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Prairie and arctic areal snow cover mass balance using a blowing snow model

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Abstract. Algorithms to calculate the threshold wind speed and the effect of exposed vegetation on saltation and to describe vertical profiles of humidity in blowing snow, permit the calculation of point blowing snow transport and sublimation fluxes using standard meteorological and landcover data or simple interfaces with climate models. Blowing snow transport and sublimation fluxes can be upscaled to calculate open environment snow accumulation by accounting for their variability over open snow fields, increase in transport and sublimation with fetch, and the effect of exposed vegetation on partitioning the shear stress available to drive transport. Blowing snow fluxes scaled in this manner were used to calculate snow mass balance and to simulate seasonal snow accumulation at a southern Saskatchewan prairie and an arctic site. Field measurements at these sites indicated that from 48% to 58% of snowfall was removed by blowing snow before melt began. Simulations suggest that the ratios of snow removed and sublimated by blowing snow to that transported were 2:1 and 1:1 at the prairie and arctic sites respectively. The resulting methodology was capable of estimating winter season mass balances for these snowpacks that compared well with snowfall and snow accumulation measurements.

1. Introduction

Snow cover has profound impacts on climate and is an extremely important variable to global water and energy cycling. There are many important interactions between snow cover and the atmosphere; one of which, wind redistribution of snow or blowing snow, is ubiquitous in windswept, exposed environments. Blowing snow transport is normally accompanied by in-transit sublimation [*Dyunin*, 1959; *Schmidt*, 1972; *Pomeroy*, 1989; *Mobbs* and *Dover*, 1993; *Déry and Taylor*, 1996; *Bintanja*, 1998]. Transport and sublimation result in losses to exposed snowcovers from erosion of from 30% to 75% of annual snowfall in prairie, steppe, and tundra environments [*Tabler*, 1975; *Tabler et al.*, 1990; *Benson* and *Sturm*, 1993; *Pomeroy et al.*, 1993, 1997]. The disposition of this eroded snow to either sublimation or transport and subsequent deposition is important to surface hydrology and atmospheric water vapor budgets. Transported snow is available for snowmelt, while that sublimated is returned to the atmosphere. Blowing snow fetch, or the downwind distance of uniform terrain that permits snow transport, determines the disposition between sublimation and transport, longer fetches promoting greater sublimation per unit area [*Tabler*, 1975; *Pomeroy and Gray*, 1995].

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Calculation of blowing snow fluxes (erosion, transport, sublimation) for a uniform area, using the presumption of horizontal steady state flow [*Pomeroy*, 1989; *Déry and Taylor*, 1996], does not provide sufficient information to calculate the snow cover mass balance over larger areas where flow at many points in the landscape will deviate significantly from steady state conditions. Schemes have been developed to calculate blowing snow fluxes over terrain of varying fetch and land use [*Pomeroy et al.*, 1993, *Déry et al.*, 1998] and varying terrain [*Pomeroy et al.*, 1997; Liston *and Sturm*, 1998; *Essery et al.*, 1999]. For example, the computationally intensive schemes presented by *Pomeroy et al.* [1993], *Liston and Sturm* [1998], and *Essery et al.* [1999] calculate fluxes using physically-based algorithms for a series of downwind control volumes; the downwind boundary condition at one control volume is the upwind boundary condition of the next. A drawback of these schemes is that some of the input parameters (threshold wind speed for transport, occurrence of blowing snow] are not normally available from meteorological records. A simpler scheme [*Pomeroy et al.*, 1997] uses climatological transport and sublimation relationships to approximate monthly snow fluxes over major landscape types with provision for snow redistribution between landscapes. However, the empirical basis and monthly time step make this scheme unsuitable for interaction with many atmospheric and hydrological models. *Li and Pomeroy* [1997a] present a method to directly calculate threshold conditions for snow transport from the meteorological history of snowpacks, while *Li and Pomeroy* [1997b] calculate the probability of blowing snow occurrence using similar data. Adaptation of these methods, permits the application of physically based blowing snow algorithms driven by standard meteorological data sets and provides a method for scaling blowing snow fluxes from a point to a larger uniform area. It is the purpose of this paper to outline and demonstrate practical, computationally efficient techniques to apply physically based blowing snow algorithms to points and areas in prairie and arctic environments. The techniques employ upscaling, transport threshold wind speed, humidity gradient, and snowfall correction parameterizations that are derived from observations, while other techniques are developed from theory.

2. Model Development, Data Requirements, and Structure

The physics of snow transport and sublimation involve phase change, two-phase flow and rapid energy and mass transfers in the atmospheric boundary layer just above the

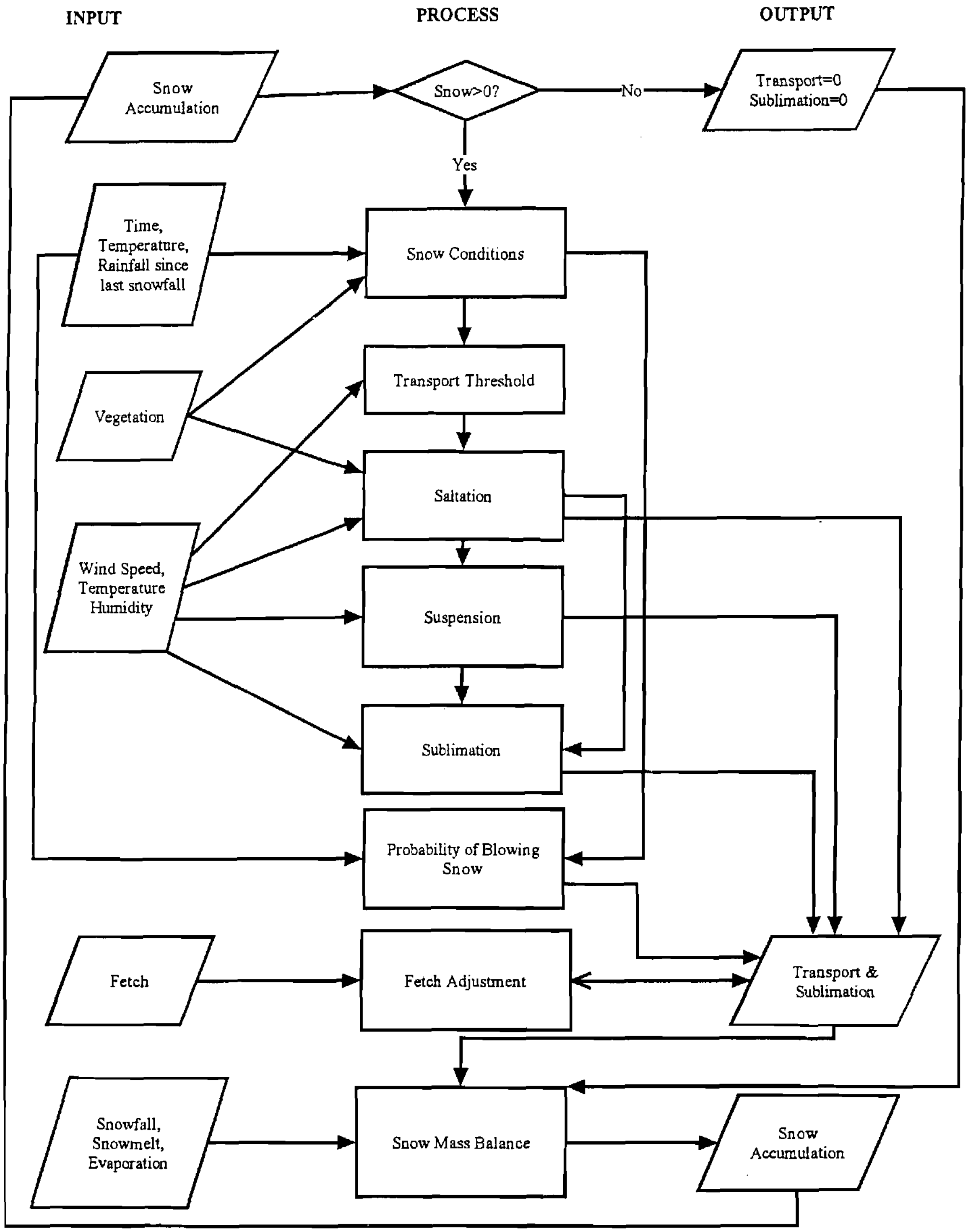


Figure 1. Blowing snow model structure, input data, and output.

snowpack. The prairie blowing snow model (PBSM) was first developed in 1987 as a single column mass and energy balance that calculates blowing snow transport and sublimation rates [Pomeroy, 1988, 1989] and later extended to include a snow cover mass balance for the case of two dimensions [Pomeroy et al., 1993]. The model presented here uses a modified, single column calculation with new methods to calculate the inputs to this scheme and to scale the fluxes from a point to a landscape in a snow mass balance calculation. The model structure, input data and output are shown in Figure 1 and described in the following sections.

2.1. Snow Mass Balance

The snow mass balance on a uniform element of a landscape is the result of the distribution and divergence of blowing snow fluxes surrounding the element and within the element [*Pomeroy et al.*, 1997]. The following, upscaled, mass balance can be drawn over an element of a landscape having fetch distance, x (m),

$$\frac{dS}{dt}(x) = P - p\left[\nabla F(x) + \frac{\int E_B(x)dx}{x}\right] - E - M , \qquad (1)$$

where dS/dt is surface snow accumulation (kg m^{-2} s^{-1}), P is snowfall (kg m^{-2} s^{-1}), p is the probability of blowing snow occurrence within the landscape element, F is the downwind transport rate (kg m^{-1} s^{-1}), E is snow surface sublimation (kg m^{-2} s^{-1}), E_B is blowing snow sublimation (kg m^{-2} s^{-1}), and M is snow melt (kg m^{-2} s^{-1}). To aggregate the fluxes from each landscape element up to larger-scale values, fluxes are normally weighted by the respective areas of the landscape elements [*Pomeroy et al.*, 1997; *Liston and Sturm*, 1998; *Essery et al.*, 1999]. A cross-sectional view of the control volume of equation (1) is shown in Figure 2. Application of the blowing snow algorithms to solve for the snow mass balance requires calculating each term of equation (1). M and E will not be greatly discussed in this paper but require the blowing snow model to be coupled to models of melt and snow surface sublimation. In general, $E << E_B$ [*Schmidt*, 1982; *Pomeroy and Essery*, 1999]. *Granger and Male* [1978] found E was strongly influenced by net radiation and had a melt season (April) mean of 0.13 kg m^{-2} d^{-1} in a prairie environment. *Pomeroy et al.* [1998] suggest that even smaller values prevail in March in this region.

2.2. Snowfall P

Snowfall undermeasurement due to wind effects, wetting losses and unrecorded trace events has been the subject of extensive investigation [*Goodison*, 1978, 1982; *Sveruk*, 1992]. In Canada, Nipher-shielded cylinders are most often used to collect snowfall, with measurements of accumulated snowfall water equivalent in the cylinder made every 6 hours. The Nipher shield reduces undermeasurement due to wind compared to that of an unshielded gauge, but undermeasurements still occur. A correction procedure recommended by *Goodison et al.* [1998] is used for uncorrected precipitation records. The upward revision of annual snowfall is of the order of 31% in the southern Canadian prairies and 64% to 161% in the high arctic [*Pomeroy and Goodison*, 1997]. The procedure involves first dividing measured P by a dimensionless Nipher catch ratio (R), where

$$R = 0.01\,[100 - 0.387\,u(z)^2 - 2.022\,u(z)] \qquad (2)$$

and u is wind speed at a height z (10 m for the empirical coefficients shown). Equation (2) is based on observations in $u(10) < 8$ m s^{-1}; for wind speeds greater than this, a constant gauge undercatch is assumed. Figure 3 shows the Nipher gauge undercatch function for wind speed as implemented for

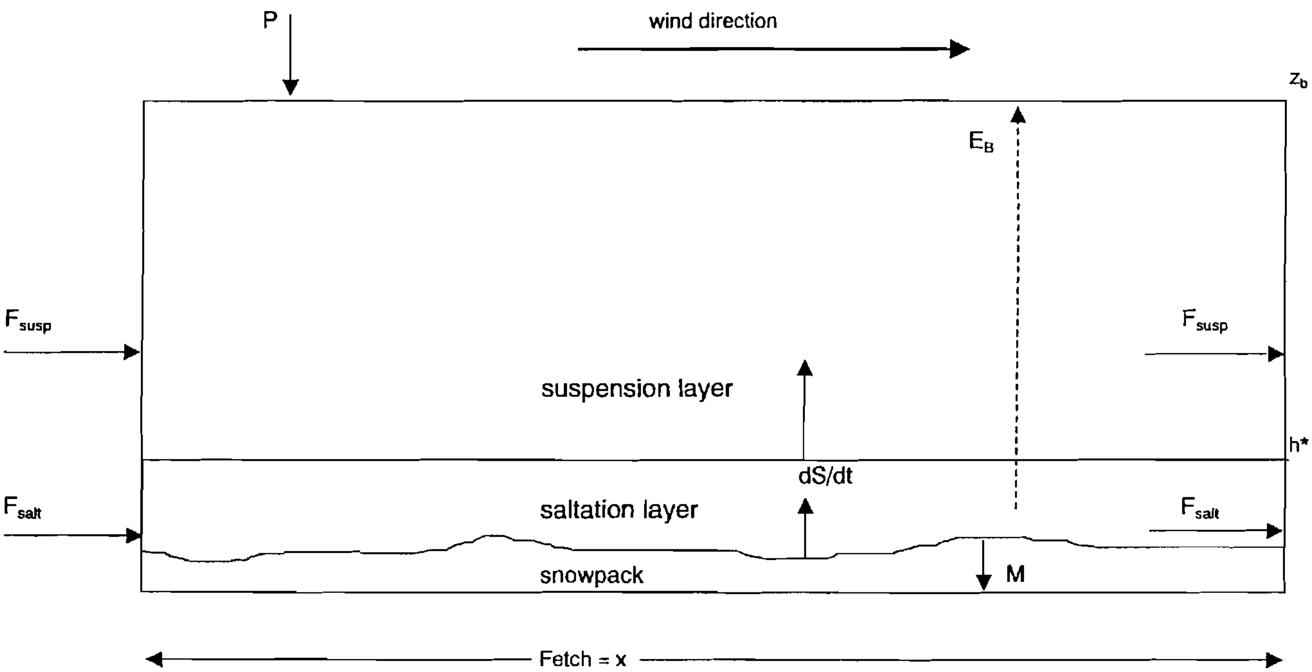


Figure 2. Model fluxes, boundaries and control volume concept. Schematic of mass fluxes considered, with reference to a control volume for the snow mass balance over a uniform fetch.

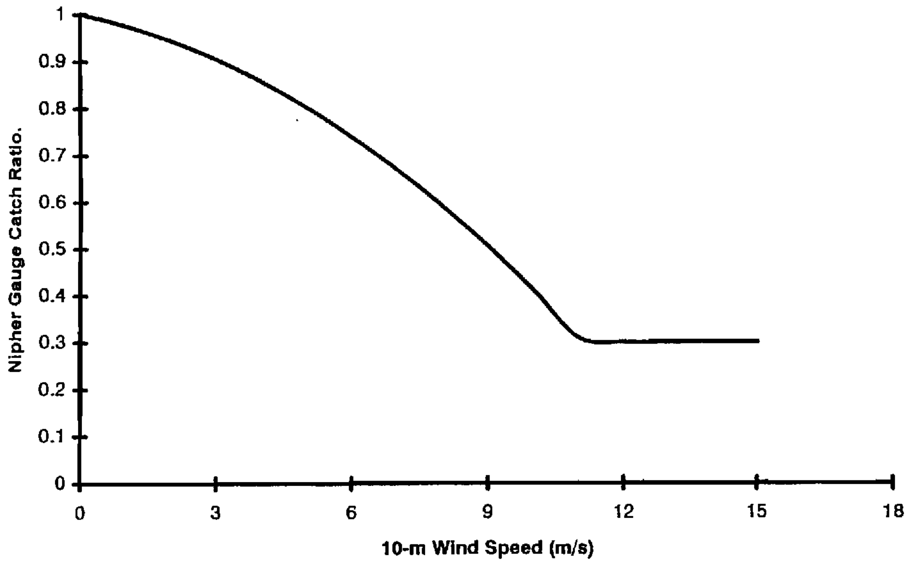


Figure 3. Nipher gauge undercatch function versus 10-m wind speed as implemented in the model.

the blowing snow model. A systematic gauge wetting loss of 0.15 mm snow-water equivalent (SWE) per day is added if snowfall occurs and a trace measurement loss of 0.15 mm SWE is added for each 6-hour period where trace snowfall is recorded but no amount given [*Goodison et al.*, 1998].

2.3. Fully Developed Snow Transport F

For brevity, only the essential equations describing transport fluxes are listed here along with key aspects of their implementation, a full listing is given by *Pomeroy et al.* [1993]. Snow erosion is initiated by saltating snow, from which an upward flux and downwind flow of snow particles develops. Downwind transport fluxes are composed of the sum of snow blowing in saltation and suspension layers of the atmosphere and are calculated initially for a point where flow is "fully developed" within the specified boundary layer height. As shown in equation (1), the divergence in transport flux along a fetch is an important component of snow accumulation.

The blowing snow saltation layer is very close to the surface and must become established before other blowing snow processes (suspension, sublimation) can proceed. Though the transport rate of saltating snow is relatively low, this process controls the erosion of snow and is highly sensitive to snowpack physical structure, and depth and to exposed vegetation. F_{salt} (kg m^{-1} s^{-1}) is calculated using a shear-stress partitioning method first derived by *Bagnold* [1941], applied to blowing snow by *Schmidt* [1986] and developed for the snow saltation layer with vegetation by *Pomeroy and Gray* [1990],

$$F_{salt} = \frac{c_1 e \rho u^*_t}{g}\left(u^{*2} - u^{*2}_n - u^{*2}_t\right) \quad . \qquad (3)$$

Pomeroy and Gray [1990] found that c_1 is the ratio of saltation velocity, u_p, to friction velocity ($u_p/u^* = 2.8$), e is the efficiency of saltation, $1/(4.2\,u^*)$, ρ is atmospheric density (kg m^{-3}), g is acceleration due to gravity (m s^{-2}), u^* is the atmospheric friction velocity (m s^{-1}), and the subscripts n and t refer to that portion of u^* applied via the shear stress ($\rho\, u^{*2}$) to the nonerodible roughness elements (nonerodible friction velocity) and to open snow surfaces at the transport threshold (threshold friction velocity), respectively. The friction velocity is calculated from wind speed, presuming a logarithmic wind speed profile. A form of equation (3) is used in many blowing snow models [*Déry and Taylor*, 1996; *Bintanja*, 1998; *Liston and Sturm*, 1998].

During blowing snow, the aerodynamic roughness height z_0 is not constant but is controlled by the height of saltating particles. Following *Owen* [1964], the saltation height is controlled by the vertical velocity of initial particle trajectories leaving the snow surface and therefore proportional to $u^{*2}/(2g)$. *Pomeroy and Gray* [1990] found a linear relationship between calculated saltation height and measured z_0 for snow covers without vegetation; this relationship is consistent with earlier empirical findings [*Tabler*, 1980]. For snow with vegetation the model uses a formulation from *Pomeroy and Gray* [1995], where

$$z_0 = \frac{c_2\, c_3\, u^{*2}}{2g} + c_4\, \lambda \qquad (4)$$

and c_2 is the square root of the ratio of initial vertical saltating particle velocity to the friction velocity, proposed by P. R. Owen (unpublished manuscript and presentation, 1980, referenced by *Greeley and Iversen*, [1985]) to be 1.6, c_3 is the ratio of z_0 to saltation height, found by *Pomeroy and Gray* [1990] to be 0.07519, and c_4 is a drag coefficient found by *Lettau* [1969] to be 0.5. The parameter λ is the dimensionless roughness density [*Lettau*, 1969] and is a function of vegetation number density N (number m^{-2}), the vegetation stalk diameter d_v (m), the height of vegetation h_v (m), and the snow depth (m), where

$$\lambda = N\, d_v \left(h_v - \frac{S}{\rho_s}\right) \qquad (5)$$

and snow depth is snow accumulation S divided by ρ_s, snow density (kg m^{-3}). Snow density has a very small covariance with depth for depths < 0.6 m on the Canadian prairies and can be found as [*Shook and Gray*, 1994]:

$$\rho_S = \overline{\rho_S} + \frac{B}{d_S} \tag{6}$$

where the mean density is 239 kg m^{-3}, B is 2.05 kg m^{-2}, and d_S is snow depth (m). The last term in equation (6) is the covariance between depth and density. Values for B were determined from a large archive of snow surveys [*Shook and Gray*, 1994]. For depths > 0.6 m, covariance between depth and density is significant and a combination of Shook and Gray's relationship with that of *Tabler et al.* [1990] provides a mean density of 456 kg m^{-3} and B of -128 kg m^{-2} [*Pomeroy and Gray*, 1995].

Li and Pomeroy [1997a] examined transport threshold wind speeds for blowing snow at low vegetation sites (mown grass, pavement) on the Canadian prairies. Presuming a "pre transport" z_0 of 0.2 mm, their results indicate an average u^*_t for wet snow of 0.37 m s^{-1} with a range from 0.26 to 0.52 m s^{-1} and a range for dry snow from 0.15 to 0.4 m s^{-1}. Wet snow was defined as that which had received above freezing temperatures or rainfall since the last snowfall. The dry snow threshold increased with increasing air temperature most rapidly near the melting point. The relationship developed by Li and Pomeroy, expressed to solve for threshold friction velocities assuming z_0 = 0.2 mm, is

$$u^*_t = 0.35 + \frac{T}{150} + \frac{T^2}{8200} \quad , \tag{7}$$

where T is air temperature (°C) measured at 2-m height. When the calculated snow depth is < 1 cm, the threshold is considered higher than the wind speed.

Raupach et al. [1993] developed an algorithm that relates the geometry of exposed vegetation to partitioning of shear stress at the surface, where u^*_n is found as

$$u^*_n = u^*(\beta\lambda)^{0.5}(1+\beta\lambda)^{-0.5} \quad , \tag{8}$$

where β is the dimensionless ratio of element to surface drag. They found that β was ~170 from studying wind erosion of soil; presuming that shear stress partitioning is valid for both soil and snow, this value was used. The effect of air temperature on the threshold friction velocity and of snow accumulation on the nonerodible friction velocity for a given vegetation cover is shown in Figure 4. Higher air temperatures and/or lower snow accumulations result in the requirement for stronger winds for u^* to overcome the combined effects of u^*_t and u^*_n in resisting saltation transport (Figure 4a and 4b). From equation (3), u^{*2} must exceed u^{*2}_t + u^{*2}_n for saltation to occur.

The suspended layer of blowing snow extends from a reference height h^* near the top of the saltation layer to the top of the blowing snow boundary layer. Presuming that suspended snow is carried downwind at the same velocity as the wind [*Schmidt*, 1982], and that suspended snow diffuses from near the top of the saltation layer, the mass flux, F_{susp} (kg m^{-1} s^{-1}), may be written as [*Pomeroy and Male*, 1992],

$$F_{susp} = \frac{u^*}{k}\int_{h^*}^{z_b} \eta(z)\ln\left(\frac{z}{z_0}\right)dz \ , \tag{9}$$

where k is von Kármán's constant (0.4), η is the mass concentration of suspended blowing snow, and z_b is the upper boundary limit (5 m for fully developed flow). Presuming a balance between upward diffusion of snow by turbulence, diffusion from the saltation layer, consumption by sublimation, and settling of suspended particles, the steady state η at height ($z + dz$) may be found from that at z as

$$\eta(z+dz) = \eta(z)\left(\frac{z+dz}{z}\right)^{\frac{w'(z)z}{\kappa_s(z)}} \quad , \tag{10}$$

where w' is an effective vertical snow particle velocity at height z, and κ_S is the eddy diffusivity for blowing snow particles at height z.

Sublimation, entrainment from saltation, particle inertial effects, and settling confound the direct calculation of w' and κ_S [*Pomeroy*, 1988; *Déry et al.*, 1998]. Values for these coefficients have not been measured but are subject to a wide range of suggestions [*Schmidt*, 1982; *Pomeroy and Male*, 1992; *Déry et al.*, 1998]. *Pomeroy and Male* [1992] measured vertical profiles of suspended snow mass concentration with and without concurrent snowfall. By defining a dimensionless vertical diffusion velocity w^*, equation (10) was used with these measurements to find the variation in w* with height:

$$w^* = \frac{w'(z)z}{\kappa_S(z)} = \frac{\dfrac{F_V(z)}{\eta(z)} - \omega(z)}{\kappa_S(z)} = c_5\, z^{c_6} \ , \tag{11}$$

where F_v is the vertical mass flux at height z (equivalent to the surface erosion flux less any sublimation from the surface to height z), ω is the terminal fall velocity (of particle of mean mass), and the coefficient c_5 and exponent c_6 are -0.8412 and -0.544, respectively (r^2 = 0.82, n = 324).

Equation (9) requires blowing snow concentrations at height h^*, the lower boundary condition, as calculated by *Pomeroy and Gray* [1990] and *Pomeroy and Male* [1992]. Suspended snow diffuses from a reference concentration set by saltation dynamics, so suspended fluxes are also influenced by the threshold and nonerodible friction velocities. Suspended mass fluxes are ~5-10 times higher than those in the saltation layer, the difference increasing with increasing friction velocity [*Pomeroy et al.*, 1993].

2.4. Fully Developed Blowing Snow Sublimation E_B

The sublimation calculation for a column of blowing snow over a unit area of land is based on a vertical integration of the sublimation rate of a single ice particle having mean mass m at height z, the mean particle mass being determined using a two-parameter gamma distribution of particle size and the change in this distribution with height [*Pomeroy et al.*, 1993]:

$$E_B = \int_0^{z_b} \frac{1}{m(z)}\frac{dm}{dt}(z)\,\eta(z)\,dz \quad . \tag{12}$$

Sublimation in equation (12) is calculated for both saltation and suspension layers. *Schmidt* [1972] has shown that the sublimation rate of a single blowing snow particle can be calculated as a balance of radiative energy exchange, convective heat transfer to the particle and water vapor from the particle, and latent heat exchange during phase change. *Schmidt* [1972, 1991], *Pomeroy* [1989], *Pomeroy et al.* [1993], *Déry and Taylor* [1996], and *Liston and Sturm* [1998] describe the full sublimation calculation scheme in detail. The key constituents of the particle energy and mass balance are the convective exchanges, which are balanced using turbulent transfer parameters (dimensionless Sherwood, Sh, and Nusselt, Nu, numbers) calculated from the particle Reynolds

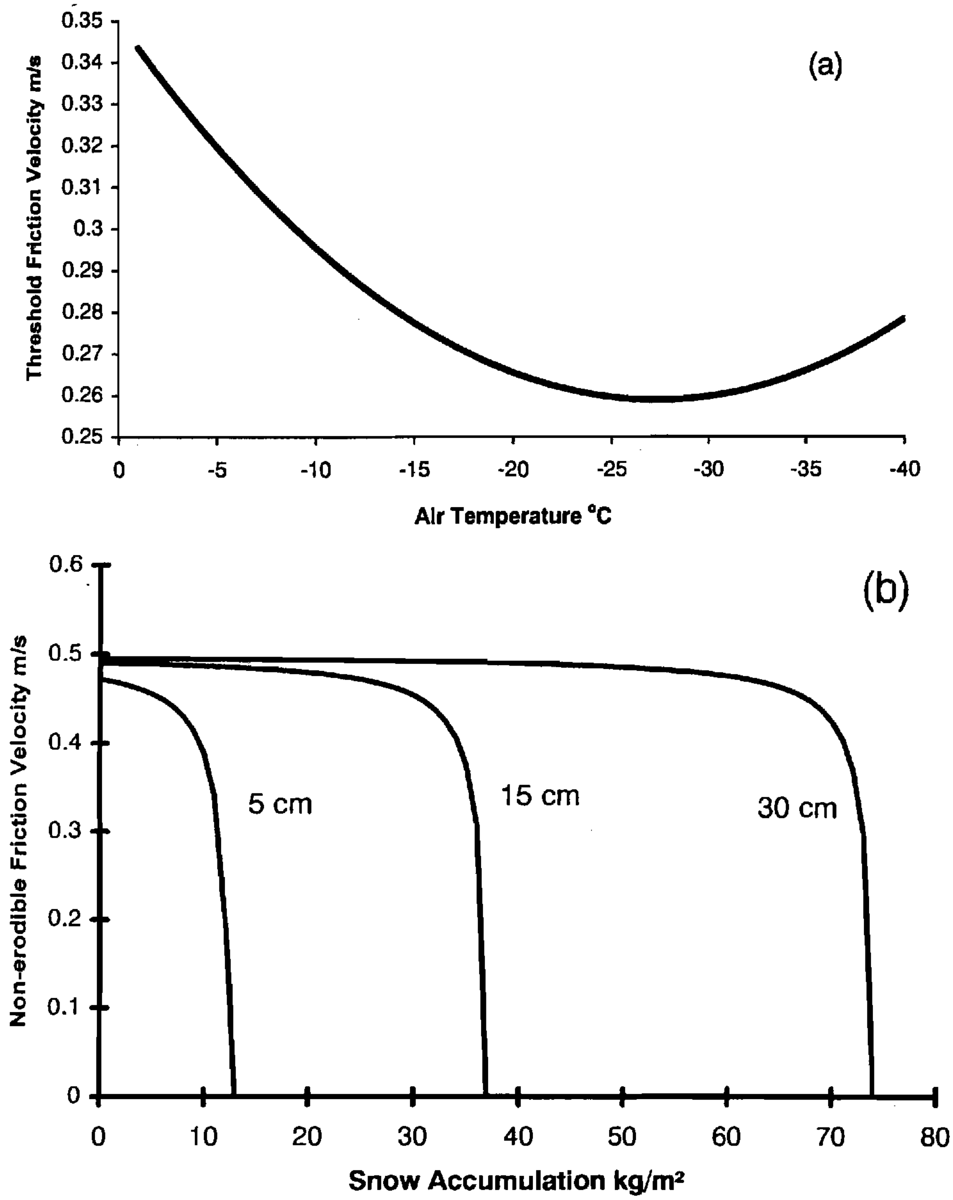


Figure 4. Saltation parameters. (a) Threshold friction velocity and air temperature relationship and (b) nonerodible friction velocity and snow water equivalent as modeled presuming cereal stubble vegetation with $u^* = 0.5$ m s^{-1}, $\beta = 170$, $Nd_v = 0.96$ m^{-1}, and $h_v = 0.05$ m, 0.15 m, and 0.3 m.

number and the assumption of thermodynamic equilibrium for the particle as

$$\frac{dm}{dt}(z) = 2\pi r(z) D\, Sh(z) \rho_a \{q(z) - q_s[T(z)]\}$$
$$= 2\pi r(z) \frac{\lambda_t}{h_s} Nu(z)[T_s(z) - T(z)] \quad , \quad (13)$$

where $r(z)$ is the radius of the particle of mean mass at height z, D is diffusivity of water vapor in air (m^2 s^{-1}), q is specific humidity, λ_t is thermal conductivity of air (J m^{-1} s^{-1} K^{-1}), and h_s is latent heat of sublimation (J kg^{-1}). The subscript s refers to the snow particle surface, where air is presumed to be saturated and at the ice bulb temperature. Equations (12) and (13) are driven using vertical gradients of air temperature, water vapor deficit, wind speed, blowing snow mass concentration, and particle radius. Important differences in the implementation can derive from the assumed vertical ... [*Déry and Taylor*, 1996].

Equation (13) shows that values of ambient temperature and humidity are critical to modeling blowing snow sublimation. There is presently a divergence of thought in the field regarding appropriate methods to set these parameters. Several recent models calculate changes to atmospheric energy and water vapor budgets that are due solely to blowing snow sublimation [*Mobbs and Dover*, 1993; *Déry et al.*, 1998]. The resulting negative feedback to sublimation (decreased water vapor deficit and temperature) that these models predict results in a substantive reduction in sublimation rates over time and with increasing fetch from initial conditions. For instance, for $u(10) = 15$ m s^{-1}, *Déry et al.* [1998] predicted an increase in relative humidity (RH) with respect to ice at $z = 1$ m of from 70% (nominal value) to 93%, and a corresponding decline in T (air temperature) at $z = 1$ m from -10 (nominal value) to -10.4°C as fetch increased from 0 to 2 km. The predicted vertical gradients for humidity were also strong, with an RH of 91% at $z = 1$ m increasing to

98% at z = 0.1 m for a 1-km fetch, a substantial increase from the uniform profile set as an initial condition.

Observations from both continental and maritime mountain, prairie and arctic sites in North America and Europe suggest that the increase in humidity and decrease in temperature associated with blowing snow do not normally cause near saturation in the lower boundary layer and that in many cases the temperature and undersaturation increase due to enhanced mixing and advection of dry air during blowing snow storms as initially stable boundary layers are mixed [*Schmidt*, 1982; *Pomeroy*, 1988, 1991; *Essery et al.*, 1999]. For illustration, two examples are drawn from the measurements of *Pomeroy* [1988] in a large open snow-covered field west of Saskatoon, Saskatchewan (52°N, 107°W), with a fetch (for these measurements) in excess of 2 km (Figure 5). As shown in Figure 5, midday on January 26 1987, a sharp increase in $u(3)$ from 2-3 to 7-11 m s^{-1} initiated blowing snow. $T(2)$ rose from -8°C to -2°C and a stable temperature gradient of about 0.35 °C m^{-1} developed. Humidity was initially well-mixed but developed a strong gradient, indicative of sublimation, of -3% RH m^{-1}. RH(2) (with respect to ice) rose from 55% to 63% over the blowing snow event but that near the surface, RH(0.08), remained fairly steady at 65%, well below saturation. Midday on February 27, $u(3)$ increased from near 5 to over 10 m s^{-1}, initiating blowing snow (Figure 5). Air temperatures rose from -18°C to -7°C by mid-event and then slowly declined to -10°C by late evening. During the rise in temperature the T gradient was insignificant; later in the event a gradient of 0.35 °C m^{-1} developed. RH rose from an initial 2 m value of 65% to a sharp, unexplained peak of 90% and then rapidly fell to 63%, after which an RH gradient indicative of an upward flux of water vapor developed (-2% RH m^{-1}) but faded later in the event as RH(2) peaked at 70%. By contrast, for $u(10)$ = 15 m s^{-1} and nominal T and RH in the mid-range of these measurements and fetches from 0.1 to 10 km, *Déry et al.*, [1998] predicted the development of stable temperature gradients of from 0.0005 to 0.05 °C m^{-1} and RH gradients of from -5.5 to -2% RH m^{-1} for the same height range. These field examples demonstrate the operation of a phenomenon much more complex than that which can be described alone by the sublimation negative feedback hypothesis of *Mobbs and Dover* [1993] and *Déry et al.* [1998] and suggest that thermodynamic blowing snow models may need to incorporate a more comprehensive treatment of exchange processes to produce realistic development of temperature and humidity fields in blowing snow.

Because of the uncertainty in deriving temperature and humidity fields during blowing snow from thermodynamic considerations, the model uses measured values of air temperature and relative humidity from locations with large open fetches (e.g., airports, meteorological stations on open plains or tundra) where approximate steady state conditions for blowing snow sublimation have already developed. To adjust the measured temperature and humidity for heights other than that of measurement (2 m), profiles of air temperature and water vapor deficit measured during an extensive field program at Loreburn, Saskatchewan (51°N, 106°W), were used [*Pomeroy and Male*, 1987]. Loreburn is located in an open grain-growing region with a fetch for

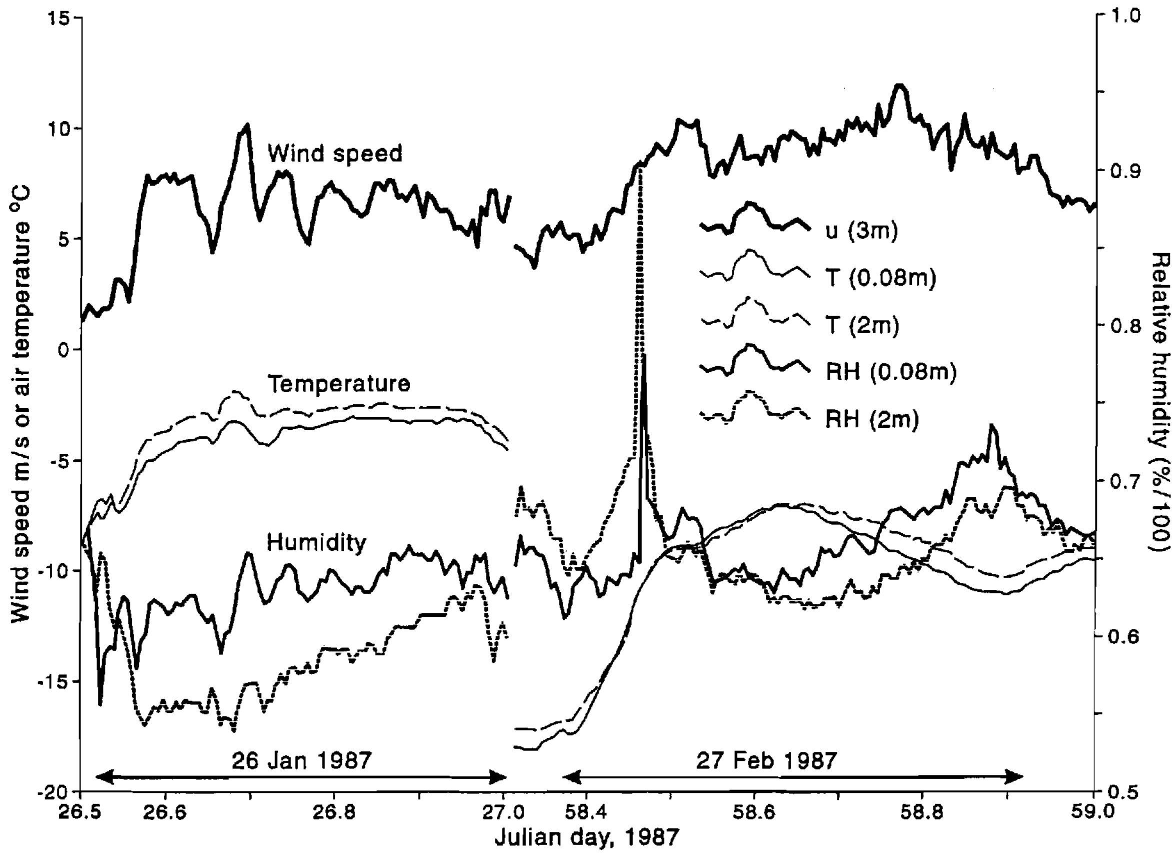


Figure 5. Evolution of wind speed, temperature and relative humidity fields during blowing snow events near Saskatoon, Saskatchewan, January 26 and February 27, 1987.

blowing snow of at least 3 to 4 km, depending on wind direction and snow depth. Over two seasons at Loreburn, air temperatures measured from 0.1 to 3 m height during strong blowing snow events ($u(10) > 10$ m s^{-1}) showed no consistent gradient, with both stable and unstable profiles developing in response to air mass characteristics, hence $T(z) = T(2)$ in the model. Water vapor deficit during strong blowing snow did show a consistent vertical profile at Loreburn, with undersaturation, σ becoming smaller as height above the snow surface decreases as demonstrated in Figure 6. The best fit profile is

$$\sigma(z) = \sigma(2)[1.019 + 0.027\ln(z)] \quad . \tag{14}$$

Equation (14) is also plotted in Figure 6 along with a set of measurements of humidity with respect to ice (7.5-min averages) made with calibrated, shielded and ventilated

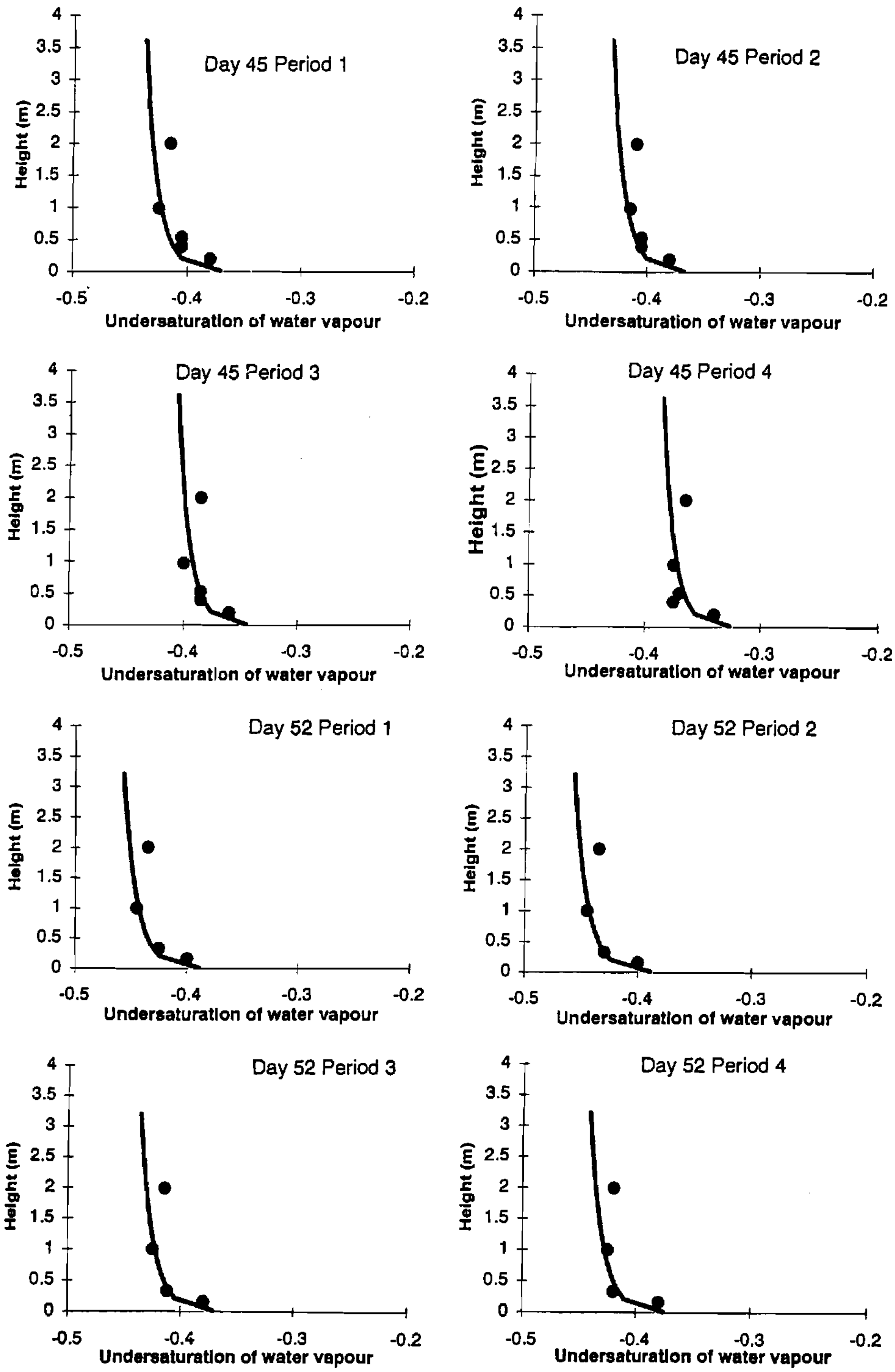


Figure 6. Sequential vertical 7.5-min profiles of undersaturation of water vapor measured at Loreburn, Saskatchewan, during strong blowing snow events: measurements (points) and modeled profiles (lines), February 14 (day 45) and 21 (day 52), 1986.

Honeywell lithium-chloride dewcells during blowing snow at Loreburn [*Pomeroy*, 1988].

The blowing snow model with this parameterization of sublimation and water vapor and temperature gradients and measured 2-m values of T and RH was evaluated using field measurements of latent heat flux from an eddy correlation device east of Saskatoon during two midwinter blowing snowstorms. The model estimate of sublimation matched measurements well, with a mean error of -0.83 W m^{-2} and a peak latent heat flux of 60 W m^{-2} [*Pomeroy et al.*, 1999].

2.5. Nonsteady Fluxes: Fetch Effects

The transport and sublimation fluxes shown in section 2.4 presume steady-state flow and a fully developed boundary layer up to a 5-m height. These conditions are normally met for fetches of several kilometers. To calculate the height of the upper boundary for suspension for fetches shorter than that required for full development, *Pomeroy* [1988] used a formulation from *Pasquill* [1974] where

$$z_b(t) - z_b(0) = k\, u^*\, t \tag{15}$$

and t is elapsed time. Time can be expressed in terms of distance x, using

$$t(x) = x\{u[z_b(t(0))]\, u[z_b(t(x))]\}^{-0.5} \quad . \tag{16}$$

Substituting the logarithmic wind speed profile equation into equations (15) and (16) provides a solution for z_b as a function of fetch x,

$$z_b(t) = z_b(0) + k^2[x(t) - x(0)]\left[\ln\left(\frac{z_b(0)}{z_0}\right)\ln\left(\frac{z_b(t)}{z_0}\right)\right]^{-0.5} . \tag{17}$$

The measurements of increase in transport rate with fetch by *Takeuchi* [1980] suggest that appropriate values for reference values $z_b(0)$ and $x(0)$ are 0.3 and 300 m, respectively. The use of a reference well above the saltation layer is to permit development of steady-state saltation and to find a height where Pasquill's assumption of equal particle and atmospheric velocity variance is valid. Equation (1) is used to integrate the fluxes to produce averaged values over fetch x appropriate for an areal snow mass balance.

2.6. Scaling Fluxes From Point to Area

Field observations of blowing snow show that it is extremely unsteady over space and time. Over a continuous snow cover, transitory patches of snow transport and non-transport may be observed during even strong blowing snow storms, and time series of blowing snow fluxes at a point show considerable variation and intermittency [*Schmidt*, 1986; *Pomeroy*, 1988]. Observations of intermittent and spatially variable snow transport are likely associated with small-scale variation in snow cover properties and boundary-layer flow that are not addressed by the uniform fetch and constant wind speed assumptions of most blowing snow models. *Shook and Gray* [1994, 1996] have shown that snow depth and extent demonstrate "fractal" characteristics and are self-similar over scales of < 100 m. Other snow properties important to blowing snow also scale with depth. The fractal nature of snow cover suggests that Gaussian distributions may be able to describe the spatial variation in blowing snow fluxes.

The probability distribution of blowing snow occurrence provides a potential scaling function for upscaling fluxes from point to area. Occurrence probability is developed from observations over time but can be applied to space if the area of application is uniform in its mean characteristics, a requirement met by many landscapes found in the prairie and arctic environments. Areal fluxes are found as in equation (1), where p is the probability of occurrence of blowing snow.

Li and Pomeroy [1997b] concluded that the probability of occurrence of blowing snow with respect to wind speed approximates a cumulative normal distribution, described by the mean u_{mean} (location parameter), and the standard deviation δ (scale parameter) of wind speed, u, as,

$$p = \frac{1}{\delta\sqrt{2\pi}} \int_0^u e^{-\frac{(u_{mean} - u)^2}{2\delta^2}}\, du \quad . \tag{18}$$

They then developed empirical descriptions for the location and scale parameters as functions of wet snow conditions, air temperature, and snow age, from a six-year data set collected at 15 meteorological stations in the Canadian prairies. For wet or icy snowpacks, the mean and standard deviation of wind speed were found to be 21 and 7 m s^{-1}, respectively. For dry snow the mean wind speed was described as

$$u_{mean} = 0.365T + 0.00706T^2 + 0.9I + 11.2\, , \tag{19}$$

where I is a snow age index equal to the natural logarithm of the hours since the most recent snowfall. Similarly, the standard deviation of wind speed was described as

$$\delta = 0.145T + 0.00196T^2 + 4.3 \quad . \tag{20}$$

The resulting blowing snow occurrence probability as a function of wind speed for various temperatures is shown in Figure 7a for $I = 2$. Our observations suggest that values of u < 7 m s^{-1} for wet snow and 3 m s^{-1} for dry snow or conditions with no snow cover result in $p = 0$. If the current wind speed has $p > 0$ and $u < u_t$, where u_t is the threshold wind speed, then u_t is recalculated as equal to u - 0.5 and used to find u_t^* for the model.

Where snow covers are continuous and unvegetated, u is the current $u(10)$. Where vegetation (grass, shrubs) extends above the snow, u is found by partitioning the shear stress applied to the surface between that applied to vegetation and that applied to snow, where

$$u = \frac{\sqrt{u^{*2} - u^{*2}_n}}{k} \ln\left(\frac{z}{z_0}\right) = \frac{u(z)}{\sqrt{1 + \beta\lambda}} \quad . \tag{21}$$

An example of the effect of exposed vegetation on the probability of blowing snow is shown in Figure 7b for an air temperature of -15°C, I = 2, and exposed wheat stubble (λ= 0.96 * h_v) with h_v increasing from 0 to 0.03 m. It is apparent that exposed vegetation has a very strong effect in reducing the probability of blowing snow occurrence. For instance, at a 15 m s^{-1} wind speed the probability of blowing snow declines from 99% to 30% as the exposed stubble height increases from 0 to 0.03 m. This modeled result is in concurrence with observations that wheat stubble tends to fill with wind-blown snow to approximately the height of stubble stalks [*Nicholaichuk et al.*, 1986].

This prairie-derived probability algorithm was used to drive a complex terrain blowing snow algorithm in the arctic,

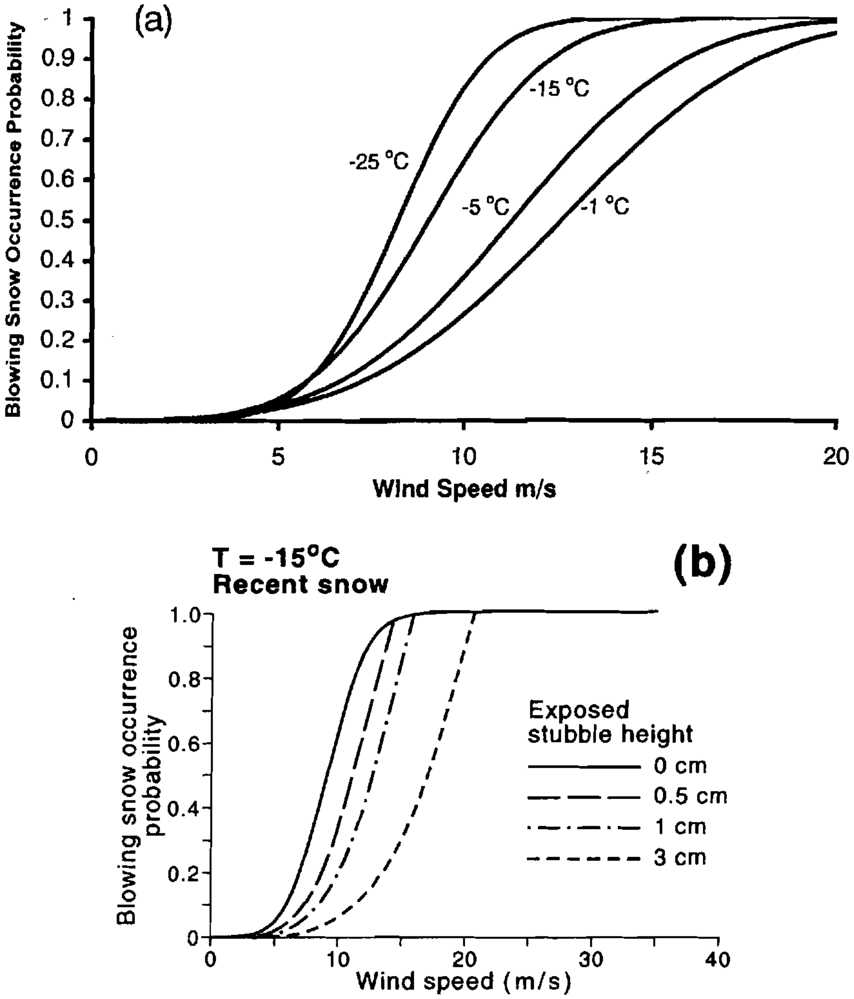


Figure 7. Blowing snow occurrence probability. (a) Modeled probability as a function of wind speed and air temperature for a snow cover 7 hours old [after *Li and Pomeroy*, 1997b]. (b) Modeled probability over snow with exposed cereal stubble stalks as a function of stubble exposure for T = -15°C, I = 2, β = 170, Nd_v = 0.96 m^{-1}, and h_v = 0, 0.005 m, 0.01 m, and 0.03 m.

using measured vegetation cover inputs [*Essery et al.*, 1999]. The distributions of end-of-season snow accumulation matched the pattern of those measured by snow survey with mean errors in estimating snow accumulation of 28% for open tundra, 2% for shrub tundra, and 10% for forests. This suggests that with caution, the probability-scaling algorithm can be applied outside of its area of development.

3. Model Sensitivity to Input Parameters

Model and algorithm sensitivity to many of the coefficients described here has been examined in previous papers [*Pomeroy and Gray*, 1990; *Pomeroy and Male*, 1992; *Pomeroy et al.*, 1993; *Déry and Taylor*, 1996; *Liston and Sturm*, 1998; *Essery et al.*, 1999]. This analysis will therefore focus on sensitivity to inputs. Input parameters may be divided into those that vary over short timescales (wind speed, temperature, humidity, and snow age) and those that are specified for a location (fetch, vegetation density, height, and stalk diameter).

Transport fluxes vary with wind speed, air temperature, and snow age over short timescales. The sensitivity of the 1000-m fetch areal transport flux, $p[F_{salt} + F_{susp}](1000)$ to 10-m wind speed for various air temperatures and snow ages is shown in Figure 8a. Wind speeds for which a transport flux of 0.001 kg m^{-1} s^{-1} is exceeded vary from 6 to 8 m s^{-1}. The transport flux increases with the fourth power of wind speed, increasing approximately an order of magnitude as wind speed increases in increments from 7 to 10, 10 to 15, and 15 to 25 m s^{-1}. The effect of temperature and snow age is only notable for low transport rates ($u(10)$ < 15 m s^{-1}) where warm snow and old snow produce the lowest transport rates. At high wind speeds the effect of temperature and snow age on transport rate is negligible. The surface snow erosion rate required to transport snow over a 1000 m fetch is 0.24 kg m^{-2} h^{-1} for a wind speed of 15 m s^{-1}. Transport rates given here are ~20% lower than that modeled by *Tabler et al.* [1990] and *Pomeroy et al.* [1993] but similar to those reported by *Dyunin and Kotlyakov* [1980] for a wind speed of 15 m s^{-1}.

In Figure 8b the 1000-m fetch areal sublimation flux $pF_{subl}(1000)$ is shown as a function of $u(10)$ for various $T(2)$ and RH(2). Sublimation rate increases with the fifth power of wind speed and is also strongly affected by temperature and humidity. Low air temperature and high relative humidity restrict sublimation. As a result the sublimation rate increases by an order of magnitude as $T(2)$ increases from -30°C to -1°C

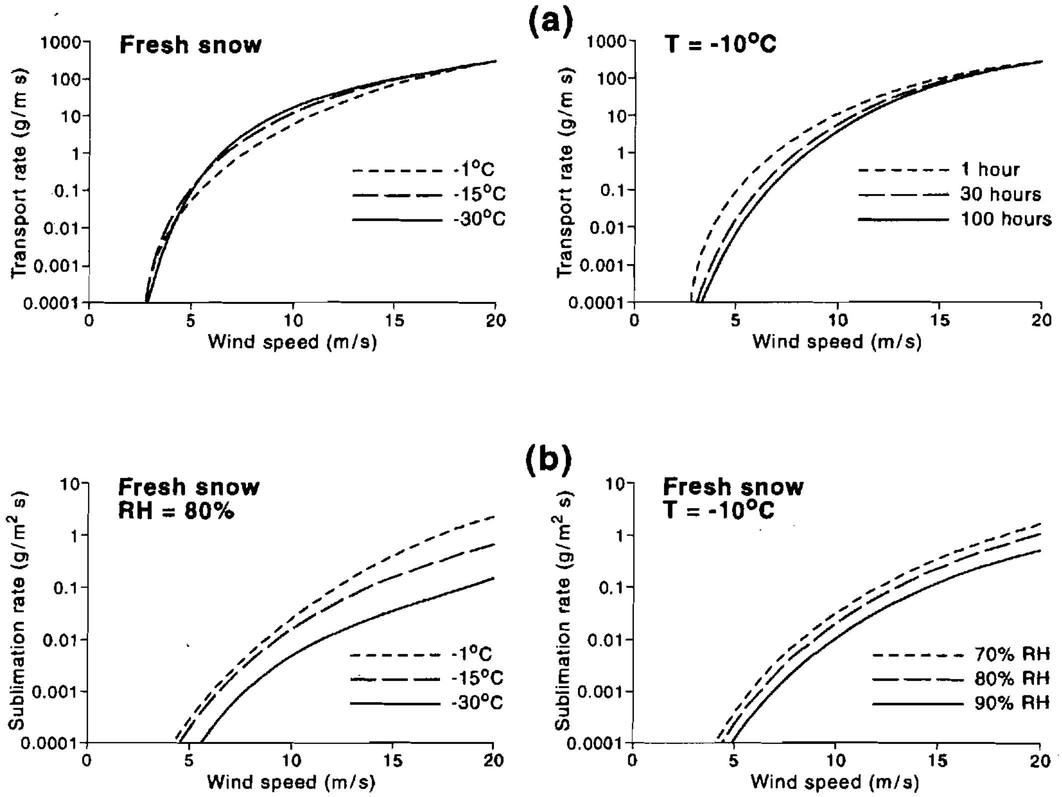


Figure 8. Transport and sublimation fluxes as a function of meteorological parameters. a) Areal downwind transport flux and 10-m wind speed for an unvegetated fetch distance of 1 km and various air temperatures and snow ages. (b) Areal sublimation flux and 10-m wind speed for an unvegetated fetch distance of 1 km and various air temperatures and relative humidities.

and triples as RH(2) drops from 90% to 70%. The surface snow erosion rate necessary to sustain the areal sublimation flux at a fetch of 1000 m is ~0.5 kg m^{-2} h^{-1} for $u(10)$ of 15 m s^{-1}, $T(2)$ of -15°C and RH(2) of 80%.

The strong sensitivity of areal transport and sublimation fluxes to wind speed, air temperature and humidity suggests that care should be taken in selecting data to run the model. For instance, the model is based partly on measurements collected at short time intervals (30 min or less). Therefore application of the model using time intervals > 1 hour [e.g. *Liston and Sturm*, 1998] is not recommended, as errors in averaging wind speeds to determine "mean" fluxes will become large. Systematic errors in measured inputs will cause relatively larger errors in the modeled fluxes; for example, a 1% error in wind speed at 15 m s^{-1} will give a 4% error in transport flux and a 5% error in sublimation flux.

An example of the effect of vegetation, exposed by variable snow accumulation S on areal snow transport and sublimation for fixed variable inputs is shown in Figure 9. Stubble density and stalk diameter were chosen from typical values for a wheat crop in Saskatchewan. A very small amount of exposed wheat stubble (1-2 cm) in this simulation is able to effectively suppress transport and sublimation. The stubble is filled when S reaches 38.5 kg m^{-2}. Blowing snow fluxes are unaffected by S greater than this value and are completely suppressed when S is < 34.5 kg m^{-2}. Between

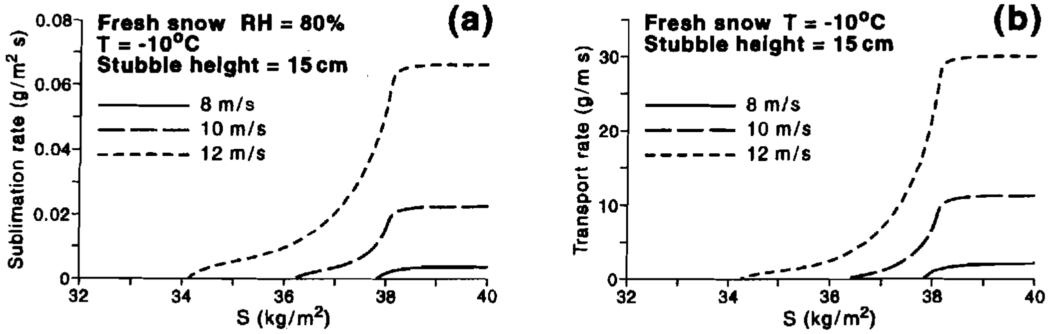


Figure 9. Areal values of (a) downwind transport rate and (b) sublimation rate for x = 1 km, $u(10)$ = 8, 10, and 12 m s^{-1}, $T(2)$ = -10°C, and RH(2) = 80%, as a function of snow accumulation S for a fixed vegetation cover.

these extremes, the rates of increase in transport and sublimation with S are similar, suggesting that while exposed vegetation restrains blowing snow, it does not significantly alter the fluxes relative to one another.

An example of variable fetch effects on areal snow transport, on erosion due to divergence of areal transport, on areal sublimation averaged over the fetch distance, and on snow erosion (negative of accumulation) is shown in Figure 10 for $u(10)$ = 10 m s^{-1}, $T(2)$ = -10°C and RH = 80%. Areal transport reached its steady state value within 500 m of fetch, causing erosion due to transport to rapidly decline with fetch for fetch distances of < 1 km and then to decrease asymptotically to a very small value as fetch became large. Areal sublimation averaged over the fetch increased sharply for fetches < 1 km, after which, increases became small with increasing fetch. An approximate steady state sublimation flux developed for fetches in excess of 3.5 km. Snow erosion due to transport and sublimation decreased sharply with fetch distance up to 1.5 km and then remained constant.

4. Model Demonstration

Application of the blowing snow model to calculate the change in surface snow mass balance (equation 1) is demonstrated in two differing environments: cultivated prairie plains near Regina, Saskatchewan (51°N, 104°W), and an arctic tundra plateau (Trail Valley Creek) south of Tuktoyaktuk, Northwest Territories (69°N, 134°W). Both sites have cold, windy winters, with the arctic site snowier (mean of 150 mm snowfall per season) and colder (winter mean air temperature -19°C) than the Regina site (113 mm and -9°C). Seasons were chosen when snowfall was high, snow cover consistent, blowing snow rather than melt dominated the midwinter mass balance, and the available data to run and evaluate the model were of high quality.

The Meteorological Service of Canada, Environment Canada, collected the Regina data at its primary synoptic station in the high snowfall winter of 1973-1974. Wind speed, air temperature, dewpoint, and precipitation type were measured and recorded every hour, and snowfall and depth of snow on the ground were measured and recorded every 6 hours, by observers near the Regina Airport. The observation site is level with an unobstructed fetch of short grass and pavement in the primary blowing snow direction (WNW). The National Hydrology Research Institute collected Trail Valley Creek data as part of MAGS, the Mackenzie Global Energy and Water Cycle Experiment (GEWEX) Study. Wind speed, air temperature, relative humidity, occurrence of snowfall, and snow depth were recorded by an automated weather station. Snow density and cumulative snowfall were measured at the beginning and end of the simulation period (February-April, 1996).

For Regina, the blowing snow model was run hourly using P, $u(10)$, $T(2)$, RH(2), and precipitation type (snow, rain) to solve for the surface snow accumulation, S as a function of ∇F and E_B. Fetch was set to 1000 m. Vegetation parameters were set for shortgrass as h = 0.01 m, N = 320 m^{-2}, and d_v = 0.003 m. The net ablation from surface evaporation and melt ($E + M$) was estimated from observed decreases in snow depth during periods with above freezing temperatures. Any melted snow was considered ice, which, though it remained on site, was no longer available to contribute to snow transport. For Trail Valley Creek, the model was run half-hourly using $u(2)$, $T(1)$, and RH(1) over a period with complete snow cover and insignificant melt to calculate ∇F and E_B. The fetch was set to 1000 m of tundra. Tundra vegetation was simulated based on plot measurements near the mast of h = 0.08 m, N = 30 m^{-2}, and d_v = 0.003 m. At the end of the simulation period, cumulative snowfall was used to calculate a net snow accumulation.

The simulations of blowing snow and accumulation and the measured snowfall and accumulation for Regina and Trail Valley Creek are shown in Figure 11. Over the winter season at Regina, snowfall was 271 kg m^{-2} of which 114 kg m^{-2} or 42% remained as surface accumulation before the seasonal melt. An estimated 14 kg m^{-2} (5% of snowfall) ablated as melt

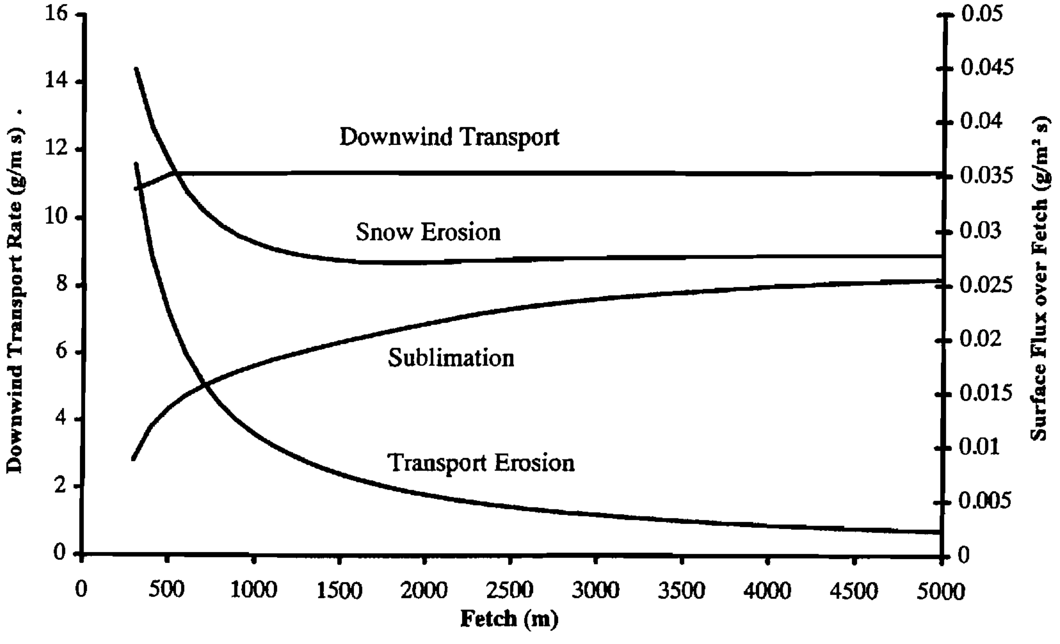
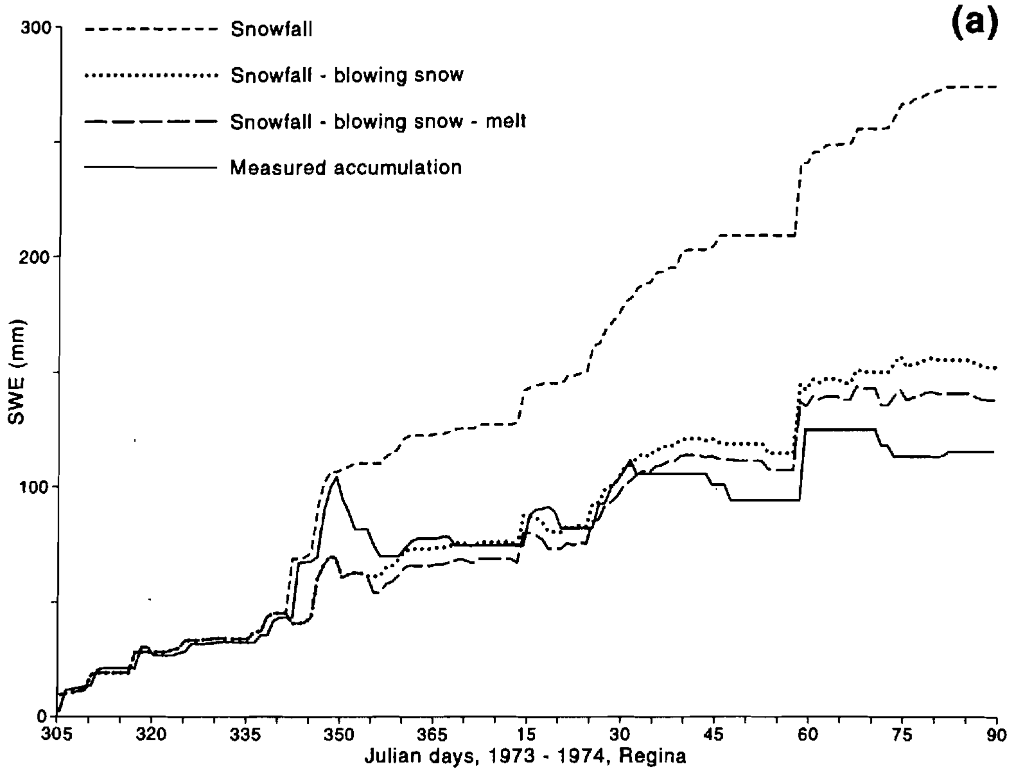


Figure 10. Areal transport, sublimation and erosion averaged over a continuous snow-covered fetch of varying length for $u(10)$ = 10 m s^{-1}, $T(2)$ = -10°C, and RH(2)=80%.



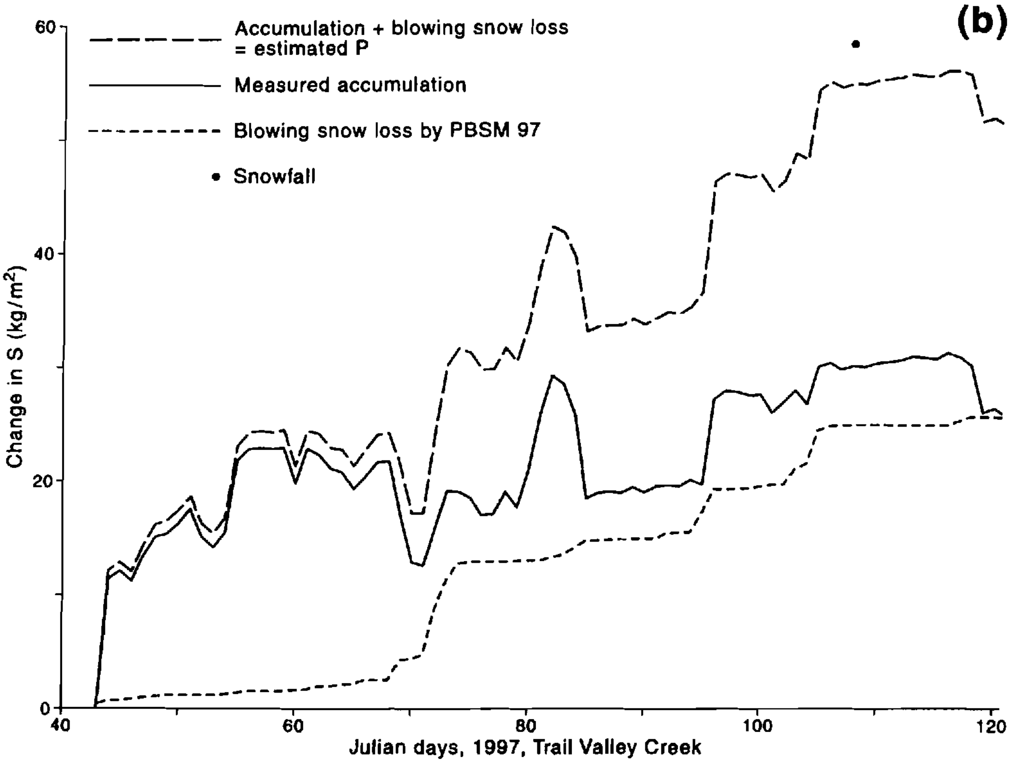


Figure 11. Blowing snow and snow accumulation measurements and simulations for Regina, Saskatchewan and Trail Valley Creek, Northwest Territories. (a) Regina. Measured seasonal snow accumulation and snowfall and modeled snowfall less blowing snow erosion and melt ablation for the winter 1973-1974. (b) Trail Valley Creek. Measured increases in snow accumulation and snowfall and modeled accumulation change plus blowing snow erosion for February-April, 1996.

or surface evaporation, 42 kg m^{-2} (or 42 Mg m^{-1} or 15% of snowfall) was eroded and transported off the fetch, and 79 kg m^{-2} (29% of snowfall) was eroded and sublimated from blowing snow, leaving an unaccounted residual of 22 kg m^{-2} or 8% of snowfall. Over the late winter simulation at Trail Valley Creek, snowfall was 58 kg m^{-2} of which 30 kg m^{-2} or 52% accumulated on the tundra. Transport and sublimation losses over the 65-day simulation totaled 11 and 13 kg m^{-2} or 19% and 22% of snowfall, respectively. The unaccounted residual was 4 kg m^{-2} or 7% of snowfall. For both cases the residual indicates an underestimation of surface snow ablation of 7% to 8%, which may be due to underestimation of blowing snow transport or sublimation or neglect of other ablative processes.

5. Discussion and Conclusions

Regardless of the calculation scheme used to estimate transport and sublimation, parameterizations of the type demonstrated in this model are necessary to apply blowing snow calculations to natural surfaces and to upscale point estimates to the landscape or climate model grid scales. Upscaling is often dependent on the nature of associations between driving parameters; these associations may be restricted to specific environments or in some cases be general [*Blöschl*, 1999]. For instance, the algorithm for calculating the increase in blowing snow fluxes with fetch is not restricted to application in any specific environment as it is based on the physics of plume dispersion. Similarly, the algorithm for calculating the influence of exposed vegetation on shear stress partitioning is based on aerodynamic drag theory and has been transferred in application from blowing dust in Australia to blowing snow in Canada. However, the probability and threshold algorithms are observationally based and have been tested in only two climate regions (prairie, arctic). Their evaluation in other environments is important to assessing whether they can provide a general scaling technique.

Tests in prairie and arctic environments support the model's performance for calculations of snow accumulation using measured wind speed, air temperature and humidity over a season or for shorter periods. Field measurements suggest that the assumption of undersaturation of water vapor near to the snow surface in the blowing snow boundary layer is robust in these environments. At sites in these regions, 48% to 58% of snowfall was removed by blowing snow from the snow cover over the test periods examined, demonstrating that the impact of blowing snow on snow mass balance is quite significant. For snow removed by blowing snow the ratio of sublimation to transport loss was 2:1 for the prairie site and approximately 1:1 for the arctic site. These tests should be extended to other climates and environments. Further research, with careful field observations, will be required to include the effect of complex terrain on these scaling techniques and to more fully evaluate the impact of the evolution of temperature and humidity fields within the blowing snow boundary layer on snow accumulation patterns.

Notation

B snow depth-density covariance parameter, kg m^{-2}.
c_1 saltation velocity proportionality constant, u_p/u^*_t, dimensionless.
c_2 square root of the ratio of initial vertical saltating particle velocity to u^*, dimensionless.
c_3 ratio of z_0 to saltation height, dimensionless.
c_4 vegetation roughness coefficient, m.
c_5 coefficient relating height to vertical diffusion velocity, m^{-1}.
c_6 experimentally determined exponent.
d_S depth of snow, m.
d_V stalk diameter, m.
D diffusivity of water vapor in air, m^2 s^{-1}.
e saltation efficiency, dimensionless.
E sublimation from the snow surface, kg m^{-2} s^{-1}.
E_B sublimation from blowing snow, kg m^{-2} s^{-1}.
E_{BA} seasonal sublimation from blowing snow kg m^{-2}.
F blowing snow downwind transport rate, kg m^{-1} s^{-1}.
F_A annual blowing snow transport, kg m^{-1}.
F_{salt} saltating snow transport rate, kg m^{-1} s^{-1}.
F_{susp} suspended snow transport rate, kg m^{-1} s^{-1}.
g gravitational acceleration constant, m s^{-2}.
h_V vegetation height, m.
h_S latent heat of sublimation, J kg^{-1}.
h^* lower boundary reference height for suspension, m.
I natural logarithm of time since snowfall, hours.
k von Kàrmàn's constant.
M melt flux, kg m^{-2} s^{-1}.
N vegetation density, m^{-2}.
Nu Nusselt number, dimensionless.
p probability of blowing snow occurrence.
P snowfall flux, kg m^{-2} s^{-1}.
q specific humidity, dimensionless.
q_S saturated specific humidity, dimensionless.
r radius of snow particle of mean mass, m.
R Nipher catch ratio, dimensionless.
RH relative humidity, %.
S snow accumulation on the ground, kg m^{-2}.
Sh Sherwood number, dimensionless.
t time, s.
T air temperature, °C.
T_s temperature of a snow particle surface, °C.
u wind speed, m s^{-1}.
u_{mean} mean of wind speed, m s^{-1}.
u^* friction velocity, m s^{-1}.
u^*_t threshold friction velocity for transport, m s^{-1}.
u^*_n non-erodible element friction velocity, m s^{-1}.
u_p saltation downwind velocity, m s^{-1}.
w' effective vertical snow particle velocity in suspension, m s^{-1}.
w^* vertical snow particle velocity ratio, dimensionless.
x fetch distance, m.
z height above the snow surface, m.
z_0 aerodynamic roughness height, m.
z_b upper boundary height for blowing snow, m.
β ratio of roughness element to surface drag, dimensionless.
δ standard deviation of wind speed, m s^{-1}.
η blowing snow mass concentration, kg m^{-3}.
κ_S eddy diffusivity for blowing snow particles, m^2 s^{-1}.
λ roughness density, dimensionless.
λ_t thermal conductivity of air, J (m s K)$^{-1}$.
ρ atmospheric density, kg m^{-3}.
ρ_S snow density, kg m^{-3}.

σ undersaturation of water vapor (1-RH/100), dimensionless.

Ω snow particle of mean mass terminal fall velocity, m s^{-1}.

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