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Turbulent fluxes during blowing snow: field tests of model sublimation predictions

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Abstract:

Sublimation fluxes during blowing snow have been estimated to return 10–50% of seasonal snowfall to the atmosphere in North American prairie and arctic environments. These fluxes are calculated as part of blowing snow two-phase particle transport models with provision for phase change based upon a particle-scale energy balance. Blowing snow models have normally been evaluated based upon their ability to reproduce diagnostic mass flux gradient measurements and regional-scale snow redistribution patterns and snow mass. Direct evidence is presented here that large latent heat fluxes (40–60 W m^{-2}) that result in sublimation rates of 0·05–0·075 mm snow water equivalent $hour^{-1}$, are associated with mid-winter, high-latitude blowing snow events. For events with wind speeds above the threshold level for snow transport, these fluxes are in the range of those predicted by the Prairie Blowing Snow Model. The fluxes are well in excess of those found during spring snowmelt and which can be predicted by standard bulk aerodynamic transfer equations, suggesting that blowing snow physics will have to be incorporated in land surface schemes and hydrological models in order to properly represent snow surface mass and energy exchange during blowing snow events. Copyright © 1999 John Wiley & Sons, Ltd.

KEY WORDS blowing snow; sublimation; latent heat; turbulent transfer; snow cover; Canadian Prairies

INTRODUCTION

Algorithms that describe snow processes should be important components of cold regions land surface schemes and hydrological models. Land surface schemes prescribe the lower boundary layer of general circulation and numerical weather models and increasingly comprise an important part of hydrological models. One such process is transport and sublimation of blowing snow by wind. Snow transport redistributes snow covers and produces notable in-transit sublimation of blowing snow (Dyunin, 1959; Schmidt, 1972; Pomeroy, 1989). Reported annual fluxes of blowing snow sublimation range from 15% to over 40% (depending on climate, fetch and land cover) of annual snowfall on the Canadian Prairies (Pomeroy and Gray, 1995), 20% to 47% of annual snowfall on tundra in the Western Canadian Arctic (Pomeroy *et al*., 1997; Essery *et al*., 1999) and 9% to 32% of annual snowfall on the Alaska north slope, depending on topography (Benson, 1982; Liston and Sturm, 1998).

Land surface schemes and hydrological models do not presently incorporate blowing snow processes, because existing snow transport and sublimation parameterizations are not well-known and accepted and because direct flux measurements that would demonstrate the need for such parameterizations have not been

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available. The objectives of this paper are to examine the importance of coupled blowing snow transport and sublimation processes in defining atmospheric water vapour fluxes during a snow accumulation period in a Prairie environment.

THEORY

For a control volume consisting of the surface snow cover and the blowing snow mass above it, the mass and energy fluxes are coupled by latent heat transfer between solid and vapour phases (Figure 1), where

$$Q_E = -\lambda(E + E_B) \tag{1}$$

and Q_E is the turbulent latent heat flux directed towards the control volume, λ is the latent heat of sublimation, E is evaporation flux from the surface snow cover and E_B is sublimation of blowing snow. The latent heat flux is a component of the energy balance which can be written as:

$$\frac{dU}{dt} = Q^* + Q_H + Q_E + \nabla \cdot Q_P + \nabla \cdot Q_A + Q_G \tag{2}$$

where U is internal energy, Q^* is net radiation and Q_H is the turbulent sensible heat flux to the control volume, $\nabla \cdot Q_P$ is the divergence of energy transported to and from the control volume by precipitation or blowing snow, $\nabla \cdot Q_A$ is the divergence of energy carried by turbulent transfer to the control volume, and Q_G

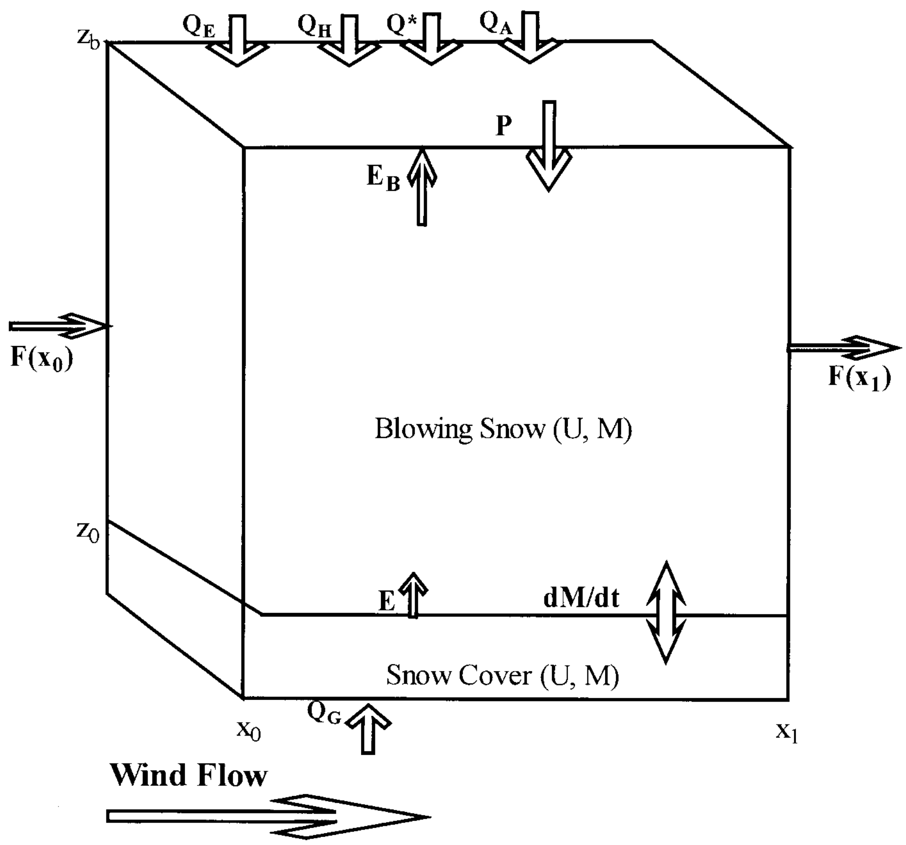


Figure 1. Control volume of blowing and surface snow with energy and mass fluxes indicated

is the heat flux from the ground to the snowpack base. The mass balance of a blowing snow and surface snow control volume can be written as:

$$\frac{dM}{dt} = P - \nabla \cdot F - E - E_B \tag{3}$$

where M is snow mass, P is snowfall flux and F is horizontal blowing snow mass flux. This mass balance is linked to equation (2) through equation (1). Considerable research has gone into the evaluation of each term of equations (2) and (3) and will be briefly summarized below.

Q^* is found as balance of incoming and outgoing short and long wave radiation, including the effects of the blowing snow layer on radiation exchange in the control volume. Whilst blowing snow scatters shortwave radiation (Pomeroy and Male, 1988), the surface layer albedo is relatively unaffected (Yamanouchi and Kawaguchi, 1985), though downward net radiation decreases by 2 W m^{-2} per 1 m s^{-1} increase in wind speed above 13 m s^{-1} (Yamanouchi and Kawaguchi, 1985). This effect is possibly similar to that observed by Granger and Male (1978) and Halberstam and Schlieldge (1981) in which a layer of moist air of differing temperature than the surface affected longwave emission. Radiative characteristics of open snow covers are otherwise well-known (Male and Granger, 1981). Internal energy change U during blowing snow is associated primarily with changes in snowpack temperature and does not normally involve phase change, as wet snow covers are cohesive and require extremely high winds to sustain transport (Li and Pomeroy, 1997a,b). However, blowing snow particles are presumed to rapidly reach the ice bulb temperature upon entrainment and this energy flux has not been considered in energy balance calculations. Ground heat fluxes depend on the season and can be large in early winter. For blowing snow conditions, $\nabla \cdot Q_P$ is small because rainfall dramatically increases snow surface cohesion and immediately suppresses transport, snowfall is usually eroded and transported, and over large fields or fetches snow transport divergence is small. Advection of energy may be important for short fetches, particularly where exposed vegetation upwind may sustain a much higher net radiation than an open snow cover. Over continuous snow cover and long fetches, $\nabla \cdot Q_A$ may be considered negligible.

Over stationary snow surfaces the turbulent fluxes in land surface schemes and physically-based hydrological models are often calculated using Dalton-type bulk transfer equations (e.g. Verseghy *et al.*, 1993), where:

$$\begin{aligned} Q_H &= \rho_a c_p C_H u(z)[\theta_a(z) - \theta_o] \\ Q_E &= \rho_a \lambda C_E u(z)[q(z) - q_o] \end{aligned} \tag{4}$$

and ρ_a is atmospheric density, c_p is specific heat of air, C_H and C_E are the bulk transfer coefficients for sensible and latent heat respectively, u is the horizontal wind speed at height z, θ is the potential air temperature (a, ambient; o at the snow surface) and q is specific humidity (at the snow surface, o, it is presumed to be saturated). When stability is not a concern, the bulk-transfer coefficients (also considered the inverse of 'resistances') depend upon the chosen roughness lengths for momentum (z_o) and heat or water vapour transfer (z_{oH}, z_{oE}), where, substituting subscript Y for either H or E:

$$C_Y = \frac{k^2}{\ln\left(\frac{z}{z_o}\right)\ln\left(\frac{z}{z_{oY}}\right)} \tag{5}$$

and k is the von Kármán constant (0·4). Male (1980) notes that bulk transfer formulae provide poor estimates of Q_E during low wind speeds and high radiation because of the occurrence of shallow stable layers directly over the snow cover and unstable layers above the stable layer. This problem should occur less often during blowing snow conditions because of high wind speeds and strong mixing. For most model applications, z_o is given a fixed value from a table, and z_{oY} is found as a ratio of z_o.

A similar form can be used to describe the calculation of energy transfer with blowing snow particles:

$$
\begin{aligned}
Q_{EB} &= \int_0^{z_b} \frac{1}{m(z)} 2\pi r(z)\rho_a \lambda DSh(q(z) - q_p)\eta(z)dz \\
Q_{HB} &= \int_0^{z_b} \frac{1}{m(z)} 2\pi r(z)D_T Nu(T(z) - T_B)\eta(z)dz
\end{aligned} \tag{6}
$$

where T is air temperature, the subscript B refers to blowing snow particles, m is the mean mass of a blowing snow particle at height z above the snow surface, z_b is the top of the blowing snow column, r is the radius of a particle of mean mass at height z, D is the diffusivity of water vapour in air, Sh is the Sherwood Number, D_T is the thermal conductivity of air, Nu is the Nusselt Number and η is the mass concentration of blowing snow at height z. The sensible and latent heat fluxes to a blowing snow particle are balanced presuming thermodynamic equilibrium and that T_B is at the ice bulb temperature (Thorpe and Mason, 1966). Equation (6) is parameterized for Prairie and Arctic environments, implemented in the Prairie Blowing Snow Model (PBSM) and coupled to snow mass transport equations through r and η (Pomeroy *et al.*, 1993; Pomeroy and Li, 1997). Similar forms are used by Déry *et al.* (1996; 1998) and Liston and Sturm (1998). Note that if E is small, then Q_E is determined by equation (6).

The divergence of blowing snow transport $\nabla \cdot F$ is found by comparing blowing snow transport fluxes entering and leaving the control volume. If the control volume is oriented along flow lines and if its top is defined as the height of the blowing snow column, then divergence is found as the difference of blowing snow entering the upwind edge and leaving the downwind edge of the control volume of downwind length $x_1 - x_0$;

$$
\nabla \cdot F = \frac{F(x_1) - F(x_0)}{x_1 - x_0} \tag{7}
$$

Techniques to calculate the blowing snow mass flux (saltation and suspension) as a function of meteorological conditions, snow conditions and fetch are presented by Pomeroy and Gray (1990), Pomeroy and Male (1992), Pomeroy *et al.* (1993), Pomeroy and Gray (1995), Li and Pomeroy (1997a,b) and Pomeroy and Li (1997) and are compiled in PBSM. Note that as the length scale $x_1 - x_0$ becomes large $\nabla \cdot F$ becomes very small.

EXPERIMENTAL SITES AND DATA COLLECTION

Kernen Farm (500 m asl) is east of the City of Saskatoon (52 °N, 107 °W) in the central southern half of the Province of Saskatchewan, Canada. The farm is situated on an open, flat, lacustrine plain, which is cropped to cereal grains and pulse crops under the practice of dryland farming (Shook and Gray, 1997). Trees are limited to farmyards, which are located several kilometres distant from the site. The climate is sub-humid and typical of northern prairies with cold winters and continuous snow cover from late November through early April.

Blowing snow experiments were conducted in December 1998 and January 1999 at a level site with continuous snow cover over fields of uniform short vegetation or fallow. During the measurement periods, the sites were frequently manned, which provided a high confidence in the observations.

Energy balance and related parameters were measured and recorded on a Campbell 23X datalogger half-hourly using eddy correlation equipment at a 2 m height (Campbell Scientific 'CSAT' sonic anemometer, 'Krypton' hygrometer and fine wire thermocouple), Radiation and Energy Balance Systems net radiometers and soil heat flux plates, NRG cup anemometer, Everest infrared thermometer, Campbell/Vaisala platinum resistance hygrothermometers, Campbell Scientific ultrasonic snow depth gauge and University of Saskatchewan blowing snow particle counters. The instrument models, resolution and estimated accuracy are provided in Table I.

Table I. Instrument models used, parameter measured, resolution, time response, estimated accuracy. Information assembled from the manufacturers and tests at NHRC

Instrument	Parameter	Time Response	Resolution	Estimated Accuracy
Campbell CSAT3 Sonic Anemometer	w', u', v', T'	60 Hz	0·25–2 mm s^{-1} 0·002 °C	$< \pm 5$ W m^{-2} using 23X covariance
12·7 μm Type E Thermocouple	T'	30 Hz	0·01 °C	$< \pm 3$ W m^{-2} using 23X covariance
Campbell KH20 Krypton Hygrometer	$\rho_v{}'$ (vapour pressure)	100 Hz	0·001245 g m^{-3}	$< \pm 5$ W m^{-2} using 23X covariance
Campbell/Vaisala HMP35CF Hygrothermometer	T, RH	15 s	0·1 °, %	$< \pm 0{\cdot}2$ °C, $< \pm 3\%$ RH
Campbell SR50 Ultrasonic Snow Depth Gauge	Depth of snow	60 s	0·1 mm	1 cm
REBS Q7 Net Radiometer	Q^*	1 s	0·013 W m^{-2}	± 2 W m^{-2}
REBS Soil Heat Flux Plate	Q_g	1 s	0·047 W m^{-2}	± 2 W m^{-2}
Everest 4000 Infrared Thermometer	T_s	10 s	0·1 °C	± 2 °C
University of Saskatchewan Snow Particle Detector	Number flux of blowing snow particles	60 s (totalization)	0·033 #/sec	$\pm 10\%$

The Campbell/Vaisala HMP35CF hygrothermometer is modified from the original Vaisala gauge by Campbell Scientific Canada for low temperature performance. It was calibrated by Campbell Scientific Canada and the National Hydrology Research Centre (NHRC) for a range of humidities at temperatures greater than −20 °C; the calibration was used as a polynomial in the datalogger programme. We feel that the device does not experience the low temperature deviations described by Anderson (1994) for the HMP35A; in any case very small sublimation values were measured and predicted at temperatures below −20 °C. Cup anemometers were tested in the field against wind tunnel calibrated anemometers to develop a correction. The eddy correlation equipment is calibrated by the manufacturer (e.g. Krypton Hygrometer against a chilled mirror hygrometer) and designed for operation to −30 °C. The CSAT3 sonic anemometer has an internal signal processing quality control procedure which involves a six fold oversampling to eliminate noise from precipitation (e.g. blowing snow) particles near sensor heads. It also provides a quantitative signal quality diagnostic; the diagnostic was acceptable for the measurements used. A fine wire thermocouple was used to provide fast air temperature measurements before and in the early part of each blowing snow event. However a few hours into each event the thermocouple was disintegrated by the severe conditions and fast air temperature was then calculated using the sonic anemometer. The accuracy of temperature measurement using the sonic anemometer is not quite as good as the thermocouple so the two measures were compared for the period before thermocouple disintegration and a calibration curve for each event fitted to minimize error.

To calculate sensible and latent heat fluxes using eddy correlation equipment the surface boundary layer is assumed to be in a steady state where vertical fluxes dominate turbulent exchange with the surface. Q_H and Q_E are found from the covariances of the instantaneous measurements of T' and q' with the instantaneous vertical wind velocity w', where

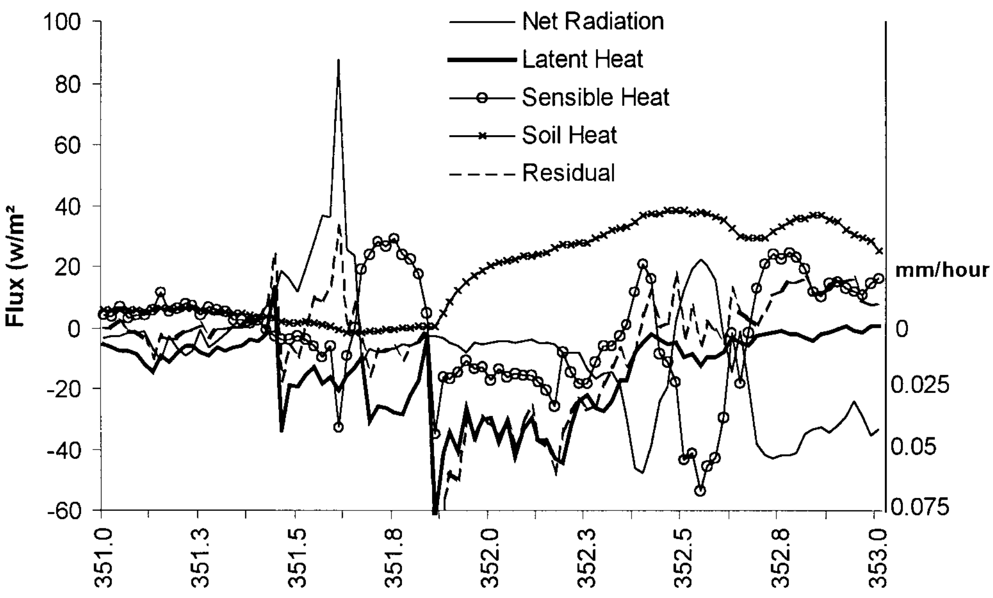
$$\begin{aligned} Q_H &= \rho_a c_p \overline{w'T'} \\ Q_E &= \rho_a \lambda \overline{w'q'} \end{aligned} \tag{8}$$

The limitations and assumptions of this procedure are discussed in standard textbooks such as Oke (1978) and Stull (1988).

MEASUREMENTS

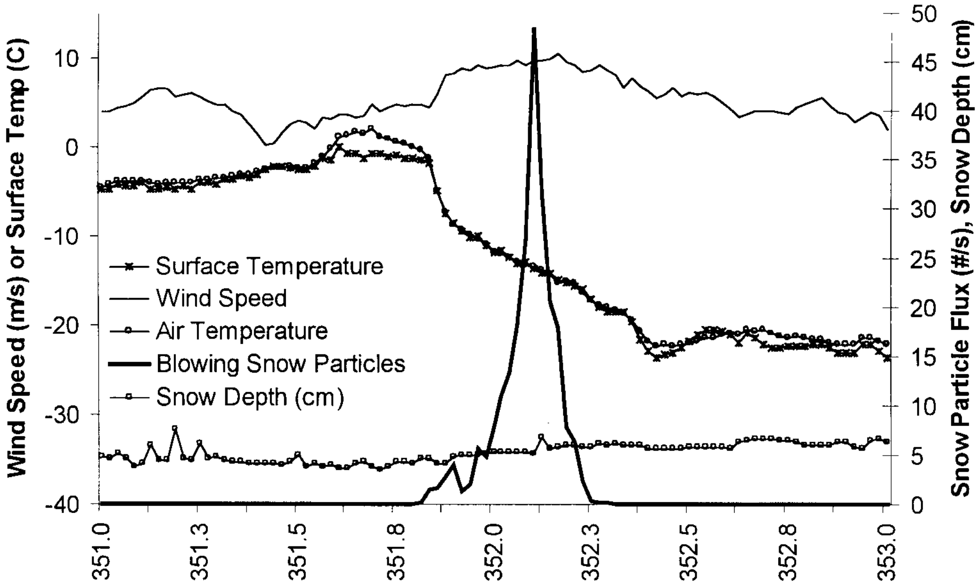
The measured energy balances and related hydrological and atmospheric parameters are shown in Figures 2 and 3 for both a stronger and a weaker blowing snow event with concurrent snowfall. Positive values indicate fluxes to the control volume. The blowing snow events show characteristic Prairie weather patterns with highly variable meteorology associated with the passage of frontal systems. In particular the stronger event was associated with the entry of an Arctic high pressure system; the northwest winds associated with these systems results in the majority of blowing snow transport and sublimation in this region (Pomeroy and Gray, 1995). For the stronger event (Figure 2), on JD 351 the air temperature warmed to slightly above freezing, then dropped to $-24\,^{\circ}$C in about 12 hours during a period of strong winds (peak at 1·3 m height, wind 11 m s^{-1}; friction velocity 0·95 m s^{-1}). During this cooling period snowfall and blowing snow were recorded around midnight and the lower boundary layer remained well-mixed. Vapour pressure dropped from 0·5 to 0·05 kPa with relative humidity (with respect to water) declining from 97% to 78%. Such a decline in relative humidity is not consistent with the model result of Déry *et al.* (1998) who suggest an increase in humidity towards near-saturation as a blowing snow event proceeds. The blowing snow event did not result in an observed increase in snow depth, and density after the event reached 140 kg m^{-3} due to impact of saltating snow particles and subsequent sintering.

During short periods around mid-day net radiation was small but positive (peak Q^* 20–90 W m^{-2}), however at other times it was negative, smaller than -10 W m^{-2} during cloudy nights and dropping as low as -40 W m^{-2} (Figure 2). Turbulent fluxes were enhanced over that expected from smooth snow covers because of exposed sparse vegetation, and were strongly affected by the occurrence of blowing snow. During the



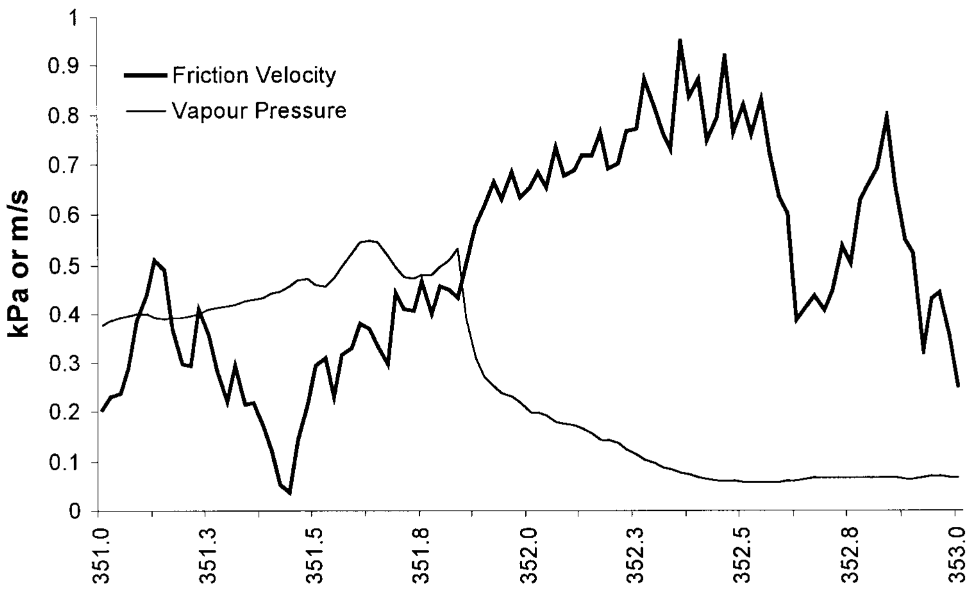
a)

Figure 2(a). Measurements at Kernen Farm near Saskatoon, 17–18 December 1998. Net radiation (1 m), latent heat (2 m), sensible heat (2 m), soil heat flux ($-0{\cdot}05$ m) and residual energy ($dU/dt - \nabla \cdot Q_p - \nabla \cdot Q_a$)



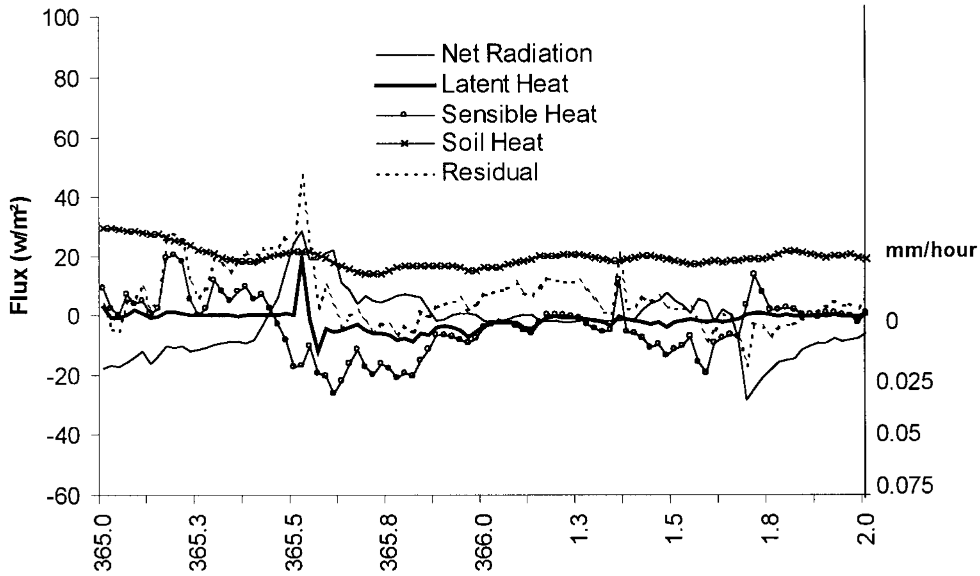
b

Figure 2(b). Measurements at Kernen Farm near Saskatoon, 17–18 December 1998. Snow surface temperature, wind speed (1·3 m), air temperature (1·3 m), blowing snow particle flux (0·2 m) and snow depth



c)

Figure 2(c). Measurements at Kernen Farm near Saskatoon, 17–18 December 1998. Friction velocity (2 m) and water vapour pressure (1·3 m)



a)

Figure 3(a). Measurements at Kernen Farm near Saskatoon, 31 December 1998–1 January 1999. Net radiation (1 m), latent heat (2 m), sensible heat (2 m), soil heat flux (−0·05 m) and residual energy ($dU/dt - \nabla \cdot Q_p - \nabla \cdot Q_a$)

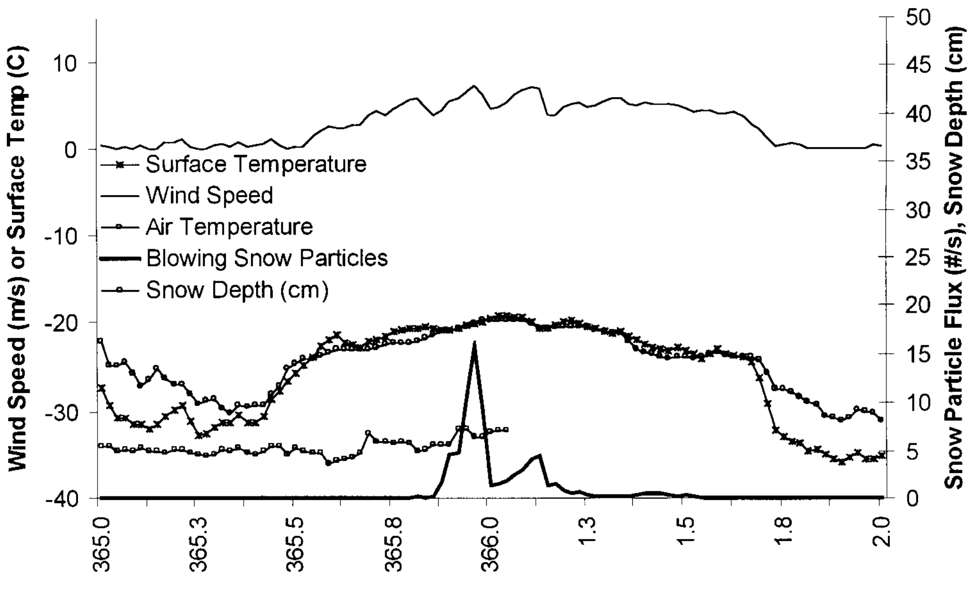


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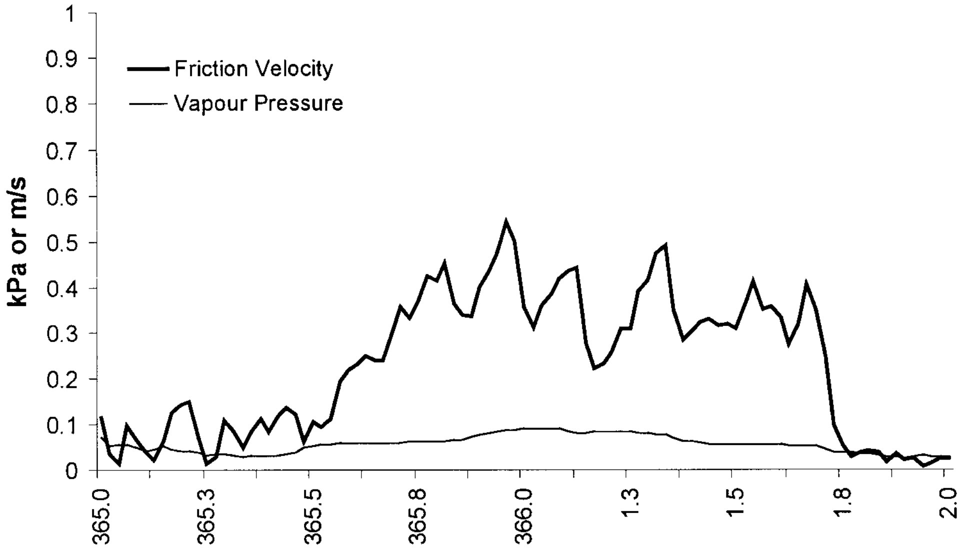
Figure 3(b). Measurements at Kernen Farm near Saskatoon, 31 December 1998–1 January 1999. Snow surface temperature, wind speed (1·3 m), air temperature (1·3 m), blowing snow particle flux (0·2 m) and snow depth



c)

Figure 3(c). Measurements at Kernen Farm near Saskatoon, 31 December 1998–1 January 1999. Friction velocity (2 m) and water vapour pressure (1·3 m)

blowing snow event the latent heat flux peaked at -60 W m^{-2}, and showed a smaller peak during the above-freezing air temperatures early on JD 351, but was otherwise small during non-blowing snow periods. The latent heat flux during blowing snow was over twice the maximum levels measured from continuous, melting prairie snow covers under high radiation spring conditions by Granger and Male (1978). The sensible heat flux mimicked the pattern and direction of latent heat during blowing snow but at one-half the magnitude. Sensible heat peaked at -50 W m^{-2} during a cold, relatively calm, high radiation period (JD 352·6). During cold, relatively calm, negative net radiation periods, sensible heat was generally slightly positive, peaking at 30 W m^{-2} but most often less than 20 W m^{-2}. A notable flux in this early winter period was the positive ground heat flux into the snowpack which became prominent (30–40 W m^{-2}) upon cooling of snow and air.

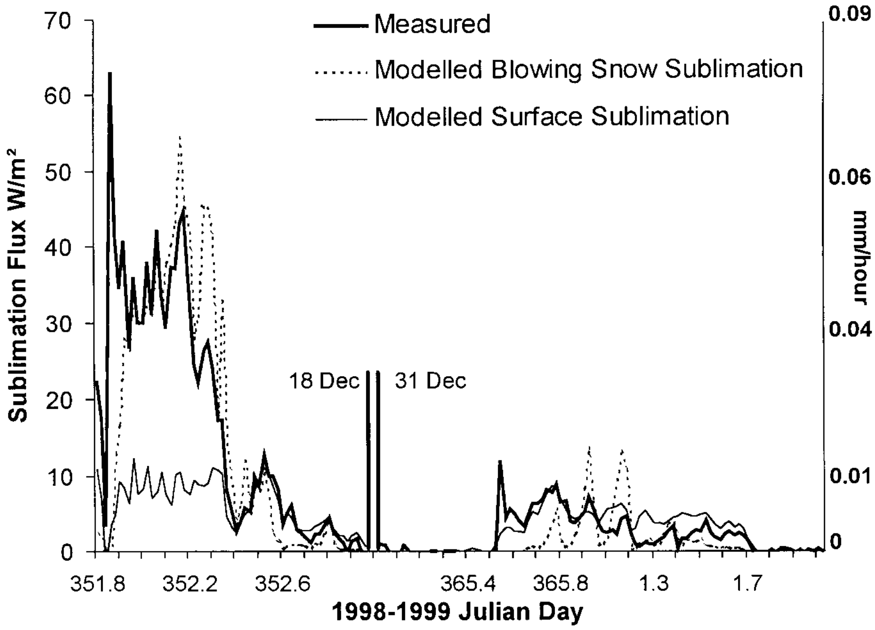
Measured latent heat fluxes indicate sublimation rates of between 0·05 and 0·075 mm snow water equivalent/hour during the blowing snow event. Such rates if persistent, could lead to sublimation of almost 2 mm water equivalent/day, a significant loss of snow from the dry prairie environment and sufficiently high to cause the seasonal snow ablation in this region reported by Pomeroy and Gray (1995). The high sublimation rates during blowing snow were distinctive from fluxes measured before and after in that they were not well balanced by measured vertical energy fluxes. The 'residual' ($dU/dt - \nabla \cdot Q_p - \nabla \cdot Q_a$) flux was large during blowing snow and very similar to the latent heat flux. Such an imbalance in 'standard' energy fluxes was measured by Harding and Pomeroy (1996) during sublimation of intercepted snow from a coniferous forest canopy (using different eddy correlation instrumentation) and confirmed with a separate measurement of snow sublimation from a weighed, suspended tree. They suggested possible energy sources such as release of stored canopy heat or convective cells in the canopy. Similar apparent 'imbalances' in the energy budget have been observed over wet forest canopies where evaporation exceeds available energy (McCaughey *et al.*, 1997). Shuttleworth (1989) has suggested meso-scale advection accounts for the additional energy. The rapid cooling in this case could have provided a heat source in snowpack and control volume internal energy. It is worthy of future consideration whether turbulent sweep-ejection processes in the blowing snow boundary layer at longer time scales than the covariance calculation (5 minutes) could have advected warmer drier air into the lower boundary layer and sustained both the relatively unsaturated conditions and high sublimation rates.

During the weaker blowing snow event near midnight on JD 365, air temperatures varied between -25 and -20 °C and were well mixed (Figure 3). Wind speed peaked at over 7 m s^{-1} ($u^* = 0{\cdot}55$ m s^{-1}) providing 'just-above' threshold conditions for snow transport (Li and Pomeroy, 1997b). Relative humidity held near 80% and vapour pressure ranged from 0·05 to 0·1 kPa, with little response to the onset of blowing snow. The snow depth sensor indicated a small accumulation of snow before it failed during the blowing snow event. Snow density remained low during this event with an average a few days later of only 110 kg m^{-3}. Net radiation during blowing snow was negligible and the daily peak previous was less than 30 W m^{-2}. Sensible and latent heat fluxes were smaller than -10 W m^{-2} (sublimation flux less than 0·01 mm/hour), with similar temporal patterns. Much larger turbulent fluxes had occurred earlier on JD 365 during a positive net radiation period. Ground heat flux was between 15 and 20 W m^{-2} during this period, providing a relatively large and steady source of energy. The residual energy fluctuated around zero.

FIELD EVALUATIONS OF MODELS

Winter latent heat flux is of interest to both hydrologists and meteorologists because of its impact on dry winter atmospheres and effect on evolution of snow cover and supply of spring melt water through overwinter sublimation loss. An evaluation of the Prairie Blowing Snow Model calculation of latent heat flux due to blowing snow sublimation (equation (6)) and that calculated using surface bulk transfer formulations (equation (4)) with a fixed roughness height, was conducted using the measurements. The models have very few variables that can be pre-set. For roughness length modelling in both schemes, 1 cm of exposed vegetation above the snowpack was presumed. The blowing snow model (Pomeroy and Li, 1997) was driven by measured air temperature and humidity and wind speed (1·3 m height); the upper boundary layer for sublimation calculations was artificially set to 1 m which roughly corresponded to the greatest height for generation of water vapour that could be detected by the eddy correlation system (at 2 m). Concurrent or fresh snowfall was assumed. The bulk transfer latent heat calculation used measured snow surface and air temperature (1·3 m), measured humidity at 1·3 m, measured wind speed, an assumed roughness length of 1 mm and assumed saturation at the snow cover surface. The effective roughness length for water vapour transfer was considered 0·1 of that for momentum as is standard practice in land surface schemes (Essery, 1997). No stability correction (e.g. by Richardson number) was employed because of the low measurement height and near-neutral conditions. As both models used measured values of temperature and humidity at heights near 1 m, both included any 'feedback effects' from blowing snow sublimation as postulated by Déry *et al*. (1998). Evidence of such effects may have been masked by meso-scale advection and other exchange processes modified by two-phase flow. Certainly the development over time of nearly-saturated conditions as predicted by Déry *et al*. was not observed, however a much stronger cooling than that predicted by their model was observed during and after the stronger blowing snow event and is associated with the advent of an arctic cold front.

The latent heat flux measured by the eddy correlation system and calculated using the blowing snow model and the Dalton-type bulk transfer scheme are shown in Figure 4 for the two blowing snow events. It is instructive to consider both the time series pattern of modelled and measured fluxes and the comparison of modelled and measured fluxes. During the first blowing snow event, measured sublimation fluxes were large (range 0·038–0·076 mm h^{-1} or 30–60 W m^{-2}) and are strongly under-predicted by the bulk transfer calculation (typically 0·013 mm h^{-1} or 10 W m^{-2}) but better-predicted by the blowing snow model (range 0·038–0·07 mm h^{-1} or 30–55 W m^{-2}). As shown in Table II, the mean error of PBSM during the first event was $-1{\cdot}4$ W m^{-2} (0·0018 mm h^{-1}) whilst that for the surface bulk-transfer formulation was $-13{\cdot}0$ W m^{-2} (0·016 mm h^{-1}) with correlation coefficients of 0·71 and 0·51 respectively. During the second blowing snow event, measured sublimation fluxes were smaller ($<0{\cdot}013$ mm h^{-1} or 10 W m^{-2}) and were correlated better



a)

Figure 4(a). Measured sublimation flux, modelled blowing snow sublimation (PBSM) and modelled surface sublimation (bulk transfer). Time series

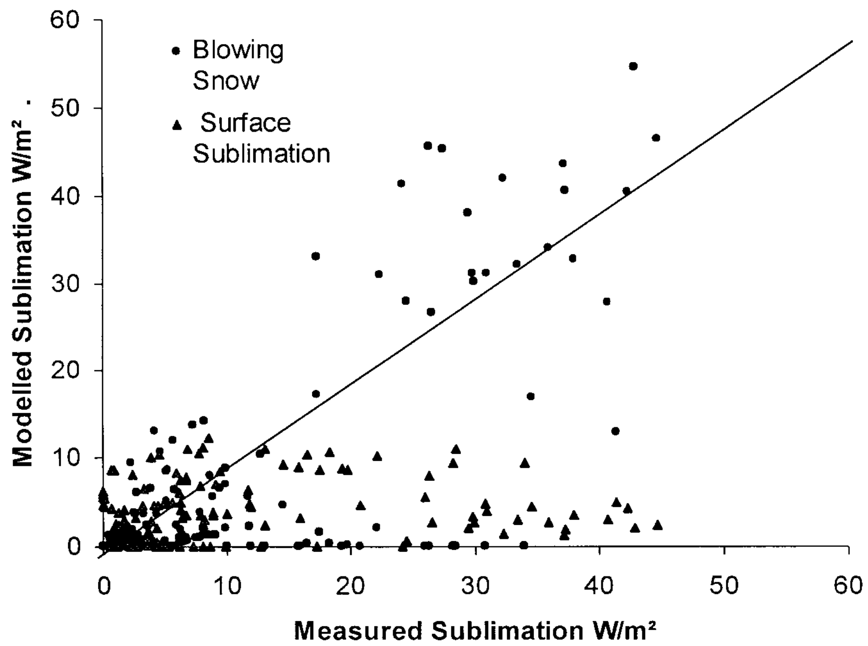
Table II. Prairie Blowing Snow Model and surface bulk transfer model performance in calculating latent heat flux compared to measured latent heat fluxes during the blowing snow measurement periods

Julian Day	r^2		Mean Error (W m^{-2})		Standard Deviation of Error	
	PBSM	Bulk Transfer	PBSM	Bulk Transfer	PBSM	Bulk Transfer
351·8–352·8	0·71	0·51	−1·4	−13·0	12·6	14·2
365·8–1·65	0·49	0·70	−0·3	1·6	3·1	1·4
Overall	0·82	0·69	−0·8	−3·7	7·6	10·4

to the surface bulk transfer calculation (correlation coefficient 0·7 vs. 0·49 for PBSM) but PBSM had the smaller mean error, −0·3 W m^{-2} vs. 1·6 W m^{-2} for the bulk-transfer model. Over the periods shown, PBSM (r^2 = 0·82, mean error = −0·83 W m^{-2}, standard deviation of error = 7·6 W m^{-2}) performed better than the surface bulk-transfer model (r^2 = 0·69, mean error = −3·7 W m^{-2}, standard deviation of error = 10·4 W m^{-2}). The important observation however, is that as blowing snow events become stronger the error in not including blowing snow physics in turbulent exchange and snow mass balance calculations becomes larger.

IMPLICATIONS AND CONCLUSIONS

Most hydrological models and land surface models do not presently characterize blowing snow redistribution and sublimation; the results presented here suggest however, that consideration of blowing snow physics



b)

Figure 4(b). Measured sublimation flux, modelled blowing snow sublimation (PBSM) and modelled surface sublimation (bulk transfer). Modelled versus measured values

is necessary to characterize latent heat fluxes during wind transport of snow. Mid-winter latent heat fluxes during blowing snow can reach 60 W m^{-2} for even a moderately windy snow storm on the Canadian Prairies; these fluxes are over twice that found over melting prairie snow covers in high radiation conditions and can result in sublimation rates of nearly 2 mm snow water equivalent/day. Latent heat flux during blowing snow was poorly predicted by a commonly-employed surface bulk turbulent transfer scheme, even when driven with measured wind speed, surface and atmospheric temperatures and humidity, but can be relatively-well simulated using the Prairie Blowing Snow Model. The improvement in simulation using the blowing snow model increased with increasing wind speed and amounted to 35 W m^{-2} (0·044 mm h^{-1}) during the peak blowing snow period. The variance in the difference between blowing snow model predicted and measured latent heat fluxes during blowing snow and the imbalance in measured energy fluxes during the strong blowing snow event suggest the need for further improvements in the understanding and parameterization of multi-phase blowing snow flow.

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