DEVELOPMENT OF THE PRAIRIE BLOWING SNOW MODEL FOR APPLICATION IN CLIMATOLOGICAL AND HYDROLOGICAL MODELS

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ABSTRACT

The Prairie Blowing Snow Model (PBSM) is a single column, physically-based, mass and energy balance that calculates blowing snow transport and sublimation rates. New features of the model include a correction for snowfall undermeasurement, mid-winter snowmelt, snow density estimates and estimation of the effect of exposed vegetation on aerodynamic roughness height during blowing snow. Importantly, the model now includes algorithms to estimate the threshold wind speed for snow transport and to estimate the probability of blowing snow occurrence. The probability of blowing snow occurrence is used to scale the blowing snow fluxes horizontally so that estimates from a 1 x 1 km surface area control volume may be extrapolated to larger scales, varying fetches and vegetation types. As a demonstration, the model is operated using standard meteorological data from meteorological stations in the Canadian Prairies and compared to snow accumulation measurements at these stations. The model is also used to estimate snow accumulation for comparison to extensive field snow surveys from the Bad Lake experimental basin in west-central Saskatchewan. It is shown that the new version of PBSM can provide estimates of snow accumulation, melt, transport and sublimation from hourly to seasonal time periods for varying land surfaces.

INTRODUCTION

Redistribution of snow by wind, termed *blowing snow,* is ubiquitous in windswept, exposed landscapes. Where snow is a large component of annual precipitation, blowing snow processes can be critical in determining the water balance at both small and large scales. Wind redistribution is normally accompanied by in-transit sublimation, with substantive losses to the snowcover of exposed sites resulting from erosion and subsequent relocation or sublimation (Dyunin, 1959; Tabler, 1975; Tabler et al., 1990; Benson and Sturm, 1993; Pomeroy et al., 1993; 1997).

The physics of snow transport and sublimation involves phase change, two-phase flow and rapid energy and mass transfers in the atmospheric boundary layer just above the snowpack. The Prairie Blowing Snow Model (PBSM) was first developed in 1988 as a single column mass and energy balance that calculates blowing snow transport and sublimation rates (Pomeroy, 1989). The model is composed of physically-based algorithms developed with the aid of an extensive observational study (Pomeroy and Gray, 1990; Pomeroy and Male, 1992). PBSM was later extended to include a snowcover mass balance for determining seasonal snow accumulation, transport and sublimation for the case of two-dimensions with varying agricultural land use, snowdepth and fetch (Pomeroy et al., 1993). The mass balance PBSM was successfully field-validated at small and large scales. These early versions of PBSM were dependent upon manual observations of blowing snow events as recorded at weather stations to determine periods for which blowing snow was occurring and to determine the threshold shear stress for saltation calculations. To overcome the restrictive data requirements, a climatological version of PBSM was developed from the hourly outputs of the mass balance version. Monthly sums of transport and sublimation for 1 km fetches of fallow and stubble land use were compared to monthly mean climate variables in order to develop simple predictive equations using meteorological conditions in the Canadian Prairies (Pomeroy and Gray, 1994). The climatological version can be easily applied using monthly means of wind speed, air temperature, humidity, snowfall and depth of snow on the ground, and has been distributed and validated in the Arctic, outside of its region of development (Pomeroy et al., 1997).

Recent developments in the mass balance version of PBSM have permitted direct calculation of the threshold condition for transport from the meteorological history of the snowpack (Li and Pomeroy, 1997a) and the probability of blowing snow occurrence (Li and Pomeroy, 1997b). These additions permit use of PBSM with standard meteorological datasets and provide a simple method for scaling blowing snow fluxes from a point to a larger uniform area. It is the purpose of this paper to outline new developments in PBSM, to demonstrate the use of PBSM for estimating snow redistribution

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and sublimation using standard meteorological records and to test PBSM at different scales using mass balances of snow water equivalent on the ground and snowfall measurements.

NEW DEVELOPMENTS IN PBSM

Snowfall Corrections

Snowfall undermeasurement due to wind effects, wetting losses and unrecorded trace events has been the subject of extensive investigation with corrections recently published. For Canadian measurements Nipher-shielded cylinders are often used to collect snowfall with measurements of accumulated snowfall in the cylinder made every six hours. The Nipher shield reduces undermeasurement due to wind compared to that of an unshielded gauge, but undermeasurements still occur. A correction procedure published by Goodison et al. (1997) has been incorporated in the PBSM and is used where uncorrected precipitation records must be used. The upward revision of annual snowfall is the order of 31% in the southern Canadian Prairies and 64% to 161% in the High Arctic (Pomeroy and Goodison, 1997)

Snowmelt Estimation

For validation and sequential operation through a season the PBSM must determine the remaining snowcover on the ground. At this stage two simple, provisional snowmelt routines have been incorporated to determine snow water equivalent losses from mid-winter melts. These melt events are normally driven by large scale advection because low sun angles and short days reduce solar radiation inputs (Shook, 1995). Where snow depth measurements are unavailable, a degree-day algorithm is used. Where the PBSM is integrated with other hydrological or meteorological models an energy balance routine such as that proposed by Gray and Landine (1988) is recommended.

Threshold Wind Speed for Transport

Li and Pomeroy (1997a) examined threshold (initiation) conditions for blowing snow transport using a dataset created from six years of hourly observations at 16 meteorological stations in the Canadian Prairies. They found an average threshold wind speed (all wind speeds listed are at the 10-m height) for wet snow of 9.9 m s^{-1} with a range from 7 to 14 m s^{-1}. Wet snow was defined as that which had received above-freezing temperatures or rainfall since the last snowfall. The range of values for wet snow was not related to any other meteorological variable. Transport thresholds for dry snow varied from 4 to 11 m s^{-1} and varied with air temperature, decreasing with decreasing air temperature down to -25 C and then slightly increasing with further decreasing air temperature. The decrease with decreasing air temperature is most rapid near the melting point. The relationship developed by Li and Pomeroy to describe the threshold wind speed at 10-m height, $U_t(10)$, is,

$$U_t(10) = 9.43 + 0.18\,T + 0.0033\,T^2 \tag{1}$$

where T is air temperature measured at 2-m height in degrees C. The variation in mean observed and modelled threshold wind speed with air temperature is shown in Fig. 1 for both fresh and aged snow. Whilst average values for aged snow are slightly higher than for fresh snow the difference does not merit separate mathematical treatment of the two conditions. A transport threshold algorithm has been incorporated in PBSM that uses Eq. 1 for dry snow and $U_t(10)$ = 9.9 m s^{-1} for wet snow. With appropriate roughness heights $U_t(10)$ is used to calculate the threshold shear stress for saltation flux calculations.

Steady-state Boundary Layer

Energy and mass transfer during blowing snow affects the atmospheric boundary layer, removing heat and increasing water vapour via turbulent transfer from blowing snow particles. Several studies have modelled a decrease in water vapour deficit and air temperature during blowing snow or along an increasing fetch of uniform blowing snow transport (Mobbs and Dover, 1993, Dery and Taylor, 1997). PBSM does not explicitly model boundary layer energy and water vapour balances but uses *measured* values of air temperature and relative humidity at locations with large open fetches (airports, meteorological stations on the Prairies) where approximately steady-state boundary layer conditions have developed. To adjust the measured temperature and humidity for heights other than that of measurement (2 m), profiles of air temperature and water vapour deficit measured during an extensive blowing snow measurement programme at Loreburn, Saskatchewan are used (Pomeroy and Male, 1987). Loreburn is located in an open grain growing region with a fetch for 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, hence in PBSM temperature is retained at the 2-m value. Water vapour deficit during strong blowing snow did show a consistent vertical profile at Loreburn, with undersaturation becoming smaller as height above the snow surface decreases as shown in Fig. 2. The best fit profile is,

$$\sigma(z) = \sigma(2)\,(1.019 + 0.027 \ln z) \qquad \textbf{(2)}$$

where σ is the undersaturation of water vapour and z is height above the snow surface (m). Equation 2 is also plotted in Fig. 2. This equation is a correction to Eq. 61 listed by Pomeroy and Gray (1995). Note that saturated conditions are not normally reached in most of the blowing snow boundary layer for typical conditions during blowing snow on the Canadian Prairies (65% > relative humidity < 95%), in contrast to the results of the modellers (Mobbs and Dover, 1993; Dery and Taylor, 1997) who do not consider large scale advection during blowing snow and make a variety of untested assumptions in arriving at their blowing snow boundary-layer models..

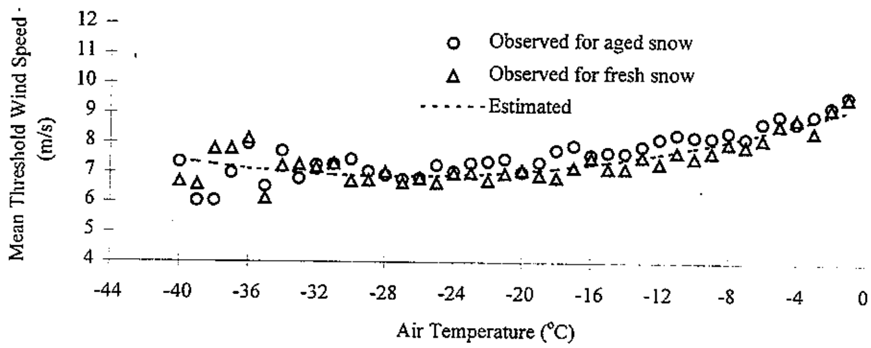

Fig. 1. Measured and modelled mean threshold wind speeds for blowing snow transport.

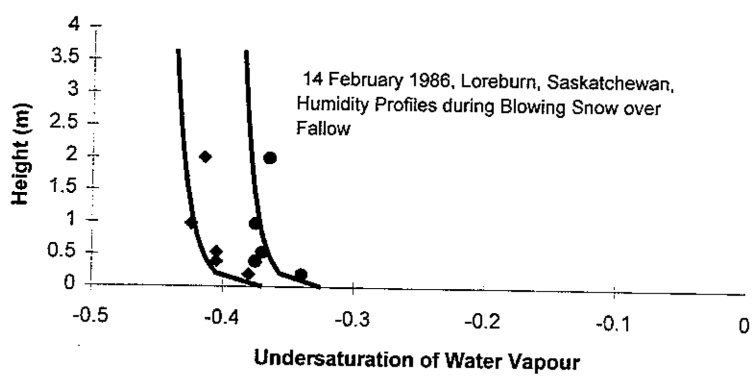

Figure 2. Vertical profiles of undersaturation of water vapour measured at Loreburn, Sask. during strong blowing snow, and modelled profiles.

<u>Scaling Fluxes from Point to Area.</u>

"Scaling-up" Single Column Calculation of Fluxes. Previous versions of PBSM have been distributed with a spatial resolution corresponding to land units of 100 m or smaller (Pomeroy et al., 1993; 1997). Explicit blowing snow flux and snow mass balance calculations for each land unit were made given its vegetation cover, fetch characteristics and blowing snow fluxes entering from adjacent land units. This approach has provided useful information on the variation in blowing snow fluxes with fetch and land use but is computationally intensive and inappropriate for many modelling applications. To simplify the model computations, PBSM here is run in a single column for a 1 km uniform fetch and

scaled-up to provide fluxes first for a 1x1 km uniform area, then for other areas of differing vegetation cover and fetch length.

Uniform Areas. The PBSM application in this paper calculates blowing snow fluxes at a point for mean hourly meteorological conditions presuming a 1 km fetch. However observations of the phenomenon show that it is extremely unsteady over space and time, in that over a uniform field of snow, 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. The occurrence of intermittent time series of fluxes and patches of snow transport over fields is likely associated with small-scale variation in snowcover properties and boundary-layer flow that are not addressed by the uniform upwind fetch and constant hourly wind speed assumptions of PBSM. To develop a scaling function for extrapolating from steady conditions at a point to "true" conditions over a relatively uniform area, the probability of blowing snow occurrence was examined by Li and Pomeroy (1997b). In examining the same dataset used to determine threshold wind speeds they found that the probability of occurrence of blowing snow with respect to wind speed approximates a cumulative normal distribution, described by the mean and variance of wind speed as,

$$P = \frac{1}{\delta\sqrt{2\pi}} \int_0^{U_i} e^{-\frac{(\bar{U} - U_i)^2}{2\delta^2}} dU \qquad (3)$$

where P is the probability of blowing snow occurrence, U_i is the current wind speed (10-m), and $\bar{U}$ is the mean and δ^2 is the variance of wind speed (location and scale parameters of normal distribution). From the six year dataset Li and Pomeroy found empirical descriptions for the location and scale parameters as functions of wet snow conditions, air temperature and snow age. For wet or icy snowpacks the mean and variance of wind speed were found to be 21 and 49 m s^{-1} respectively. For dry snow the mean wind speed was found to fit,

$$\bar{U} = 0.365T + 0.00706T^2 + 0.9I + 11.2 \qquad (4)$$

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 found to fit

$$\delta = 0.145T + 0.00196T^2 + 4.3 \qquad (5)$$

The resulting blowing snow occurrence probability as a function of wind speed for various temperatures is shown in Fig. 3 along with observed probabilities for I=2. While the occurrence probability was developed from a time series, it can be applied to space if the area of application is uniform in its mean characteristics. The occurrence probability is used by PBSM as a scaling parameter for weighting calculated steady-state fluxes over uniform space, e.g. a probability of occurrence of 70% would result in fluxes being multiplied by 0.7 in scaling from point to large uniform area. Values of $\bar{U}$ less than 7 m s^{-1} for wet snow and 4 m s^{-1} for dry snow are assigned P=0 as are conditions with no snowcover. If the current wind speed has P>0 and $U_i<U_t(10)$ then $U_t(10)$ is recalculated as equal to U_i-0.5.

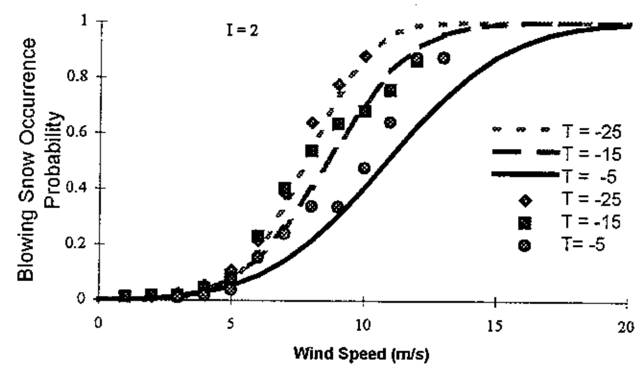

Figure 3. Blowing snow occurrence probability as a function of wind speed and air temperature for a snowpack 7 hours old: model and measurements

Areas of Varying Vegetation Cover. The probability-scaling method shown above presumes snowcovers with no exposed vegetation. To calculate the probability of blowing snow occurrence for terrain with vegetation exposed above the snowpack the aerodynamic effects of exposed vegetation must be taken into account in the calculation of a probability distribution. Raupach et al. (1993) proposed calculation schemes for relating the geometry of exposed vegetation to partitioning of shear stress at the surface, where the ratio of the shear stress applied to exposed roughness elements, τ_n, to the total shear stress, τ, is found as

$$\frac{\tau_n}{\tau} = \frac{\beta \lambda}{1 + \beta \lambda} \tag{6}$$

where β is the ratio of element to surface drag and λ is the roughness density found as the number density of exposed roughness elements (# m^{-2}) multiplied by the exposed silhouette area of the roughness elements. Raupach et al. suggest that β is approximately 170.

Equation 6 can be used to find wind speed associated with the shear stress applied to the snow surface under exposed vegetation, U_{iS}, as,

$$U_{iS} = \frac{u_*\left(1 - \sqrt{\frac{\tau_n}{\tau}}\right)}{k} \ln\left(\frac{z}{\frac{u_*^2}{163.3} + 0.5\lambda}\right) \tag{7}$$

where u^* is the shear stress (m s^{-1}), k is von Kármán's constant (0.4) and z is the height of U_{iS} (m). The roughness height is approximated using a relationship by Pomeroy and Gray (1990) for blowing snow roughness ($u^{*2}/163.3$) added to Lettau's (1969) vegetation roughness parameterisation, 0.5λ. The wind speed from Eq. 7 is used as U_i in Eq. 3 to scale the blowing snow probability for terrain where vegetation is exposed. An example is shown in Fig. 4 for an air temperature of -15 C, I=2, exposed wheat stubble vegetation (λ=0.96*stubble height) 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 25 m s^{-1} wind speed the probability of blowing snow declines from 100% to 30% as the exposed stubble height increases from 0 to 0.01 m. This modelled result is in concurrence with observations in Saskatchewan that wheat stubble tends to fill with wind-blown snow to approximately the height of stubble stalks (Pomeroy and Gray, 1995).

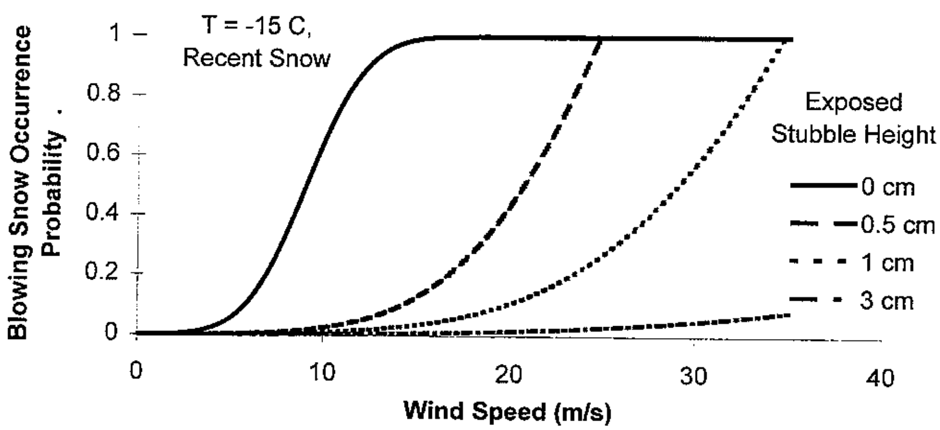

Figure 4. Probability of blowing snow over snowpacks with exposed cereal stubble stalks.

Areas of Varying Fetch. PBSM fluxes vary with fetch (Pomeroy et al., 1993; Pomeroy and Gray, 1995); the amount of snow eroded and sublimated versus that eroded and transported to the end of the fetch increases with increasing fetch distance. It is not appropriate to simply set the control volume of a single column version of PBSM equal to large fetches without mass balance controls on surface boundary conditions as the application by Pomeroy et al. (1993). They found that the erosion flux does not vary substantially with increasing fetch beyond 1 km and that vertical fluxes of blowing snow sublimation and snow erosion and subsequent transport to the end of the fetch follow characteristic seasonal patterns controlled by the surface snowpack mass balance. Pomeroy et al. (1997) described equations that parameterise the variation in the vertical fluxes of sublimation and erosion to transport with fetch distance with respect to the 1-km fetch fluxes. The application of these equations is shown in Fig. 5, where Q_T is the downwind transport rate (kg m^{-1} s^{-1}), x is an incremental downwind distance, Q_E is the unit area sublimation flux (kg m^{-2} s^{-1}) and all fluxes are normalised to the values calculated from the single column PBSM with fetch set to 1-km.

Distribution over Landscapes

In distributing PBSM the landscape is segregated into "sources" and "sinks" of blowing snow (Pomeroy et al., 1997) with the following mass balance in each landscape type specifying the interaction between source and sink areas,

$$Q_A(F) = Q_P - \frac{dQ_T}{dF}(F) - Q_E(F) - Q_M \tag{8}$$

where Q_A is the snow accumulation flux (kg m^{-2} s^{-1}), Q_M is the snow melt flux (kg m^{-2} s^{-1}) and F is the fetch distance (m). Note that Q_T entering a fetch (F=0) is equal to the Q_T leaving the previous fetch at the maximum fetch distance of the previous landscape type. To aggregate the fluxes for each landscape type up to larger average values the fluxes are weighted by the respective areas of the landscape types.

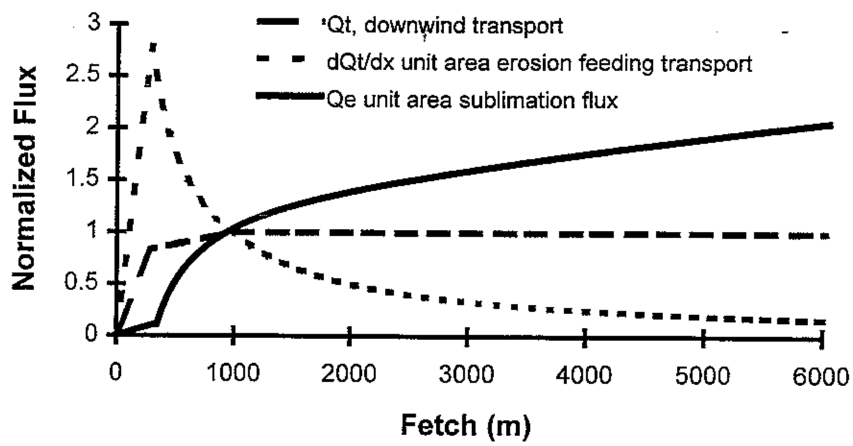

Figure 5. Change in blowing snow transport and sublimation fluxes with increasing fetch.

APPLICATION AND EVALUATION OF PBSM ROUTINES

Simulation for Climate Stations

As a first application, the new PBSM routines are run using standard datasets from Atmospheric Environment Service climate stations in the Canadian Prairies: hourly wind speed, air temperature, relative humidity, presence of snowcover, precipitation type and six-hourly snowfall. The winter seasons of 1973-74 and 1974-75 were chosen for application because fetch characteristics of the stations are excellent for those years, data quality is good and the years represent average (1974-75) and high (1973-74) annual snowfall amounts. The model was run using snowfall corrections and revisions as described above for 1-km fetches of uniform short grass terrain, with melt estimated from decreases in snowdepth during above freezing temperatures. The model is uncalibrated in that measured snow depths are not used to adjust the calculated snowcover mass balance during or after blowing snow events. The modelled snow accumulation was compared to estimates of snow water equivalent (SWE) made from measured snow depth at the stations (six-hourly) converted to SWE using prairie snow density equations developed from extensive measurements by Shook and Gray (1994). Simulations and SWE estimated from snowdepth for Calgary, Alberta; Regina and Prince Albert, Saskatchewan; and Portage la Prairie, Manitoba are shown in Fig. 6. The stations represent northern prairie snow climates ranging from chinook-prone western foothills (Calgary), to windy central plains (Regina), northern boreal forest transition (Prince Albert) and eastern deciduous transition (Portage la Prairie). Table 1 shows the annual snowfall, transport and sublimation for each station. It is evident from these data that:

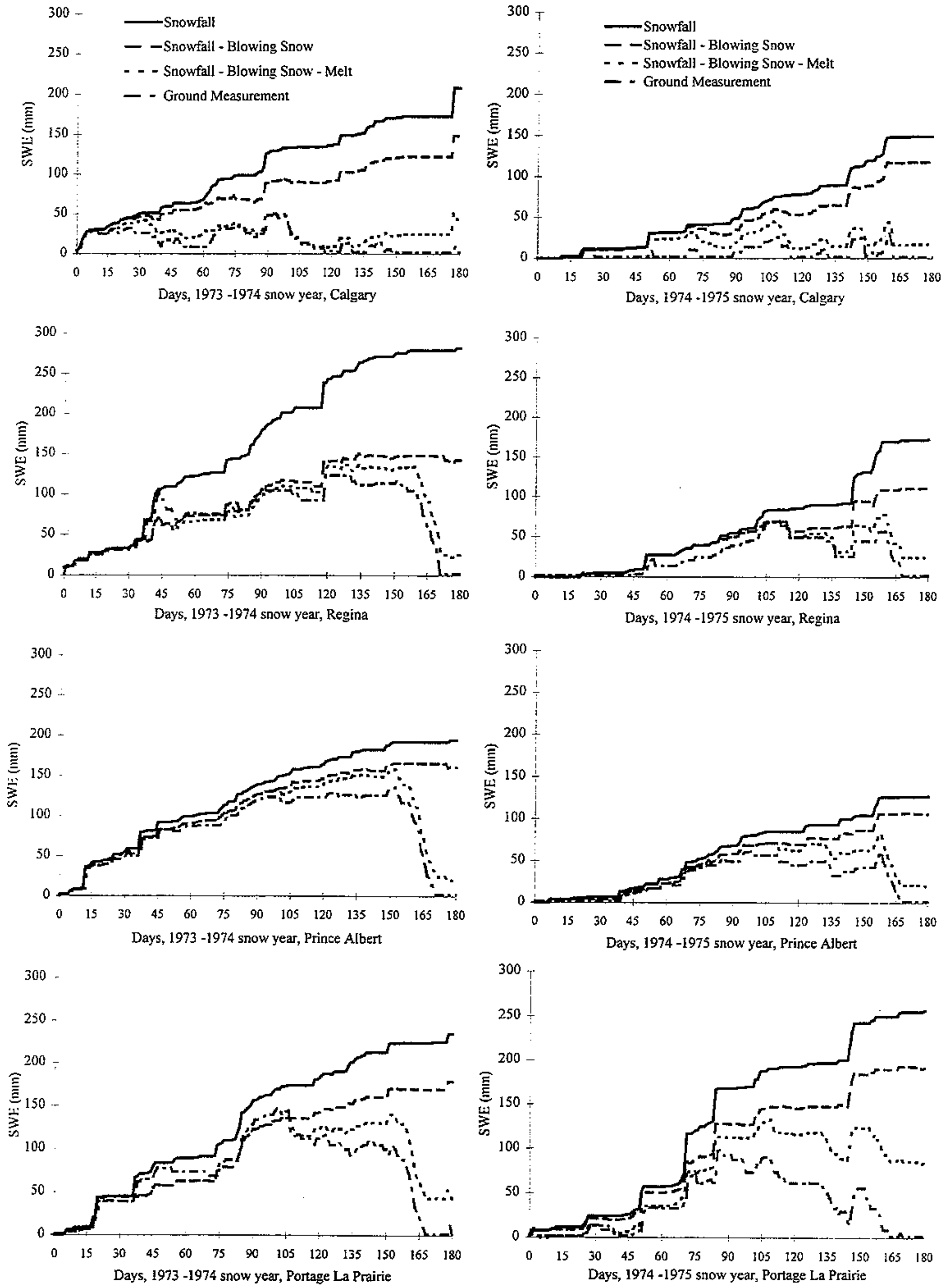

Figure 6 Simulated blowing snow fluxes, snow melt and surface snow mass balance and estimated snowfall and snow water equivalent for Climate Stations in the Prairie Provinces of Canada. Model simulations assumed fetch=1 km with short vegetation.

a) sublimation and transport ablate 12%-33% and 3%-16% respectively of annual snowfall over the winter,

b) the temporal pattern of snow accumulation is modelled correctly by PBSM though there are cumulative errors in the seasonal accumulation of varying magnitude, and

c) calculations of ablation due to mid-winter melt alone would lead to substantial errors (16%-49% of snowfall) in estimated SWE at the end of the winter .

The model works relatively better where wind speeds are highest (Regina, Calgary) and snowfall amounts are largest (1973-74), however correspondence of the modelled mass balance to the estimated SWE should not be taken as strict confirmation or rejection of model performance because of potential errors in the estimated SWE due to density assumptions and the location of snow survey lines at the climate stations.

Table 1. Seasonal modelled snow fluxes and snowfall for climate stations, mm SWE/year.

Station	Snowfall	Melt	Transport	Sublimation
Calgary 1973-74	208	105	12	48
Calgary 1974-75	149	101	4	27
Regina, 1973-74	282	117	46	93
Regina, 1974-75	172	86	16	44
Prince Albert, 1973-74	194	141	9	24
Prince Albert, 1974-75	128	85	5	16
Portage la Prairie, 1973-74	235	136	20	37
Portage la Prairie, 1974-75	257	107	24	41

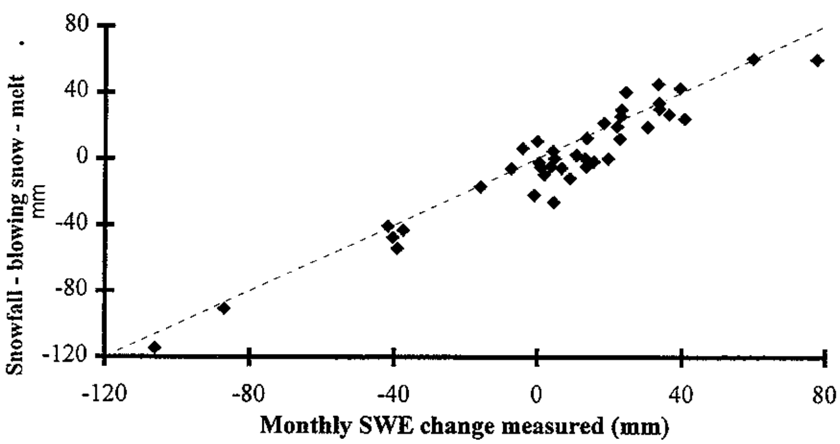

Figure 7. Monthly fluxes of snowcover mass balance by PBSM and changes in estimated snow water equivalent for 4 prairie climate stations over 2 years.

Because errors in accumulative blowing snow fluxes and snowcover mass balances become magnified as the seasonal simulation progresses, monthly fluxes of Q_A(1 km) and the monthly change in estimated SWE are shown in Fig. 7. The comparison follows the 1:1 correspondence line fairly well, with a correlation coefficient of 0.96 and mean overestimate of accumulation of 5.9 mm SWE per month. Such correspondence between ground measurements and modelled ablation for an uncalibrated run with an empirical melt routine is quite promising.

Simulation for the Bad Lake Basin

Bad Lake in west central Saskatchewan was the site of an International Hydrological Decade research basin that operated from 1967 through 1986. In the mid-1970's extensive snow surveys were conducted in the basin and a well-serviced meteorological station was operated (Gray and Granger, 1988). The basin is primarily covered with cereal crops and pasture with a tableland dissected by sharp gullies vegetated with bushes. Snow surveys in the basin and basin land cover classes have been described by Steppuhn and Dyck (1974), Steppuhn (1976) and Shook (1995).

PBSM was used to calculate snow accumulation, sublimation and transport fluxes in source (fallow and stubble fields) and sink (valley sides covered with shrubs) areas of the Creighton Tributary coulee of Bad Lake basin for the years 1973-74 and 1974-75 when snow surveys were made to evaluate model results. The degree-day melt routine was used as daily snow depths were not measured. Over the two years, annual blowing snow sublimation and transport were respectively 17-24% and 10-15% of snowfall for fallow and stubble plateau landscapes with mean fetches of roughly 1 km. Two large area snow surveys were conducted in the spring before the seasonal melt on summer fallow (no vegetation) and cereal stubble (stubble height 0.25 m) landscapes of Creighton Tributary. The annual simulation of blowing snow fluxes and the areal average of the measured SWE are shown in Fig. 8 and show a remarkably good correspondence with a cumulative error of 9 mm SWE or 3.5% of annual snowfall and a tendency for the model to overestimate accumulation somewhat.

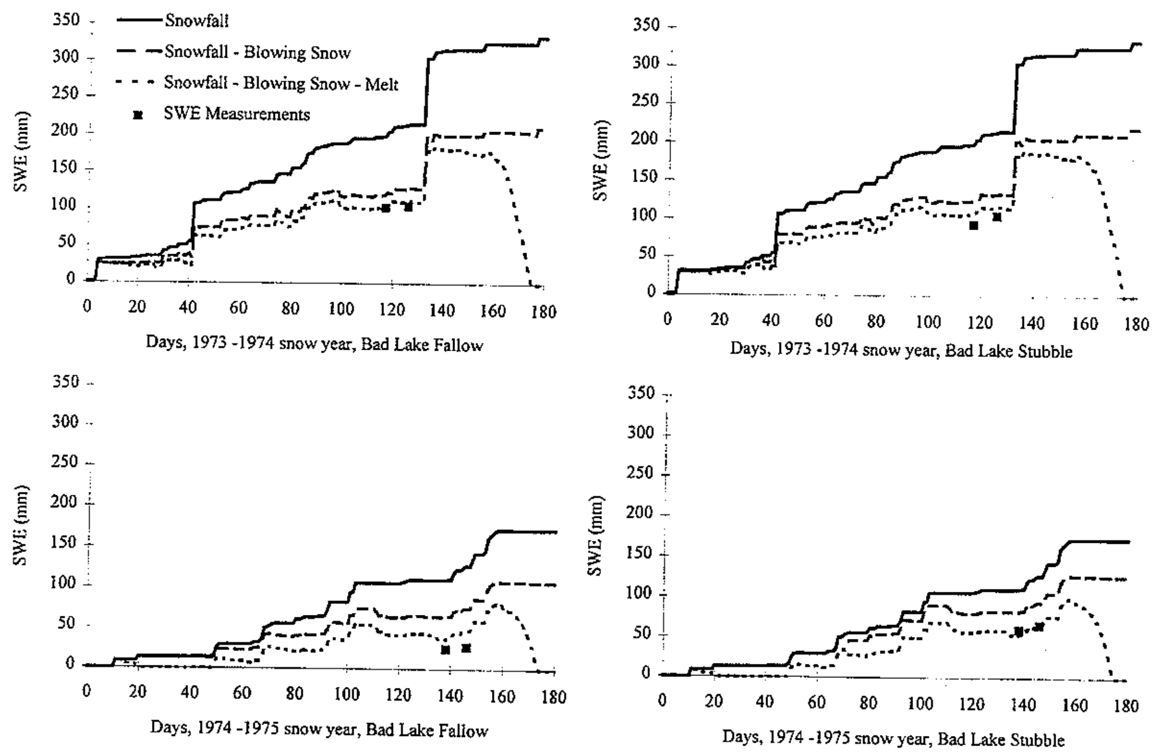

Figure 8. Seasonal simulation of blowing snow fluxes and snowcover mass balance along with measured snowfall and area measurements of snow water equivalent for fallow and stubble field source areas, Creighton Tributary, Sask.

Measurements and simulations of snow accumulation in sink areas provide a test of the simulation of accumulation of blowing snow to sinks and the estimation of transport from source areas. If good agreement is reached for measured and modelled snow accumulation for both source and sink areas then individual components of the mass balance that do not contribute to relocation such as sublimation can be indirectly confirmed. Snow accumulation in 16 sink areas (shrub-covered valley sides or E5 following Steppuhn's (1976) classification) of Creighton Tributary was surveyed extensively in February and March of 1974. Fetches of the sink areas varied from 115 to 230 m and transport out of the sinks was presumed to be negligible. Fetches of source areas adjacent to the sinks were approximately 1 km, with land use varying from complete stubble to complete fallow. A comparison of modelled accumulation and measured SWE in the sinks from the two sets of surveys is shown in Fig. 9. A mean difference (measured - modelled) of 26 mm SWE was found, indicating the model underestimated accumulation by about 6%. However, examination of Fig. 9 shows the model overestimates smaller measured SWE and overestimates larger measured SWE by up to 100 mm. This overestimation may be due to certain areas of the snow sink landscape acting as incompletely-efficient traps or to relocation between snow "traps" during this very high snowfall season. Despite the variation with individual points, the average difference between modelled and measured accumulation for source and sink landscapes is quite promising and suggests that the model can capture the essential features of blowing snow erosion, sublimation, transport and relocation in an irregular prairie environment.

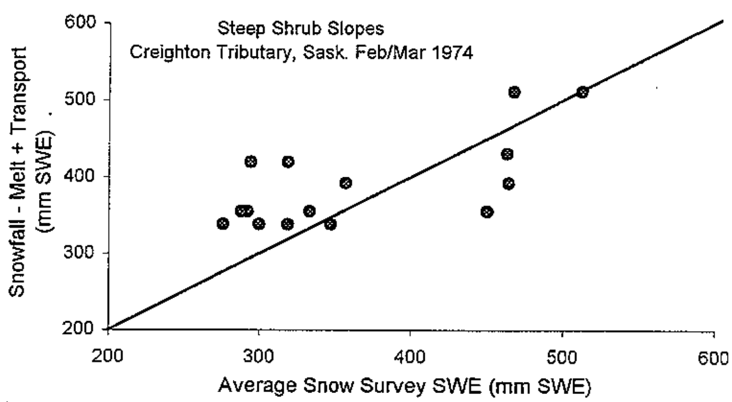

Figure 9 Estimated and measured accumulation for steep shrub-covered slopes, Creighton Tributary, Sask.

CONCLUSIONS

Further development of the PBSM has provided a physically-based model that can use standard meteorological data to estimate the timing and magnitude of blowing snow transport and sublimation fluxes and calculate a mass balance for snowpacks in windswept regions. A new statistical scaling technique permits calculation of areal fluxes from a single-column simulation of blowing snow and extrapolate the fluxes to landscapes of varying fetch and vegetation cover. Demonstration of the model in the prairies of western Canada shows that the model can simulate measured snowcovers at AES climate stations quite well over a season with monthly mean errors within 6 mm SWE/month. The model was evaluated in calculating snow relocation from and accumulation to snow source and sink landscapes in a windswept prairie basin by comparing calculations to areal snow surveys of each landscape type. In source areas differences in modelled and measured accumulation ranged up to 20 mm, averaging at 9 mm SWE or 3.5% of annual snowfall. In sink areas differences ranged up to 100 mm averaging at 26 mm SWE or an underestimation of accumulation of about 6%. The good agreement shown suggests that the model represents the essential features of snowcover development in windswept regions and can be used to provide simulations of blowing snow redistribution, sublimation and accumulation at a variety of scales and levels of resolution.

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