data shown here suggest a conceptual model that can be quantified as a simple algorithm. The varied relationships between threshold wind speed and ambient air temperature reflect the balance between the driving force of wind shear stress and the resistant force of snow for three snow resistance regimes for dry snow transport. Using statistical techniques, the following relationship is indicated from the dataset examined:

$$U_t(10) = a + bT + cT^2, \tag{3}$$

where $U_t(10)$ (m s^{-1}) is observed individual threshold wind speed measured at 10 m height, T (°C) is the ambient air temperature at 2-m height, and a, b, and c are parameters: $a = 9.43$ m s^{-1}, $b = 0.18$ m °C^{-1} s^{-1}, and $c = 0.0033$ m °C^{-2} s^{-1}. The 95% confident limits are 9.27–9.60 for a, 0.16–0.2l for b, and 0.003–0.004 for c. The standard error of estimate by Eq. (3) is 1.97 m s^{-1}.

The performance of this algorithm in estimating the threshold wind speed for dry snow is illustrated in Figs. 4 and 5 where the mean threshold wind speed and the individual threshold wind speeds from observations are plotted against that estimated by Eq. (3) using observed meteorological conditions, *for only those periods when snow is reported to be on the ground.* While the agreement between the observed and estimated thresholds may be poor for individual snow transport events (Fig. 5), a strong agreement is demonstrated between the mean threshold of individual snow transport events and those estimated (Fig. 4). Larger errors in the estimated and the observed mean threshold correspond to the lower frequency of occurrence where the number of observations are insufficient to provide average threshold conditions.

6. Discussion

There are physical reasons for a difference between the threshold wind speeds for transport of fresh snow

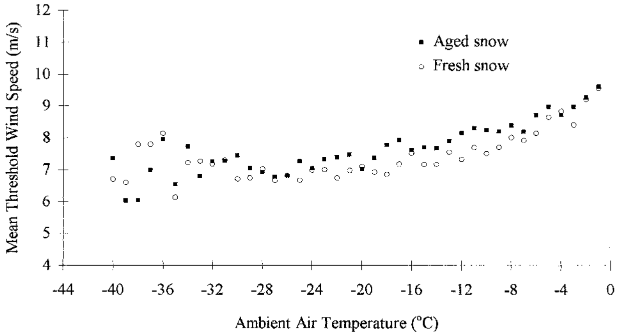

FIG. 6. Difference in the transport threshold wind speed between fresh snow and aged snow.

and aged snow as discussed earlier. An attempt was made to examine transport of fresh snow and aged snow separately in the present study. The data show that, on average, old snow requires a slightly higher wind speed to move. The mean threshold wind speed was 7.5 m s^{-1} for fresh snow, 8.0 m s^{-1} for aged snow, and 7.7 m s^{-1} for all cases. However, the difference in the transport threshold wind speed between aged snow and fresh snow for the same ambient temperature is not substantial, particularly when the temperature is below -25°C, or close to 0°C (Fig. 6). This may not indicate that the snow aging process has a negligible effect on the transport threshold because current and previous meteorological conditions are usually highly correlated to each other. The metamorphosis and sintering that harden snow occur more rapidly for higher temperatures and high wind speeds. As the independent variable is current temperature and the dependent variable is current wind speed, the autocorrelation of those variables over time means that past values need not be included in the predictive equation. The implication is that the snowpack resistance can be related to the current meteorological conditions, permitting Eq. (3) using the instantaneous temperature only to estimate the threshold wind speed.

7. Conclusions

The threshold wind speed largely depends on the snow particle bonding, cohesive, and kinetic resistance to transport, which are highly related to the snow surface temperature. This is in contrast to transport thresholds for most other aeolian and fluid particulates for which particle size is the most important factor. Threshold wind speed increases nonlinearly with ambient air temperature above -25°C, and is inversely proportional below -25°C. This inverse relation between threshold wind speeds and very cold temperatures is considered tentative because of the relatively small dataset available to test it. The majority of recorded threshold wind speeds ranges from 7 to 14 m s^{-1} with an average of 9.9 m s^{-1} for wet snow transport, and from 4 to 11 m s^{-1} with an average of 7.7 m s^{-1} for dry snow transport. The mean transport threshold wind speed is 7.5 m s^{-1} for fresh snow and 8.0 m s^{-1} for aged snow.

A physically based, empirical solution to the threshold wind speed based on standard meteorological parameters was developed from six years of hourly observations from 16 meteorological stations in Canadian prairies. The data indicate that the threshold wind speed for snow transport varies considerably even with the same ambient air temperature, but that the mean threshold is highly related to the ambient air temperature. The threshold wind speed estimated using the air temperature agrees well on an average basis. These results should be field tested with instantaneously measured transport thresholds. As the observed data show a notable variance in the occurrence of snow transport for the same temperature, a stochastic element may have to be considered in extrapolating point transport fluxes calculated with input from this threshold model to land surfaces. Future research should investigate the probability of occurrence of snow transport for specific meteorological conditions and consider this probability as a ''scaling'' parameter to extrapolate point model results to larger areas.

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