

Turbulent Fluxes During Blowing Snow: Field Tests of Model Sublimation

J. POMEROY,¹ AND R. ESSERY¹

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 ($40-60 \text{ W m}^{-2}$) that result in sublimation rates of $0.05-0.075 \text{ mm snow water equivalent hour}^{-1}$, 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.

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 (Dyumin, 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 & 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 parameterisations are not well-known and accepted and because direct flux measurements that would demonstrate the need for such parameterisations have not been available. The objectives of this paper are to examine the

¹ National Hydrology Research Centre, 11 Innovation Boulevard, Saskatoon, Saskatchewan S7N 3H5, Canada (Institute of Geography and Earth Sciences, University of Wales, Aberystwyth, Wales SY23 3DB, U.K., after 1 February 2000)

² Division of Hydrology, University of Saskatchewan, Saskatoon, Saskatchewan S7N 0W0, Canada (currently at Hadley Centre for Climate Prediction and Research, United Kingdom Meteorological Office, London Road, Barchnell, England RG12 2SY, U.K.)

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 (Fig. 1), where

$$Q_E = -\lambda(E + E_B) \quad 1)$$

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:

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

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 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:

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

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

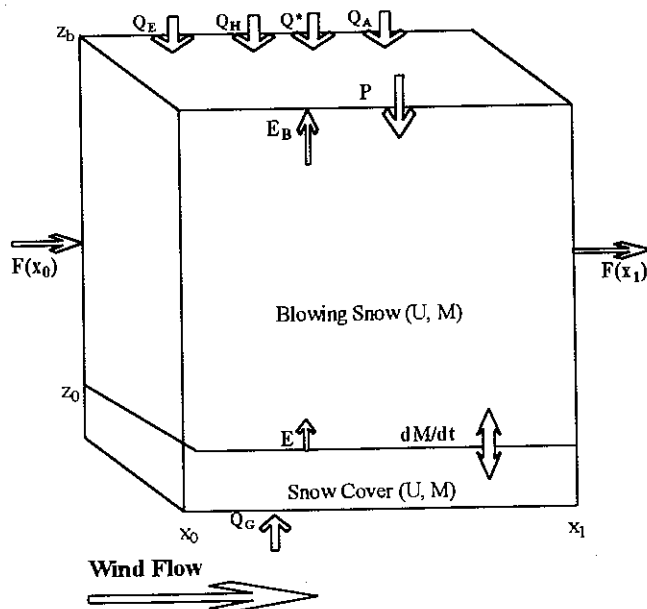


Figure 1. Control volume of blowing and surface snow with energy and mass fluxes indicated.

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 Schliedje (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, 1997b). 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 high net radiation than an open snowcover. 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:

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

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:

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

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 snowcover 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:

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

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 parameterised 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_B is determined by Eq. 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 $x_1 - x_0$;

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

Techniques to calculate the blowing snow mass flux (saltation and suspension) as a function 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; 1997b) 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.a.s.l.) 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 snowcover 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 1.

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 (Anderson, 1994), 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

Table 1. 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	< ± 5 W m ⁻² using 23X covariance
12.7 µm Type E Thermocouple	T'	30 Hz	0.01°C	< ± 3 W m ⁻² using 23X covariance
Campbell KH20 Krypton Hygrometer	ρ_v' (vapour pressure)	100Hz	0.001245 g m ⁻³	< ± 5 W m ⁻² using 23X covariance
Campbell/Vaisala HMP35CF Hygro-thermometer	T, RH	15 s	0.1 °, %	<±0.2°C, <±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 totalisation	0.033 #/sec	±10%

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 minimise 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

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

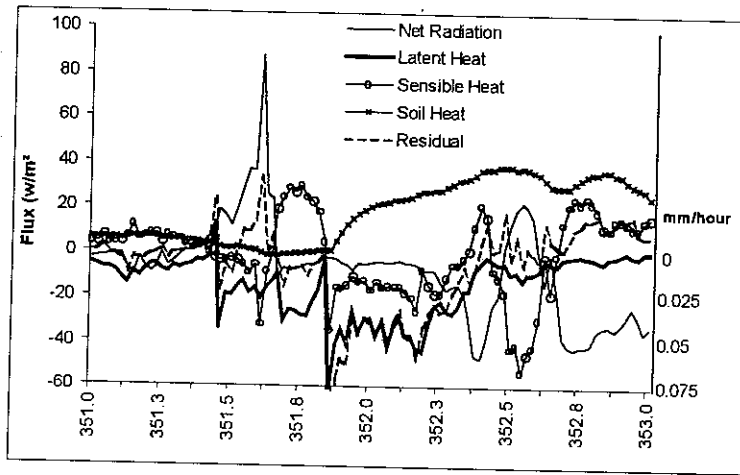
The limitations and assumptions of this procedure are discussed in standard textbooks such as Oke (1978) and Stull (1988).

MEASUREMENTS

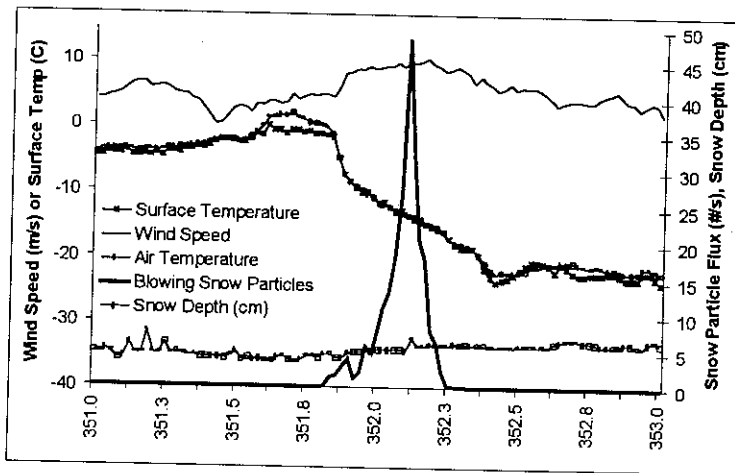
The measured energy balances and related hydrological and atmospheric parameters are shown in Figs. 2 and 3 for both a stronger and a weaker blowing snow event with concurrent snowfall. Positive values indicate fluxes to the control volume. The blowing snow events show characteristic Prairie weather 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 (Fig. 2), on JD 351 the air temperature warmed to slightly above freezing, then dropped to -24°C in about 12 hours during a period of strong winds (peak at 1.3 m height, wind 11 m/s; friction velocity 0.95 m/s). 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 $Q^* 20\text{-}90\text{ W m}^{-2}$), 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} (Fig. 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 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 ($30\text{-}40\text{ W m}^{-2}$) upon cooling of snow and air.

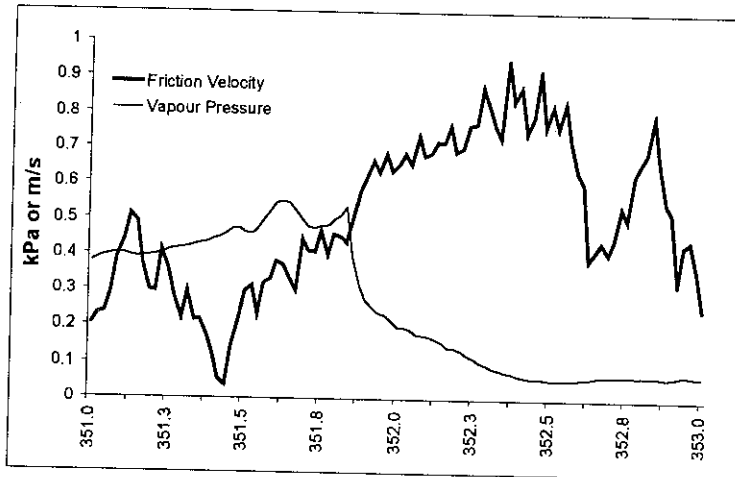
Measured latent heat fluxes indicate sublimation rates of between 0.05 and 0.075 mm snow water equivalent per hour during the blowing snow event. Such rates if persistent, could lead to sublimation of almost 2 mm water equivalent per day, a significant loss of snow from the dry prairie environment and 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 (McCaughy 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.



a)



b)



c)

Figure 2. Measurements at Kernan Farm near Saskatoon, 17-18 Dec. 1998.
 a) Net radiation (1 m), latent heat (2 m), sensible heat (2 m), soil heat flux (-0.05 m) and residual energy ($dU/dt - \nabla \cdot Q_p - \nabla \cdot Q_a$). b) Snow surface temperature, wind speed (1.3 m), air temperature (1.3 m), blowing snow particle flux (0.2 m) and snow depth. c) Friction velocity (2 m) and water vapour pressure (1.3 m).

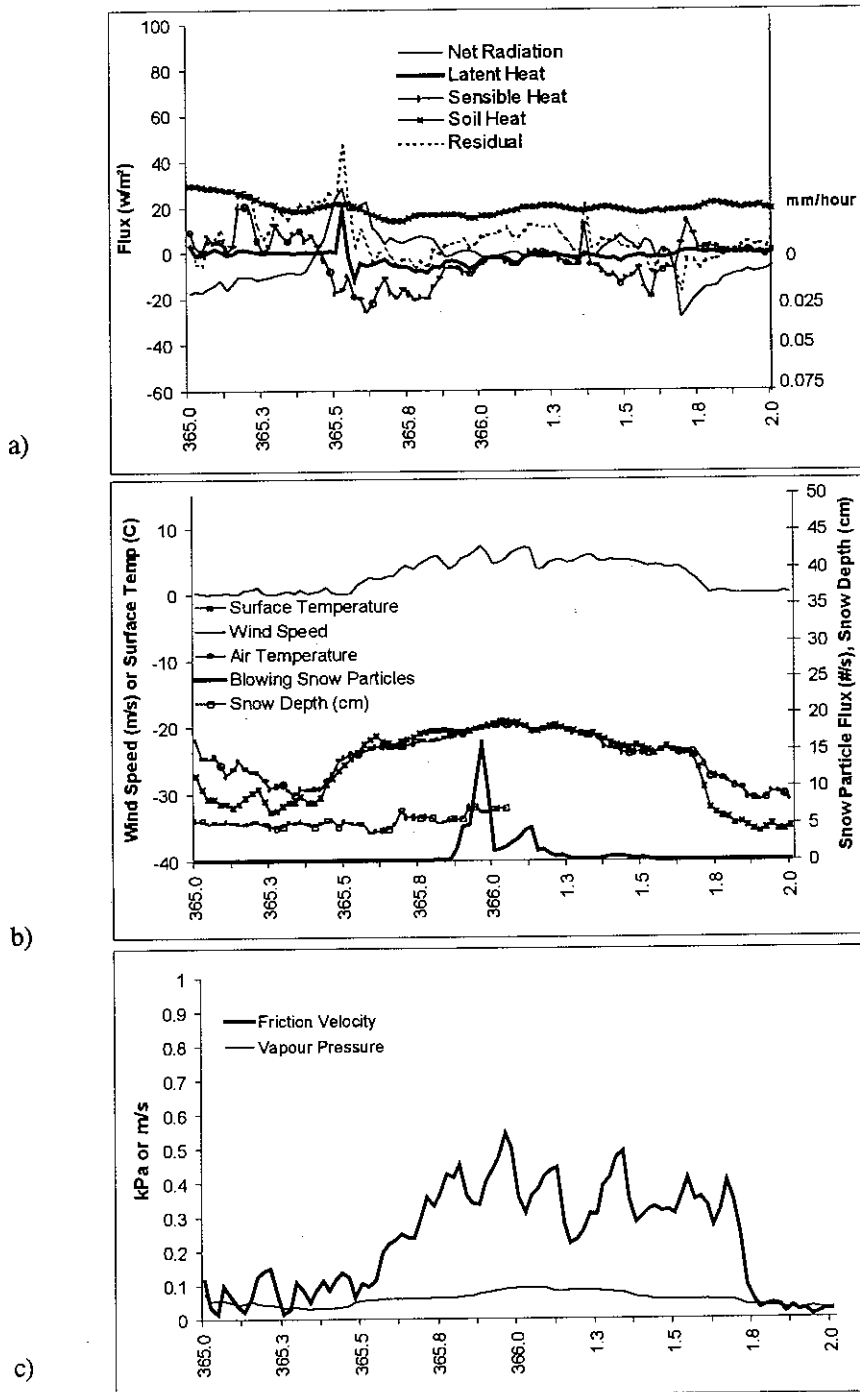


Figure 3. Measurements at Kerman Farm near Saskatoon 31 Dec. 1998 – 1 Jan. 1999.

a) Net radiation (1 m), latent heat (2 m), sensible heat (2 m), soil heat flux (-0.05 m) and residual energy ($dU/dt - \nabla \cdot Q_p - \nabla \cdot Q_a$). b) Snow surface temperature, wind speed (1.3 m), air temperature (1.3 m), blowing snow particle flux (0.2 m) and snow depth. c) Friction velocity (2-m) and water vapour pressure (1.3 m).

During the weaker blowing snow event near midnight on JD 365, air temperatures varied between -25 and -20 C and were well mixed (Fig. 3). Wind speed peaked at over 7 m/s ($u^* = 0.55$ m/s) 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 per 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 (Eq. 6) and that calculated using surface bulk transfer formulations (Eq. 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 & 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 Fig. 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 $0.038\text{-}0.076 \text{ mm hour}^{-1}$ or $30\text{-}60 \text{ W m}^{-2}$) and are strongly under-predicted by the bulk transfer calculation (typically $0.013 \text{ mm hour}^{-1}$ or 10 W m^{-2}) but better-predicted by the blowing snow model (range $0.038\text{-}0.07 \text{ mm hour}^{-1}$ or $30\text{-}55 \text{ W m}^{-2}$). As shown in Table 2, the mean error of PBSM during the first event was -1.4 W m^{-2} ($0.0018 \text{ mm hour}^{-1}$) whilst that for the surface bulk-transfer formulation was -13.0 W m^{-2} ($0.016 \text{ mm hour}^{-1}$) with correlation coefficients of 0.71 and 0.51 respectively. During the second blowing snow event, measured sublimation fluxes were smaller ($< 0.013 \text{ mm hour}^{-1}$ or 10 W m^{-2}) and were correlated better 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.

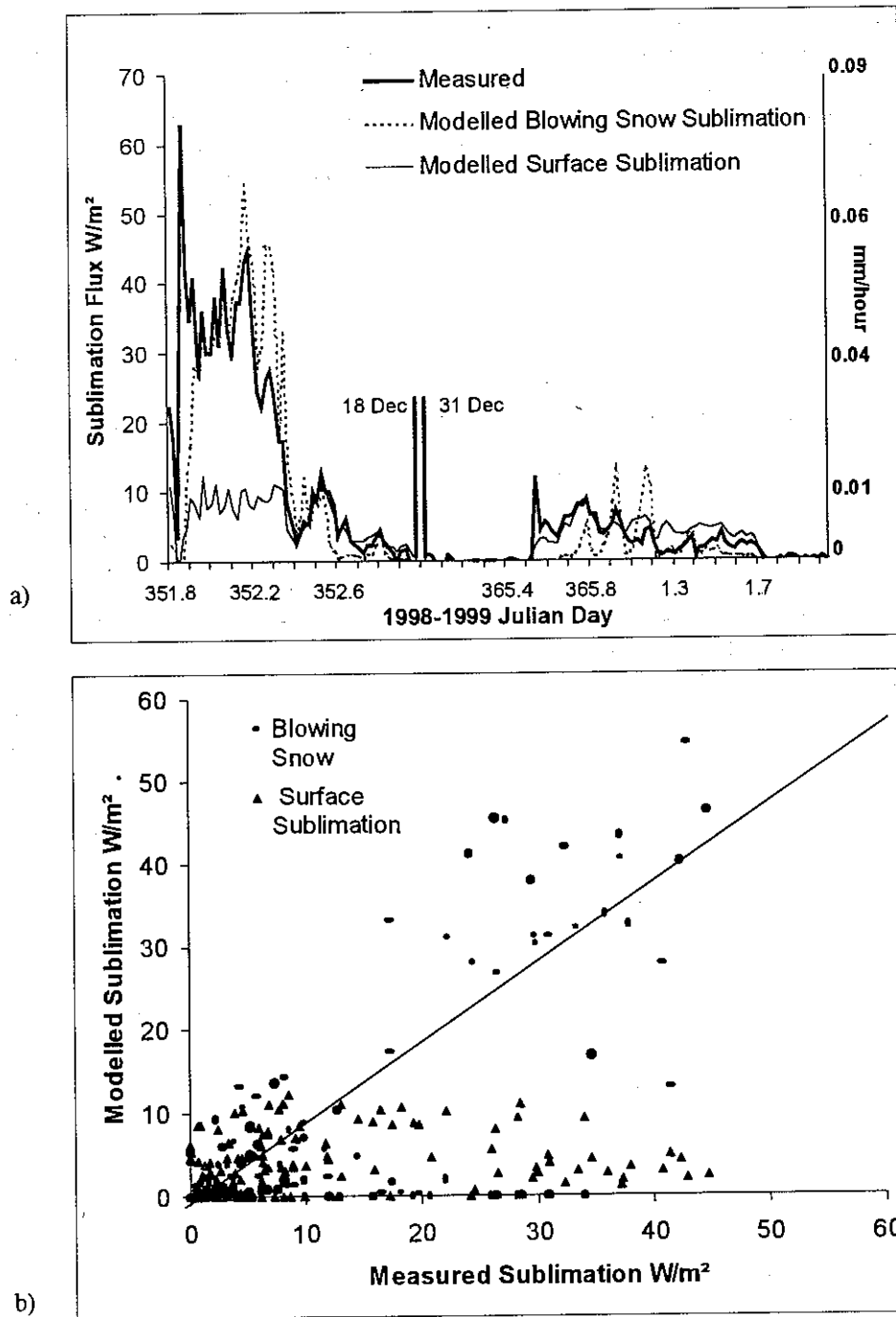


Figure 4. Measured sublimation flux, modelled blowing snow sublimation (PBSM) and modelled surface sublimation (bulk transfer). a) Time series. b) Modelled versus measured values.

IMPLICATIONS AND CONCLUSIONS

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

Table 2. 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

over melting prairie snow covers in high radiation conditions and can result in sublimation rates of nearly 2 mm snow water equivalent per 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 hour⁻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 parameterisation of multi-phase blowing snow flow.

ACKNOWLEDGEMENTS

The authors acknowledge the financial support of the Mackenzie GEWEX Programme through the Natural Sciences and Engineering Research Council of Canada and National Hydrology Research Institute, Environment Canada. Field and analytical assistance was provided by Newell Hedstrom, Dell Bayne and Brenda Toth. RE was on sabbatical from the Hadley Centre, UK Meteorological Office. Two anonymous reviewers are thanked for their thought-provoking comments which improved this paper.

REFERENCES

- Anderson, P.S. (1994) A method for rescaling humidity sensors at temperatures well below freezing. *Journal of Atmospheric and Oceanic Technology*, 11(5). 1388-1391.
- Benson, C.S. (1982) Reassessment of winter precipitation on Alaska's Arctic slope and measurements on the flux of wind-blown snow. *Geophysical Inst. Rep. UAG R-288*. Geophysical Inst., Univ. of Alaska, Fairbanks, 26 p.
- Déry, S.J. and P.A. Taylor (1996) Some aspects of the interaction of blowing snow with the atmospheric boundary layer. *Hydrological Processes*, 10. 1345-1358.
- Déry, S.J., P.A. Taylor and J. Xiao (1998) The thermodynamic effects of sublimating, blowing snow in the atmospheric boundary layer. *Boundary-layer Meteorology*, 89. 251-283.
- Dyunin, A. K. (1959) Fundamentals of the theory of snow drifting. *Izvest. Sibirsk, Otdel. Akad. Nauk. USSR* No. 12:11-24. [English translation by G. Belkov, National Research Council of Canada, Ottawa, Technical Translation 952, 1961].
- Essery, R.H.L. (1997) Seasonal snow cover and climate change in the Hadley Centre GCM. *Annals of Glaciology*, 25. 362-366.
- Essery, R.H.L., Li, L. & Pomeroy, J.W. (1999) A distributed model of blowing snow over complex terrain. *Hydrol. Processes*. In press.
- Granger, R. J. & Male, D.H. (1978) Melting of a prairie snowpack. *J. Applied Meteor.*, 17, 1833 - 1842.
- Halberstam, I. And J.P. Schieldge (1981) Anomalous behaviour of the atmospheric surface layer over a melting snowpack. *Journal of Applied Meteorology*, 20. 255-265.
- Li, L. and J.W. Pomeroy (1997a) Probability of occurrence of blowing snow. *Journal of Geophysical Research*, 102(D18). 21955-21964.
- Li, L. and J.W. Pomeroy (1997b) Estimates of threshold wind speeds for snow transport using meteorological data. *Journal of Applied Meteorology*, 36. 205-213.

- Liston, G.E. and M. Sturm. (1998) A snow-transport model for complex terrain. *Journal of Glaciology*, 44(148). 498-516.
- Male, D.H. (1980) The seasonal snowcover. In, (ed. S. Colbeck) *Dynamics of Snow and Ice Masses*. Academic Press, London. 305-395.
- Male, D.H. and R.J. Granger (1981) Snow surface energy exchange. *Water Resources Research*, 17(3). 609-627.
- McCaughey, J.H., B.D. Amiro, A.W. Robertson and D.L. Spittlehouse (1997) Forest environments. In, (eds. W.G. Bailey, T.R. Oke and W.R. Rouse) *The Surface Climates of Canada*. McGill-Queen's University Press, Montreal & Kingston. 247-276
- Oke, T.R. (1978) *Boundary Layer Climates*. Methuen and Co., London. 371 p.
- Pomeroy, J.W. (1989) A process-based model of snow drifting. *Annal. Glaciol.*, 13:237-240.
- Pomeroy, J.W. and D.M. Gray (1990) Saltation of snow. *Water Resources Research*, 26(7). 1583-1594.
- Pomeroy, J.W. and D.M. Gray (1995) *Snowcover Accumulation, Relocation and Management*. Nat. Hydrol. Res. Inst. Science Rep. No. 7, Environment Canada, Saskatoon, 144 pp. (available from the National Water Research Institute, 11 Innovation Blvd., Saskatoon, Sask., Canada)
- Pomeroy, J.W. and D.H. Male (1988) Optical properties of blowing snow. *Journal of Glaciology*, 34(116). 3-10.
- Pomeroy, J.W. and D.H. Male (1992) Steady state suspension of snow. *Journal of Hydrology*, 136. 275-301.
- Pomeroy, J.W. & Li, L. (1997) Development of the Prairie Blowing Snow Model for application in climatological and hydrological models. *Proc. Eastern Snow Conf.*, 54. 186-197.
- Pomeroy, J.W., D.M. Gray and P.G. Landine (1993) The Prairie Blowing Snow Model: characteristics, validation, operation. *Journal of Hydrology*, 144. 165-192.
- Pomeroy, J.W., Marsh, P. & Gray, D.M. (1997) Application of a distributed blowing snow model to the Arctic. *Hydrol. Processes*, 11, 1451-1464.
- Schmidt, R.A. (1972) Sublimation of wind-transported snow - a model. *Res. Pap. RM-90*, USDA For. Serv., Rocky Mtn. For. & Range Exp. Sta., Fort Collins, USA.
- Shook, K. and D.M. Gray (1997) The role of advection in melting shallow snowcovers. *Hydrological Processes*, 11. 1725-1775.
- Shuttleworth, W.J. (1989) Micrometeorology of temperate and tropical forest. *Philosophical Transactions of the Royal Society (London) B*, 324. 299-334.
- Stull, R.B. (1988) *An Introduction to Boundary Layer Meteorology*. Kluwer Academic Publishers, Boston.
- Thorpe, A.D. and B.J. Mason (1966) The evaporation of ice spheres and ice crystals. *Journal of Applied Physics*, 17, 541-548.
- Verseghy, D.L., McFarlane, N.A. & Lazare, M. (1993) CLASS - a Canadian land surface scheme for GCMs, II, vegetation model and coupled runs. *Int. J. Climatol.*, 13. 347-370.
- Yamanouchi, T. and S. Kawaguchi (1985) Effects of drifting snow on surface radiation budget in the katabatic wind zone, Antarctica. *Annals of Glaciology*, 6. 238-241.