WIND TRANSPORT OF SEASONAL SNOWCOVERS

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ABSTRACT: A model of blowing snow is developed based on recent developments in transport theory. The response of the model to meteorological input parameters is demonstrated and results of the model compared to field measurements. The model agrees with these measurements for high transport levels. In open snowcovered areas it is shown that sublimation and vertical transport of airborne snow result in significant surface snow erosion rates. Blowing snow particles can abrade and entrain soil materials and other particulates. Understanding of the blowing snow phenomenon may therefore be important in modelling contaminant transport involving snow.

1. INDTRODUCTION

This paper demonstrates the assembly of present theory on blowing snow mechanics into a calculation procedure. The results of this calculation are examined in terms of their sensitivity to meteorological parameters and compared to measurements of the density of blowing snow in the atmosphere. The comparison reflects on the adequacy of present blowing snow theory and illuminates the sensitivity of the calculation procedure. Comments will be made suggesting the effects of blowing snow on both atmospheric and snowpack chemical transport.

2. BLOWING SNOW THEORY

2.1. Fully Developed Transport and Sublimation

Blowing snow is described as being "fully developed" when mean horizontal transport rates are steady with time and the maximum snow load is carried by the air for a constant snowcover, atmospheric velocity and density. Fully developed transport exists in the lower boundary layer where the blowing snow develops over at least 500 m of flat uniform terrain with consistent snowpack conditions. When fully developed, blowing snow may be modelled in terms of a control volume of unit

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horizontal area and 10 m height. Most blowing snow transport over flat terrain occurs below this height and particles transported above 10 m have little likelihood of deposition before substantial sublimation occurs (Male, 1980; Schmidt, 1982a).

Given equal horizontal inflows and outflows of both solid and vapour states of water through the control volume, the rate of erosion from the snowpack is equal to the vertical transport rates of snow and water vapour. Since surface evaporation rates are extremely low under cold mid-winter conditions (Male and Granger, 1981), the vertical flux of water vapour is assumed equal to the sublimation rate from blowing snow particles within the control volume. In the following theoretical development, the horizontal and vertical components of transport as well as the sublimation rate of blowing snow are considered in terms of fully developed flow through a control volume.

2.2. Horizontal Transport

Horizontal transport of snow by the wind occurs via saltation and turbulent suspension. The theories describing these modes of two-phase flow and the transfer of snow particles from one mode to the other are described separately.

2.2.1. Saltation. Saltation is the bouncing of particles as they are transported in curved trajectories near the surface. In this paper the use of the term also includes any particles rolling or creeping along the surface, though these constituents are minor for a cohesive material. The initiation of a "cascade" of saltating particles requires momentum transfer from atmospheric shear forces and the impact of "randomly entrained" particles (Schmidt, 1980). It is postulated that these randomly entrained particles enter transport due to periodic extremes of fluid shear or pressure fluctuations at the surface (Dyunin, 1954). Bagnold (1973), based on earlier work (Bagnold, 1941), has provided a physically based theory of saltation transport which has been developed for application in streams (Bridge and Dominic, 1984) and blowing snow (Schmidt, 1986). The development of Bagnold's ideas stated here differs from the above approaches in the interpretation of the "threshold of saltation" and in an attempt to make the calculations amenable to meteorological measurements.

Bagnold (1973) defines the "unsuspended transport rate" (called the saltation flux Q_{salt} hereafter) as,

$$Q_{salt} = W_p \, \bar{u}_p \quad , \tag{1}$$

where $\bar{u}_p$ is the mean horizontal particle speed and W_p is the immersed weight of saltating particles over a unit area of surface. Since the buoyancy of ice particles in air is negligible, W_p can be defined as

$$W_p = \eta_{salt} \, \bar{h} \, g \quad . \tag{2}$$

where η_{salt} is the drift density (mass of blowing snow particles per unit volume of atmosphere), $\bar{h}$ is the mean height of saltating particle trajectories and g is the gravitational constant.

The shear force or stress available to support the weight of saltating particles is found by partitioning the two-phase shear stresses. At the initiation of a saltation cascade (the saltation threshold), the shear stress on the snow surface τ_t is composed of a fluid stress τ_f and a stress imparted to the surface via randomly entrained particles already in transport τ_p, such that

$$\tau_t = \tau_f + \tau_p \quad . \tag{3}$$

Note that τ_p includes the stress required to overcome dynamic friction in the momentum transfer from entrained particles to surface crystals. Thus Bagnold's dynamic friction coefficient is included in the threshold shear stress and need not constitute a separate term.

Bagnold argues that shear stress in excess of that needed to overcome particle cohesion, inertia and dynamic friction supports the weight of saltating particles. Thus

$$W_p = \tau_s = \tau - \tau_t \quad , \tag{4}$$

where τ is the total shear stress ($\tau > \tau_t$) and τ_s is the shear stress in excess of that required to initiate a saltation cascade. Combining Eqs. 2 and 4 with the definition of the friction velocity [$u^* = (\tau/\rho)^{0.5}$] yields,

$$\eta_{salt} \, \bar{h} \, g = (u^{*2} - u_t^{*2})\rho \quad , \tag{5}$$

where ρ is the density of the two-phase flow and u_t^* is the friction velocity at the threshold of saltation. Assuming a constant shear stress in the atmospheric boundary layer, u^* and u_t^* can be measured well above the saltation layer, at heights where ρ = atmospheric density.

The velocity of saltation $\bar{u}_p$ can be estimated by extrapolating the logarithmic wind profile into the saltation layer. This is done with reservation, as observations of wind velocities in saltation are limited. Analysis of particle trajectories in saltation (Kobayashi, 1972; White, 1982) show that most horizontal acceleration occurs at the top of particle trajectories. Measurements of individual snow particle trajectory heights by Kikuchi (1981) show no relationship to measured u^* ($r^2 = 0.015$), in contradiction of Owen's (1964) predictions. In the absence of acceptable theory on the mechanics of individual particles in saltation, it is suggested that $\bar{h}$ = 0.01 m. Measurements by Kobayashi (1972) and Kikuchi (1981) suggest this as a mean saltation height in blowing snow. Assuming that the effects of particle drag and preferential acceleration in the faster wind at the top of the saltation layer counteract one-another, a mean horizontal particle speed equal to the predicted wind speed at 0.5 $\bar{h}$ is suggested. Employing these assumptions,

$\bar{u}_p$ is found as

$$\bar{u}_p = \frac{u*'}{k} \ln(0.5\bar{h}/z_o) \quad , \tag{6}$$

where u*' is the friction velocity in the saltation layer, k is the von Kármán constant (0.4), $\bar{h}$ equals 0.01 m and z_o is the aerodynamic roughness height.

Combining Eqs. 1, 2, 5 and 6 provides an expression for the saltation flux;

$$Q_{salt} = \frac{(u*^2 - u_t*^2)\ u*'\ \ln(0.5\bar{h}/z_o)}{\rho_a\ k} \quad . \tag{7}$$

The two-phase friction velocity u*' can be solved in terms of the snow-free friction velocity u*, atmospheric density ρ_a and drift density in saltation η_{salt} where;

$$u*' = \frac{u*(\rho_a)^{0.5}}{(\rho_a + \eta_{salt})^{0.5}} \quad . \tag{8}$$

η_{salt} can be found using u*, Eq. 5 and the assumption of saltation layer height, where,

$$\eta_{salt} = \frac{u*^2 - u_t*^2}{\rho_a\ \bar{h}\ g} \quad . \tag{9}$$

This provides a solution for Q_{salt} in terms of the mean wind speed profiles measured above the saltation layer. Where;

$$Q_{salt} = \frac{(u*^3 - u*\ u_t*^2)\ \ln(0.5\bar{h}/z_o)}{k\left[\rho_a{}^2 + \dfrac{u*^2 - u_t*^2}{\bar{h}\ g}\right]^{0.5}} \quad . \tag{10}$$

Note that the form of this relationship bears some resemblance to Bagnold's original (1941) equation where Q_{salt} varies as the cube of the friction velocity.

2.2.2. <u>Suspension</u>. Suspended horizontal transport of blowing snow occurs when particles are supported by the vertical component of atmospheric turbulence. The vertical gradient of suspended snow can be modelled in analogy to turbulent diffusion of a gas, except blowing snow particles have appreciable terminal fall velocities in still air. Thus the characteristics of atmospheric turbulence and particle fall velocities must be known to calculate the suspended drift density at some height.

Budd's (1966) expressions estimate suspended snow transport based on the vertical diffusion of non-uniform particles. These expresssions require a reference drift density, with greater accuracy achieved when the reference is near the ground. If a particle at the top of its saltating trajectory encounters an upward vertical velocity exceeding its fall velocity, the particle will become suspended and available for vertical turbulent diffusion. The probability that a saltating particle will enter this lowest "reference" suspended particle layer can be calculated from the saltating particle drag, size distribution and the vertical turbulence spectrum near the ground.

Hunt and Weber (1979) in examining the diffusion of particles from ground level sources, note the inhomogeneity of vertical turbulence near the ground. Near a boundary, the experimentally determined mean vertical velocity $\bar{w}$ is upward and a function of the friction velocity where,

$$\bar{w} = 0.4\ u*' \quad . \tag{11}$$

The vertical turbulent velocities are normally distributed with the standard deviation also a function of u*', where

$$\sigma_w = 1.3\ u*' \quad . \tag{12}$$

Note that away from the influence of the ground $\bar{w} = 0$ and $\sigma_w = 1.1\ u*$ (Panofsky and McCormick, 1960). If we consider the top of the saltation layer near enough to the ground for Hunt and Weber's relationship to apply, the probability that a vertical turbulent velocity will exceed a particle's fall velocity can be transformed into a standard normal distribution where,

$$P(w' \geq \omega') = 1 - (2\pi)^{-.5} \int_{-\infty}^{\omega'} e^{-x^2/2}\, dx \quad , \tag{13}$$

and

$$w' = (w - \bar{w}/\sigma_w) \text{ and } \omega' = (\omega - \bar{w}/\sigma_w).$$

For some frequency distribution of particle radii in saltation $f(P_r)$, the drift density η in the lowest suspended layer z_1 can be found as a function of the drift density in saltation η_{salt} and the ratio of particle volumes in the lowest suspended layer to that in the saltation layer where,

$$\eta_{z_1} = \eta_{salt} \frac{\int_0^\infty P_r^3 \, f(P_r) \, P(w' \geq \omega') \, dP_r}{\int_0^\infty P_r^3 \, f(P_r) \, dP_r} \quad . \qquad 14)$$

The distribution of particle radii in saltation has been shown by Schmidt (1981) to fit a two parameter gamma distribution proposed by Budd (1966) for suspended blowing snow. The form of this distribution (Haan, 1977) is

$$f(P_r) = \frac{P_r^{(\alpha-1)}}{\beta^\alpha \, \Gamma(\alpha)} \, e^{(-P_r/\beta)} \quad , \qquad 15)$$

where Γ is a gamma function, α is a shape paremeter and β is a scale parameter such that the mean particle radius equals $\alpha\beta$.

Lee (1975) demonstrates that the terminal fall velocity for blowing snow particles is best calculated using Carrier's (1953) drag equations. A simple power law equation fitting this calculation is

$$\omega = 1.1 \cdot 10^7 \, P_r^{\,1.8} \, , \; (r^2 = 0.991) \qquad 16)$$

where P_r is the particle radius (m) and ω is the terminal fall velocity (m/s). This allows characterization of the probability in Eq. 13 in terms of the particle radius and friction velocity.

Substituting the gamma distribution and solving ω' in terms of the particle radius and ambient friction velocity yields,

$$\eta_{z_1} = \frac{\eta_{salt} \int_0^\infty P_r^{\,(\alpha+2)} e^{(-P_r/\beta)} [1-(2\pi)^{-.5} \int_{-\infty}^{\gamma} e^{-x^2/2} dx] dP_r}{\int_0^\infty P_r^{\,(\alpha+2)} \, e^{(-P_r/\beta)} \, dP_r} \quad , \qquad 17)$$

where $\gamma = (1.1 \cdot 10^7 \, P_r^{\,1.8} - 0.4 \, u*')/(1.3 \, u*')$. A transfer coefficient T_c between the saltating and lowest suspended drift density is defined as

$$T_c = \eta_{z_1}/\eta_{salt} \qquad 18)$$

Combining Eqs. 17 and 18 and integrating where possible gives the transfer coefficient as a function of the friction velocity and particle radius distribution in saltation. Thus

$$T_c = \frac{1}{(\alpha+2)! \; \beta^{(\alpha+3)}} \int_0^\infty P_r^{\,(\alpha+2)} e^{(-P_r/\beta)} [1 - (2\pi)^{-.5} \int_{-\infty}^{\gamma} e^{-x^2/2} \, dx] dP_r . \qquad 19)$$

The height z_1 is set 0.01 m above $\bar{h}$, reflecting the wide variance of saltating trajectory heights about $\bar{h}$. It is expected that some suspended particles are found at height $\bar{h}$ and some saltating particles at height z_1. We assume that this continuum is adequately represented by interpolation between the discrete drift densities predicted by the transfer coefficient.

Based on the assumption that the eddy diffusivity equals the diffusivity of blowing snow, Shiotani and Arai (1953) develop an expression relating the drift density η at some height z to a reference drift density at z_1 where

$$\eta_z = \eta_{z_1} \, (z/z_1)^{-\frac{\bar{\omega}}{ku*}} \quad . \qquad 20)$$

Budd (1966) shows that this relationship must be integrated over the particle size distribution, which varies with height. His relationship, corrected for flow densities in the reference suspended layer is

$$\eta_z = \eta_{z_1} [1 + \frac{\beta_\omega}{ku*'} \; \ln(z/z_1)]^{-(\alpha_\omega+3)} \quad , \qquad 21)$$

where $u*'$ is the friction velocity at height z_1 and α_ω and β_ω are the parameters for the gamma distribution of fall velocities at z_1. The gamma distribution of fall velocities at z_1 is found by transforming the particle radius distribution as shown by Pomeroy and Male (1986). The transformation assumes $\alpha = 15$ for suspended particle radii and solves for $\bar{P}_r$ at z_1 using a relationship proposed by Radok (1968) based on the measurements of Budd et al. (1966). The parameters describing the distribution of fall velocities at z_1 are $\alpha_\omega = 4.8$ and $\beta_\omega = 0.1375 \, u*'$. Note that the drift density profile is not dependent on $u*$ once the reference drift density is established. Barenblatt (1953) (as quoted by Monin and Yaglom (1965)) suggests that vertical density gradients in twophase flow create an effect similar to atmospheric stability caused by temperature gradients. While Budd's model does not address this effect, in our application friction velocities at z_1 are used to establish the reference drift density and particle size distribution. Shear stress is constant and flow density varies with height in blowing snow. However significant deviations from normal atmospheric density are limited to the lowest suspended and the saltation layer. By using $u*'$ in calculating η_{salt} and η and β_ω at height z_1, we feel that effects due to flow density gradients have been accounted for.

The mean horizontal velocity of a suspended particle is equal to the mean horizontal windspeed (Schmidt, 1982b). The horizontal mass flux of suspended snow per unit area perpendicular to the wind is therefore

$$Q_z = \eta_z u_z \quad . \qquad 22)$$

The windspeed at height z is calculated using the logarithmic profile with corrections for atmospheric stability (Webb, 1970) if necessary. The friction velocity used in calculating windspeed should be corrected for local flow density where,

$$u*' = \frac{u* \rho_a^{.5}}{(\rho_a + \eta_z)^{.5}} \qquad 23)$$

nd u* is measured at a height where η_z is negligible. Assuming a modified logarithmic profile can be used, the total suspended horizontal mass flux is

$$Q_{susp} = \frac{u*' \eta_{z_1}}{k} \int_{z_1}^{\infty} \ln(z/z_o)(1 + \frac{\beta_\omega}{k u*'_{z_1}} \ln(z/z_1))^{-(\alpha_\omega+3)} dz \quad , \qquad 24)$$

where $u*'_{z_1}$ denotes u*' at z_1. In the control volume, this equation is integrated to 10 m. The sum of Q_{salt} and Q_{susp} represent the horizontal blowing snow flux through the control volume.

2.3. Vertical Transport

An expression for the vertical transport rate through a unit area, F_s is developed by Shiotani and Arai (1967) based on their turbulent diffusion formulation. It has been applied by Porch and Gillette (1977) using dust and is,

$$F_{s_z} = \eta_z \bar{\omega}_z \quad . \qquad 25)$$

Using Eq. 16 to convert from fall velocity to particle radius and integrating over a gamma distribution of particle radii yields

$$F_{s_z} = \frac{1.1 \cdot 10^7 \eta_z (\alpha + 3.8)! \beta_z^{1.8}}{(\alpha + 2)!} \quad . \qquad 26)$$

Assuming α = 15, β_z can be found using Budd's (1966) relationship for the variation of β with height, where

$$\beta_z = 1.882 \cdot 10^{-5} \left(1/\beta_{\omega_{z_1}} + \frac{\ln(z/z_1)}{k u*'} \right)^{-0.556} \quad . \qquad 27)$$

For the calculation of F_s at the top of the control volume z = 10 m is used.

2.4. Sublimation

Increased ratios of surface area to mass and rates of ventilation result in sublimation rates from blowing snow being much greater than that of surface snow under similar environmental conditions. Dyunin (1961) initially outlined a method for the calculation of sublimation from blowing snow particles. Schmidt's (1972) comprehensive calculations are based on heat and water vapour transfer equations and the results of Thorpe and Mason's (1966) experiments on sublimation from ice spheres. In Schmidt's model the snow particles are considered ice spheres at the "wet bulb" temperature. Lee (1975) has rigorously evaluated the physics of the process and recommends turbulent atmospheric velocities relative to the ice spheres be used in determining ventilation rates. In this application, Lee's recommendations for modifications to Schmidt's model are made.

The sublimation rate from a single blowing snow particle dm/dt can be calculated as

$$dm/dt = \frac{2\pi P_r \sigma}{\frac{r}{\lambda T Nu}\left(\frac{rM}{RT} - 1\right) + \frac{1}{D \rho_s Sh}} \quad , \qquad 28)$$

where σ = the undersaturation of water vapour in the atmosphere,
r = latent heat of sublimation from ice,
M = molecular weight of water,
R = universal gas constant,
T = ambient atmospheric temperature,
λ = thermal conductivity of the atmosphere,
D = diffusivity of water vapour in the atmosphere,
ρ_s = saturation density of water vapour at T,
Nu = the Nusselt number,
and Sh = the Sherwood number.

The terms λ, D and ρ_s are sensitive to T under blowing snow conditions, while σ is sensitive to the relative humidity and T. The dimensionless Nusselt and Sherwood numbers are shown by Lee (1975) to be approximately equal for blowing snow and are predicted by the particle Reynolds number Re where,

$$Nu = Sh = 1.79 + 0.606Re^{0.5} \quad . \tag{29}$$

The particle Reynolds number is a function of its radius and relative ventilation velocity V, where

$$Re = \frac{2\ P_r\ V}{\nu} \tag{30}$$

and ν is the kinematic viscosity of the atmosphere. The ventilation rate of a suspended particle is equal to its fall velocity plus the vector sum of the root mean square turbulent wind velocities relative to the particle. For saltating particles it is assumed that mean relative velocities are much greater than turbulent relative velocities. Thus the saltating ventilation velocity is equal to the mean vertical and horizontal components of particle velocity relative to the mean wind velocity. Formulae to calculate these velocities and other parameters of Eq. 28 are presented in detail in Pomeroy and Male (1986).

The sublimation rate per unit volume of atmosphere $d\eta/dt$ for uniform conditions of wind velocity, shear stress, temperature, drift density and water vapour pressure and some distribution of particle radii is

$$d\eta/dt = N \int_0^\infty (dm/dt)\ f(P_r)\ dP_r \quad , \tag{31}$$

where N is the number of snow particles per m^3 of air. Solving for N, integrating over the gamma distribution and over height yields a general formula,

$$Q_{subl} = \frac{3}{4\pi\rho_i} \int_0^\infty \frac{\eta_z}{(\alpha+2)!\ \beta_z^{(\alpha+3)}} \int_0^\infty (dm/dt)\ P_r^{(\alpha-1)}\ e^{(-P_r/\beta_z)} dP_r\ dz\ . \tag{32}$$

Appropriate values of T, σ and windspeed are required to calculate (dm/dt) for each z. For calculation of sublimation rates in the control volume Eq. 32 is integrated over z from 0 to 10 m.

3. APPLICATION OF THEORY

The relationships describing transport and sublimation can be compiled to produce a physically based model of blowing snow. The model utilizes the framework of a control volume for which the horizontal and vertical snow mass transfers are calculated. A logarithmic wind profile modified for the effect of flow density gradients and a neutral temperature profile are assumed. A decrease in relative humidity RH with distance from the snow surface is also included. This decrease is simulated by the gradient

$$\sigma_z = \sigma_{2m}(1.02 - 0.027\ \ln(z)) \quad , \tag{33}$$

which results in an RH approximately 10% higher at 0.05 m than at 2 m height. These assumptions are generally supported by two winters of boundary layer measurements of wind, temperature and humidity during blowing snow at Loreburn.

In applying the model, reference windspeeds are set at 10 m with reference air temperatures and relative humidities at 2 m. These heights correspond to government meteorological standards in Canada and other countries. An aerodynamic snow surface roughness of z_o = 0.002 m (based on measurements over a non-vegetated snowcovered plain) is used to calculate friction velocities from these windspeeds. Thus a threshold u_{10} of 4.5 m/s corresponds to a u_t^* of 0.2113 m/s. Figure 1 shows vertical profiles of drift density in suspension and the mean drift density in saltation for a threshold u_{10} of 4.5 m/s, air temperature of -15 °C and u_{10} of 5, 7.5 10 and 15 m/s. Note that the drift density increases 10 fold for an increase in the u_{10} from 5 to 7.5 m/s and less than 5 fold for an increase from 10 to 15 m/s. This indicates that the vertical gradient of drift density becomes steeper with increasing windspeed. The vertical gradients in the saltation layer (< 0.01 m) are a result of the modelling procedure and do not reflect an actual condition.

The saltating and suspended mass fluxes to 10 m are shown in Fig. 2 as a function of windspeeds for two threshold windspeeds. Note that below a u_{10} of 9 m/s the ratio of saltating to suspended transport is approximately 10 or greater. However at windspeeds of 15 m/s this ratio has declined to less than three. Thus suspended transport increases its relative importance with increasing windspeed. At the higher threshold windspeed the ratio of saltating to suspended transport declines somewhat faster.

The total horizontal snow flux from 0 to 10 m is shown in Fig. 3 for various threshold conditions as a function of the 10 m windspeed. Note that the differing threshold windspeeds have little impact on mass fluxes at higher windspeeds. This indicates that the shear stress which is applied in overcoming surface snow particle cohesion, inertia and dynamic friction has become extremely small in comparison to the total shear stress at these windspeeds. However at lower windspeeds the mass flux is extremely sensitive to the threshold conditions.

Rates of sublimation from blowing snow particles within the control volume at a constant air temperature of -15 °C and relative humidities at 2 m of 40, 70 and 90% are shown as a function of the 10 m windspeed in Fig. 4. Increasing the relative humidity from 40 to 90% decreases the sublimation rate eight fold. The sublimation rates for a constant relative humidity of 70% and air temperatures of -35, -15 and -1 °C are shown as a function of the 10 m windspeed in Fig. 5. Increasing the air temperature from -35 to -1 °C results in a 20 fold increase in the sublimation rate. At a windspeed of 15 m/s, sublimation rates at -1 °C correspond to 1.8 mm of snow water equivalent per day. Note that this sublimation is only that which occurs in the lower 10 m of the atmosphere.

Figure 6 shows the vertical mass flux of snow through a unit area at 10 m height as a function of the 10 m windspeed. The vertical flux is calculated for three threshold conditions and while extremely sensitive to this parameter at lower windspeeds, it is relatively independent of the thresholds at high windspeeds. Daily values of the vertical flux are approximately equal to sublimation rates below 10 m for windspeeds of 15 m/s and higher, when temperatures are above -10 °C. Since the erosion rate at the snowpack surface must balance the vertical transport of water in solid and vapour states for fully developed flow, the snowpack erosion rate is equal to the sum of the sublimation and vertical mass transport rates. Thus, for a wind speed of 15 m/s, RH = 70% and T = -1 °C, an erosion of 3.6 mm of snow water equivalent per day would be expected. Particles vertically transported above 10 m are assumed to be no longer available for local deposition because of small particle size and high sublimation rates. Particles this small would only rarely be transported downward in the strong winds at 10 m.

4. MEASUREMENTS

4.1. Site

Measurements were carried out six km west of the village of Loreburn, Saskatchewan, located in the cold semi-arid grasslands of western Canada. The site is a relatively treeless plain with extensive grain growing agricultural land use. The nearest obstructions to blowing snow are a farmstead 1.8 km northwest and a small tree-fringed pond 1.2 km northeast of the site. In general, meteorological fetch is excellent. At the time of measurement, terrain within 1 km of the site was bare soil covered with a shallow snowpack.

4.2. Instrumentation

Measurements were made of mean horizontal windspeed and blowing snow particle flux at six heights, logarithmically spaced from 0.03 to 3 m and temperature and humidity at five heights, similarly spaced from 0.1 to 2 m. The instruments were located on a mast with 1.5 m metal arms connected to a central post. The mast could be raised and lowered with snowdepth and rotated toward the prevailing wind direction. "Climatronics" small alloy cup anemometers with ± 0.1 m/s precision were used to measure windspeed. Shielded and naturally aspirated thermilinear thermisters made by Yellow Springs Instrument Co. with ± 0.05 °C precision were used to measure air temperature. Shielded and uniformly aspirated "Honeywell" lithium chloride dewcells with ± 0.15 °C of dewpoint precision were used to determine absolute humidity. A blowing snow particle detector designed and built in the Division of Hydrology counts the number of particles interrupting a narrow infrared beam. The gauge is calibrated using Mie light scattering theory. For details of its design, calibration and performance see Pomeroy, Brown and Male (1985).

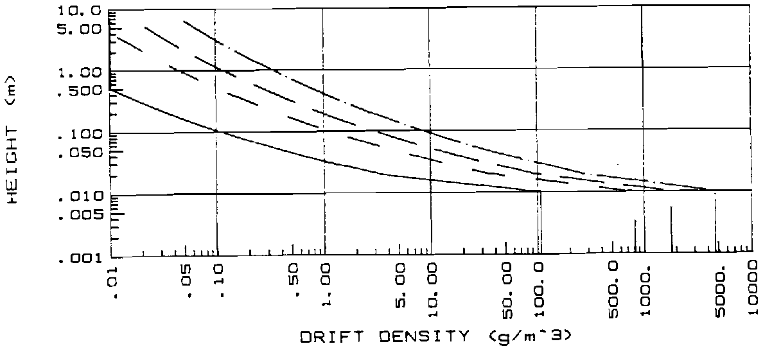

Figure 1. Vertical profiles of drift density. T = -15 °C, threshold u_{10} = 4.5 m/s, —— u_{10} = 5 m/s, ---- u_{10} = 7.5 m/s, — — — u_{10} = 10 m/s, — • — • u_{10} = 15 m/s.

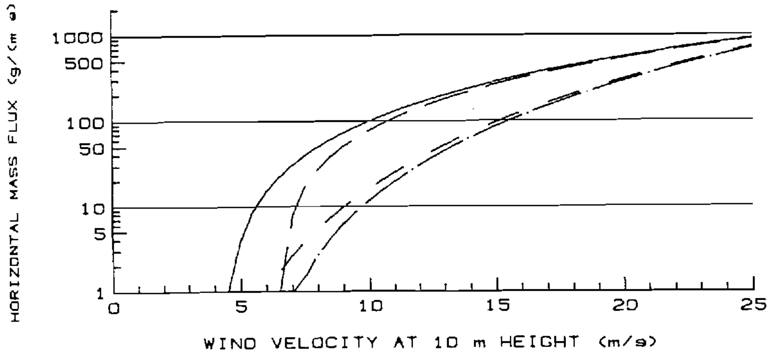

Figure 2. The horizontal flux of blowing snow carried in saltation and suspension. T = -15 °C; threshold u_{10} = 4.5 m/s, —— saltation, ---- suspension; threshold u_{10} = 6.5 m/s, — — — saltation, — • — • suspension.

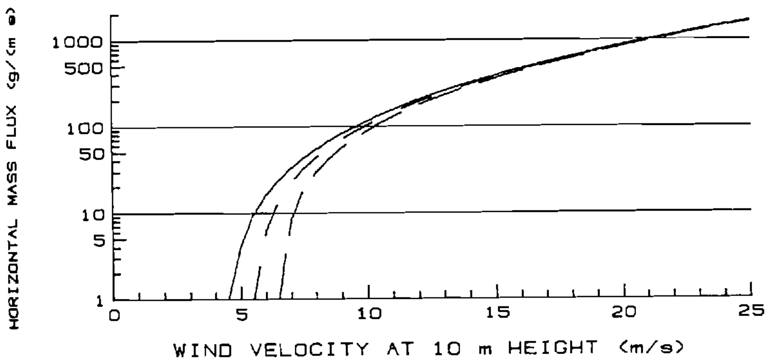

Figure 3. The horizontal flux of blowing snow to a height of 10 m. ——— threshold u_{10} = 4.5 m/s, ---- threshold u_{10} = 5.5 m/s, —— —— threshold u_{10} = 6.5 m/s.

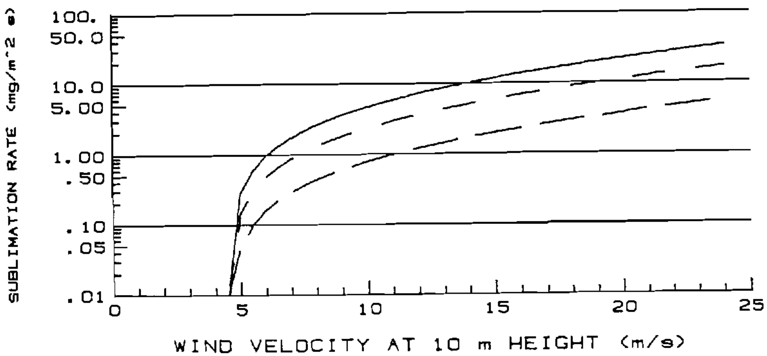

Figure 4. The rate of sublimation in a 10 m column of blowing snow. T = -15 °C, ——— RH = 40%, ---- RH = 70%, —— —— —— RH = 90%.

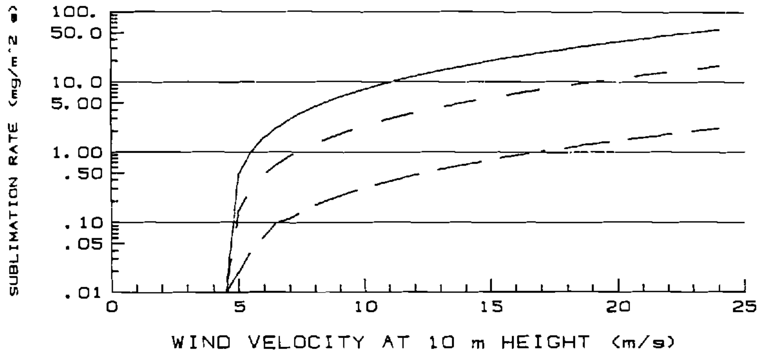

Figure 5. The rate of sublimation as a function of air temperature. RH = 70%, ——— T = -1 °C, ---- T = -15 °C, —— —— —— T = -35 °C.

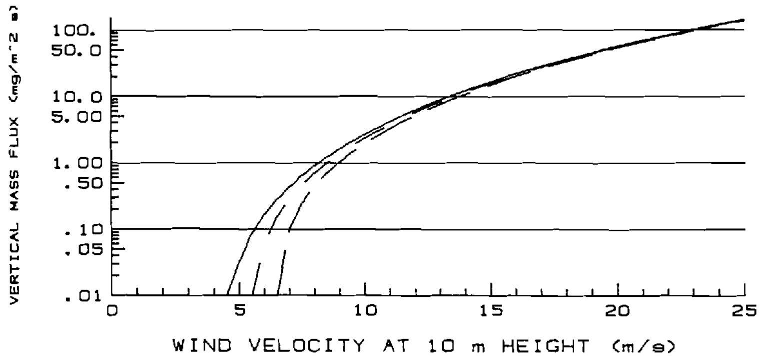

Figure 6. The vertical flux of blowing snow at 10 m height. T = -15 °C, ——— threshold u_{10} = 4.5 m/s, ---- threshold u_{10} = 5.5 m/s, —— —— —— threshold u_{10} = 6.5 m/s.

4.3. Measurement Analysis

Output of the instruments is digitized and averaged over 7.5 minute periods. The average values are stored on magnetic tape. The snow particle flux is corrected for the particle detector sampling area and divided by the corresponding windspeed to convert to drift density. Relative humidity is calculated using the dewpoint and air temperature. Richardson numbers are calculated from wind and temperature profiles, if $|Ri| > 0.005$ then Monin-Obukhov lengths are calculated. The logarithmic wind profile is corrected for stability if necessary and friction velocities and aerodynamic roughness lengths calculated.

4.4. Blowing Snow Events

Less than ideal conditions for blowing snow in the winter of 1985-86 resulted in only two days of blowing snow for which we are confident in the operation of all the instrumentation. Conditions during these two events are summarized below.

20 Feb 1986
Blowing Snow: from 0900H to 2000H
Temperatures: -32 °C at 0900H, high of -19 °C, rapid decline @ 2000H
Snow Depth: average over 300 m of bare soil with snow: 0.091 m @ 1600H
N.R.C. Canadian Hardness Gauge: 10-50, small probe @ 1500H
Snow Condition: snowfall on 17th, little drifting since then.
u_{10} threshold: 4.8 m/s
Surface Roughness: $\bar{z}_o$ = 0.003526 m

21 Feb 1986
Blowing Snow: from 0400H to 1800H
Temperatures: -14 °C @ 0400H to -25 °C @ 0900H, -20 °C @ 1800H
Snow Depth: survey identical to 20/2, 0.081 m @ 1500H
N.R.C. Canadian Hardness Gauge: 60 to +100, small probe @ 1400H
Snow Condition: no new snowfall, some windslab formation.
u_{10} threshold: 5.55 m/s @ 0400H, 5.7 m/s @ 1800H
Surface Roughness: $\bar{z}_o$ = 0.00163 m

5. COMPARISON OF MODELLED AND MEASURED RESULTS

A simulation of conditions at Loreburn can be made by using measured temperatures, humidities and windspeeds in the equations for blowing snow transport and calculating the resulting vertical profiles of drift density. Two simulations were run on data from 20 Feb. and three from 21 Feb. 1986. Measured values comprise 25 min to 1 hour averages of data collected as described in Section 2.

The assumption of a neutral wind profile is met approximately for these periods. On 20 Feb. the boundary layer is essentially neutral ($Ri = -0.0023$) and is very slightly unstable on 21 Feb. ($Ri = -0.0083$). Humidity gradients are indicative of near surface evaporation on both days with the increase in RH from 2 to .1 m being 2-3% and 3-4% on the 20th and 21st Feb. respectively. Mean windspeed gradients are not well represented by the logarithmic wind profile which was corrected for stability and atmospheric density. Calculations of z_o and $u*$ are made from wind measurements at .5, 1, 2 and 3 m, after corrections involving the Monin-Obukhov length. Reference windspeed, surface roughness, temperature and humidity values used in the simulation are shown in Table I.

TABLE I. Input data for blowing snow simulation.

Date	Time	u_{10}(m/s)	u_{10} threshold(m/s)	z_o(m)	T(°C @ 2m)	RH(% @ 2m)
20/2	1500	9.12	4.8	.003526	-20.3	63.0
	1790	8.81	4.8	.003526	-19.0	62.1
21/2	1080	8.19	5.7	.0016	-25.7	61.7
	1360	6.2	5.7	.0016	-23.4	59.5
	1700	6.5	5.7	.0016	-21.0	56.9

Results of the simulation on 20 Feb. 1986 are shown in Fig. 7 with the corresponding measured drift densities plotted as points. Measured and simulated results agree reasonably well for the drift densities at lower heights, but differences of a factor of two are present for drift densities above 1 m. The simulation on 21 Feb. 1986 is shown with corresponding measured drift densities in Fig. 8. Measured drift densities are under-estimated at lower heights for all winds and at upper heights for strong wind speeds. At wind speeds near the threshold velocities, measured drift densities at upper heights are over-estimated. It appears that for conditions of relatively high ambient windspeeds and low threshold windspeeds the model predicts the lower heights of suspended drift densities well. However, when windspeeds approach threshold levels, estimation of drift densities can be in error by an order of magnitude. Since no data are available for saltation drift density, that component of the model is not evaluated.

6. DISCUSSION

A very limited test of the model of blowing snow transport has shown good predictive capability for drift densities below 1 m when windspeeds are not near threshold values. However for the small drift densities encountered in near threshold conditions the model demonstrates poor predictive capabilities. Sources of error could include estimation of threshold windspeeds, instrument measurement error, assumptions on snow particle size distribution, equality of the diffusivities of momentum and snow particles and characterization of the saltation flux. The blowing snow model is a one-dimensional respresentation of a complex three-dimensional phenomenon and is subject to the limitations of such a

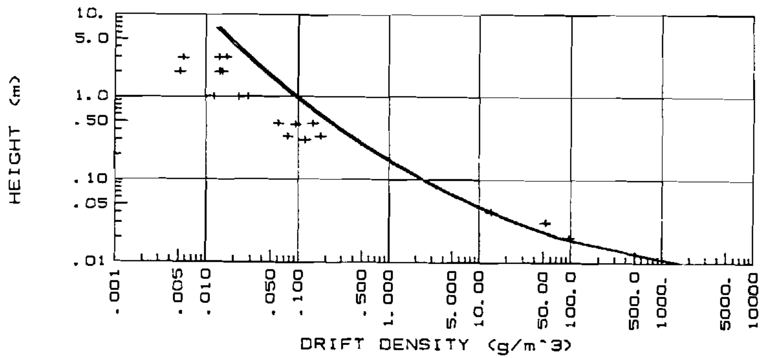

Figure 7. Drift densities measured at Loreburn and those predicted by the model, 20 Feb. 1986. Measured drift densities + @ $u_{10} \approx 8.0$–8.6 m/s; ——— predicted drift densities @ $u_{10} \approx 8.2$ and 8.6 m/s.

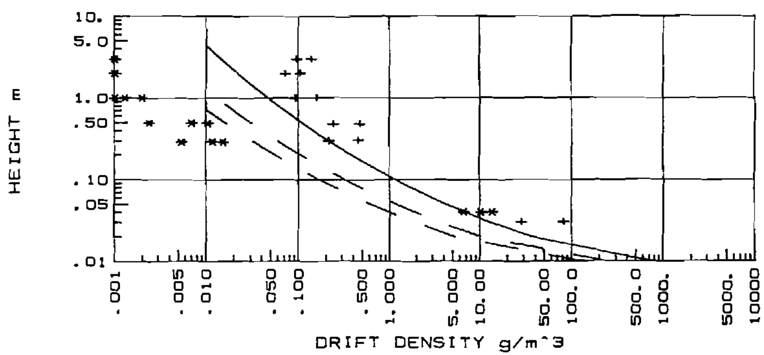

Figure 8. Drift densities measured at Loreburn and those predicted by the model, 21 Feb. 1986. Measured drift densities + @ $u_{10} \approx 8.0$ m/s, * @ $u_{10} \approx 6.0$–6.5 m/s; predicted drift densities ——— @ $u_{10} \approx 8.0$ m/s, ---- @ $u_{10} \approx 6.0$ and 6.5 m/s.

simplification. Nevertheless the success of modelling the higher drift densities is encouraging, as these values dominate the calculation of daily blowing snow transport quantities. Thus the major problem with the model is the sensitivity to threshold conditions. Figure 3 demonstrates this, where a small change in threshold conditions results in a large change in the horizontal mass flux at low windspeeds.

Quantities of blowing snow transported vertically and sublimated can be significant when a blowing snow event persists for several hours. In the western Canadian grain growing region up to several hundred hours per month of blowing snow are possible in the winter. This can be sufficient to completely ablate a 40 cm deep snowcover within one month.

Blowing snow has the potential to both scavenge small solid particulates from the atmosphere and to enhance the wind transport of the soil material. In the first instance the repeated re-entrainment and transport through the atmosphere of surface snow may allow these snow particles to scavenge small particulates while in transport. In the second instance, saltating snow particles impart a force to any exposed surface material. Where soil is exposed to saltating particles (as in blowing snow as a snowcover nears depletion) soil particles are abraded from the surface and enter saltation (Male, 1984). The evidence of this blowing soil is frequently found in "dirty" snowdrifts in Saskatchewan and can easily be observed during blowing snow because of the tone contrast between soil and snow particles. Because of their size, most soil particles eroded during blowing snow are vertically transported and not locally deposited. If a contaminant is included within the blowing snow particle, sublimation of ice will increase the relative concentration of the contaminant. Thus blowing snow may have an important role in the chemical transport to and from snowpacks.

7. CONCLUSIONS

A calculation procedure for blowing snow mass flux in the horizontal and vertical direction and sublimation of blowing snow particles has been derived and demonstrated. The mass flux calculations are extremely sensitive to threshold conditions for transport when windspeeds are low. However, the calculations are resilient in terms of threshold condition variability when windspeed is high. Sublimation rates are found to be extremely sensitive to atmospheric temperature and humidity. A limited test of the blowing snow model using field measurements demonstrates reasonable accuracy in calculating drift densities for high transport rates in the higher drift density zone of flow. However results near the threshold conditions for transport do not match the measurements well. It is suggested that the model shows promise in calculating the hydrologically significant range of blowing snow transport where ambient windpeeds are high and threshold windspeeds relatively low.

Blowing snow vertical transport and sublimation are shown to be responsible for snowpack erosion in areas of fully developed flow. Erosion rates as a result of these processes can be significant in some areas. Blowing snow may act to both concentrate contaminants in the snowpack and enhance the verical transport of small surface particulates.

It is recommended that research continue on the mechanics of blowing snow transport and that research be initiated on the effects of blowing snow on chemical transport. In addition to theoretical research, more measurements of blowing snow drift densities and particle sizes are needed to calibrate models of blowing snow transport.

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