Chapter 14

Recent Studies on the Climatology and Modeling of Blowing Snow in the Mackenzie River Basin

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Abstract This chapter presents a multi-scale analysis of the contribution of blowing snow to the hydrometeorology of the Mackenzie River Basin (MRB). A climatology of adverse wintertime weather events demonstrates that blowing snow events are rare within the forested sections of the MRB but become more frequent in the northern parts of the Basin covered by tundra, which experience the largest impacts of blowing snow transport and sublimation due to large-scale processes. A parameterization for blowing snow sublimation based on the PIEKTUK-D model and the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA-15) data is used to determine that the combined processes of surface and blowing snow sublimation deplete 29 mm yr^{-1} snow water equivalent, or about 7% of the watershed's annual precipitation. This study provides only a first-order estimate of the contribution of surface sublimation and blowing snow to the MRB surface mass balance because of limitations with the dataset and some uncertainties in the blowing snow process.

1 Introduction

Canada is renowned for its long, frigid winters which promote the accumulation of snow that is ubiquitous and even affects the lifestyle during the cold season. Despite their favorable aesthetic and cultural aspects, winter and snow also have negative impacts on Canadians. For instance, intense cold and associated high windchills remain Canada's most devastating natural hazard as about 100 people perish annually from exposure to extreme coldness (Phillips 1990). Severe winter weather such as blizzards, snowdrifting, or freezing rain storms can seriously disrupt the daily activities of millions of Canadians (Kind 1981). A spectacular example is the 1998 'Ice Storm' in Eastern Canada that caused over a billion dollars of damage and was responsible for the deaths of dozens of people (Szeto et al. 1999).

Despite their significant societal impacts, processes responsible for the formation and evolution of winter storms and the seasonal snowpack are not fully understood. One region in Canada subject to much ongoing investigation is the Mackenzie River Basin (MRB) which is snow-covered for a substantial part of the year (up to 250 days on the Arctic tundra according to Phillips 1990). Some of the processes influencing surface mass balance of the Basin are potentially linked to the transport of wind-driven snow during blizzards and other high wind events (Stewart et al. 1998). Drifting and blowing snow occur when winds surpass a certain threshold value and then erode snow from exposed surfaces to relocate it to sheltered zones such as vegetated areas or depressions. A secondary process during advection is sublimation (phase change of ice to water vapor), providing an additional source of moisture while acting as a sink of sensible heat to the atmospheric boundary layer (ABL) (Déry and Taylor 1996).

Questions remain regarding the role played by blowing snow in the surface mass balance of high-latitude regions (Lawford 1994). This chapter presents a multi-scale analysis of the contribution of blowing snow to the surface mass balance of the MRB. For completeness, the contribution of surface sublimation from the snowpack is also investigated.

2 Background

2.1 Surface Mass Balance

The annual surface mass balance for a nival regime may be expressed as (King et al. 1996):

$$S = P - E - M - D - Q_s \tag{1}$$

where S is the change in storage or accumulation of snow at the surface, P is the precipitation rate, E is the evaporation rate that includes the surface sublimation rate Q_{surf}, and M is the divergence of water after melt and runoff. The two terms associated with blowing snow are represented by D and Q_s, the horizontal divergence and sublimation of blowing snow, respectively. Note that E, Q_{surf}, and Q_s are all defined here as positive quantities when representing an upward flux of water vapor from the surface to the atmosphere. All terms in Eq. (1) are expressed in units of mm yr^{-1} snow water equivalent (SWE) unless stated otherwise.

For the MRB, precipitation and evaporation estimates vary considerably, in part due to the lack of observations in the basin, most notably in the Mackenzie Mountains. Current estimates have the annual ($P - E$) averaged for the MRB to be at 180 mm yr^{-1} (Stewart et al. 1998). Although much attention has been given to the influx and efflux of atmospheric water vapor in the MRB domain (Bjornsson et al. 1995; Lackmann and Gyakum 1996; Lackmann et al. 1998; Misra et al. 2000; Smirnov and Moore 1999), none of these authors attempted to evaluate the impact of blowing snow on the surface mass balance of the MRB as a whole. Walsh et al. (1994) and Betts and Viterbo (2000) examined the role of surface sublimation in the MRB, but they came to contradictory conclusions, with the former finding significant deposition during wintertime, and the latter strong sublimation. The lack of accurate (if any) information on the role of blowing snow and surface sublimation in the water balance of cold regions including the MRB warrants this study.

2.2 Surface Sublimation

Surface sublimation represents the continual exchange of water between the air (in the vapor phase) to or from the ice- or snowpack (in the solid phase). Following van den Broeke (1997), Q_{surf} is estimated from

$$Q_{surf} = \rho' u_* q_* \tag{2}$$

where u_* (m s^{-1}) is the friction velocity and ρ' (negative) is a conversion factor to units of mm yr^{-1} SWE. Assuming neutral stability near the surface and a logarithmic variation of wind speed with height, u_* can be obtained from (Garratt 1992):

$$u_* = \kappa U / \ln[(z+z_0)/z_0] \tag{3}$$

in which U (m s^{-1}) is the wind speed at a height z (m) above the surface, κ (= 0.4) depicts the von Kármán constant and z_0 (m) represents the aerodynamic roughness length for momentum. Similarly, the humidity scale q_* (kg kg^{-1}) is deduced from a logarithmic moisture profile by assuming saturation with respect to ice near the surface such that (Garratt 1992):

$$q_* = \kappa q_{si}[RH_i(z)-1] / \ln[(z+z_q)/z_q] \tag{4}$$

in which q_{si} (kg kg^{-1}) represents the saturation mixing ratio with respect to ice near the surface and z_q (m), taken here as equal to z_0, denotes the roughness length for moisture over snow. Surface sublimation therefore depends critically on the gradients of both humidity and wind speed near the surface. Note that when the relative humidity with respect to ice $RH_i >$ 1.0 and q_* becomes positive, deposition to the surface (or negative sublimation) is said to occur. For simplicity, we assume in this study that all surfaces are generally flat and homogeneous. Note however that interactions between vegetation canopies and snow can lead to significant variations in surface sublimation rates compared to open areas (Pomeroy and Dion 1996; Pomeroy et al. 2007).

2.3 Blowing Snow Transport

Blowing and drifting snow occur when wind speeds exceed a certain threshold value and initiate the transport of snow that was formerly at the surface. Precipitating snow may also induce blowing snow, making the source of blown snow somewhat difficult to resolve in many instances. Two substantive processes are involved during blowing and drifting snow: saltation and suspension. Saltation is snow particles bouncing along the surface at heights of a few centimeters providing then a source for snow suspension (Pomeroy et al. 1997). Suspension occurs when snow particles are entrained by turbulent motions within the ABL. In this mode, particles may rise to 100 m or more above the surface (King and Turner 1997).

Although blowing snow usually refers to suspended snow that reduces visibility at eye level and drifting snow refers to snow transport below that height, we do not distinguish between them and consider both to be blowing snow. The 10-m wind speed threshold (U_t) for initiation of transport is usually about 5 to 10 m s^{-1} (King and Turner 1997), depending on such environmental factors as temperature and moisture conditions of the snowpack and the age of the snow (Schmidt 1980). Here we follow Li and Pomeroy (1997) who found a dependence on the 2-m air temperature T_a (°C) for U_t (m s^{-1}) as

$$U_t = U_{t0} + 0.0033(T_a + 27.27)^2 \tag{5}$$

where the minimum value of the threshold 10-m wind speed, U_{t0}, is equal to 6.98 m s^{-1} and is reached at about T_a = -27°C. This parabolic equation predicts higher resistance to transport at very low temperatures and near the freezing point. Near 0°C, the snow tends to be wet, and the imbedded water leads to higher cohesion of the snowpack. On the other hand, at very low temperatures, cohesion associated with strengthening elastic and frictional forces again reduces the capacity of the wind to displace snow from the surface. The intermediate range -25°C < T_a < -10°C is defined by Li and Pomeroy (1997) as the cold cohesive regime in which wind transport of dry snow is generally most favorable.

A number of empirical relationships describing the transport of blowing snow in terms of the wind speed can be found in the literature, as summarized by both Giovinetto et al. (1992) and Pomeroy and Gray (1995). Since the results of Pomeroy and Gray (1995) are based on measurements conducted in the Canadian Prairies that extend into the southern sections of the MRB, we follow their results which express Q_t (kg m^{-1} s^{-1}) as:

$$Q_t = BU_{10}{}^{C} \tag{6}$$

where U_{10} is the 10-m wind speed (m s^{-1}), $B = 2.2 \times 10^{-6}$ kg m$^{-5.04}$ s$^{-3.04}$, and $C = 4.04$. Once the blowing snow transport rate is known, its net contribution to the surface mass balance is found by

$$D = -\rho'/\rho \nabla \cdot Q_t \tag{7}$$

with ρ (kg m^{-3}) being the air density. Note that by definition, positive values of D indicate horizontal divergence of mass through wind redistribution, and hence a sink in the mass balance equation. Given that D is inversely proportional to the fetch for blowing snow, however, this term decreases in importance for constant values of Q_t and for increasing fetches for snow transport.

2.4 Blowing Snow Sublimation

Blowing snow occurs when loose particles of snow at the surface are entrained by winds exceeding a certain threshold for transport. As particles become suspended in a sub-saturated (with respect to ice) ABL, they sublimate at relatively fast rates despite exhibiting certain self-limiting properties (Déry et al. 1998; Déry and Yau 1999a, 2001a; Gordon and Taylor 2007). The modeling of blowing snow has recently attracted much interest in the hydrometeorological community given its possible twofold impact (i.e., terms D and Q_s in Eq. 1) to the water budget of snow-covered regions (e.g., Bintanja 1998; Déry and Yau 2001b; Essery et al. 1999; Xiao et al. 2000). However, these models cannot be applied directly to long-term, global datasets as they are computationally restrictive. Therefore, to estimate blowing snow sublimation, we use a parameterization for Q_s derived by Déry and Yau (2001a) based on the development of a double-moment blowing snow model (PIEKTUK-D). The authors found that the relationship

$$\begin{aligned} Q_s = (&a_0 + a_1\xi + a_2\xi^2 + a_3\xi^3 + a_4U_{10} + a_5\xi U_{10} + a_6\xi^2 U_{10} \\ &+ a_7U_{10}{}^2 + a_8\xi U_{10}{}^2 + a_9U_{10}{}^3)/U' \end{aligned} \tag{8}$$

provided good estimates ($R^2 = 0.95$) of Q_s (here in units of mm d^{-1} SWE) at a Canadian Arctic location. Equation (8) shows that the rate of blowing snow sublimation depends on the 10-m wind speed U_{10} (m s^{-1}) and the 2-m air temperature and humidity through a thermodynamic term ξ (-1 × 10^{-12} m^2 s^{-1}) given by:

$$\xi = (RH_i - 1)[2\,\rho_{ice}(F_k + F_d)]^{-1} \quad (9)$$

where ρ_{ice} (kg m^{-3}) denotes the density of ice, and F_k and F_d (m s kg^{-1}) represent the conductivity and diffusion terms (both temperature dependent) associated with the sublimation process (Rogers and Yau 1989). Equation (8) is normalized by a factor U' (dimensionless) to remove a dependence on the saltation mixing ratio. Values for the coefficients $a_0 - a_9$ and U' are provided by Déry and Yau (2001a).

3 Methods

Since observations of adverse wintertime processes are scarce and conducted under very harsh conditions, we compile a climatology of these events using gridded reanalysis data. Thus a 15-year (from 1979 to 1993 inclusive) climatology of significant cold-season "events" for the MRB and the globe is inferred from 6-hourly European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-15) data on a 2.5° latitude × 2.5° longitude grid (Gibson et al. 1997). The presence of snow is determined from the snow depth parameter of the ERA-15 data that is based on station observations; sea ice coverage is provided by the Canadian Meteorological Centre (CMC) at the same resolution as the former.

In this work, the definition of a blowing snow event is any day when the surface was snow-covered land or sea ice (in concentration of 50% or more), the temperature was below 0°C, and the threshold velocity for transport was exceeded, as determined from Eq. (5), at any grid point. The ERA-15 data are available four times daily (0000, 0600, 1200, and 1800 UTC). A blowing snow event is considered to have occurred if the criteria for such an event are satisfied at any one of the four times. However, for all grid points we limit the number of events to a maximum of one per day. This allows a comparison of our results with observations, since significant cold-season processes are often recorded only on a daily basis, i.e., a day with or without such an event.

Once a surface sublimation or a blowing snow event has been detected, Q_{surf}, Q_s, and Q_t are computed following the equations given in the previous section. These rates are assumed constant for the 6-hourly periods of the ERA-15 data. However, we ensure that the mass eroded through any of these processes does not exceed that present at the surface as deduced from the ERA-15 data. Note that we compute RH_i only when air temperatures are below the freezing point. In the event that RH_i is > 1.0, we set Q_s to zero despite the inference of a blowing snow event. Once the data have been compiled for the 15 years that span the ERA-15 data, we present our results using a polar stereographic projection.

4 Results

4.1 Climatology of Blowing Snow Events in the MRB

Although most of the MRB is forested, the northeastern and northern sections of the Basin lie within the tundra and the southern Basin is in the Canadian prairies (Woo et al. 2007). The tundra region is conducive to frequent blowing snow events (≥ 10 per year) which decline rapidly in number within the boreal forest, where fewer than five episodes occur annually (Fig. 1). The number of occurrence increases southward as the boreal forest gives way to the open prairies. Compared with Phillips (1990), we underestimate the frequency of blowing snow events in the forested regions of the MRB, where the ERA-15 displays a negative wind speed bias (Déry and Yau 1999b).

4.2 Contribution of Blowing Snow to the Mass Balance of MRB

Of the three processes examined in this study, only surface sublimation contributes significantly to the overall water budget of the MRB (Fig. 2a). Blowing snow sublimation reaches no more than a few millimeters per year SWE along the Arctic coast and remains negligible throughout most of the Basin (Fig. 2b). Minimal values of the blowing snow transport and divergence rates also prevail within the boreal forest in association with the reduced wind speeds there (Fig. 3). Thus, we find that all three terms combine to erode 28.8 mm yr^{-1} SWE from the MRB, equivalent to 7% of the annual precipitation input into the Basin (Fig. 4). Note also that this value is comparable with the sublimation rates reported at four individual meteorological stations within the MRB (Déry and Yau 2002). In the summer, both surface and blowing snow sublimation approach zero as the Basin remains nearly void of snow (Fig.5). Peak values of blowing snow (surface) sublimation are achieved during the month of January (April).

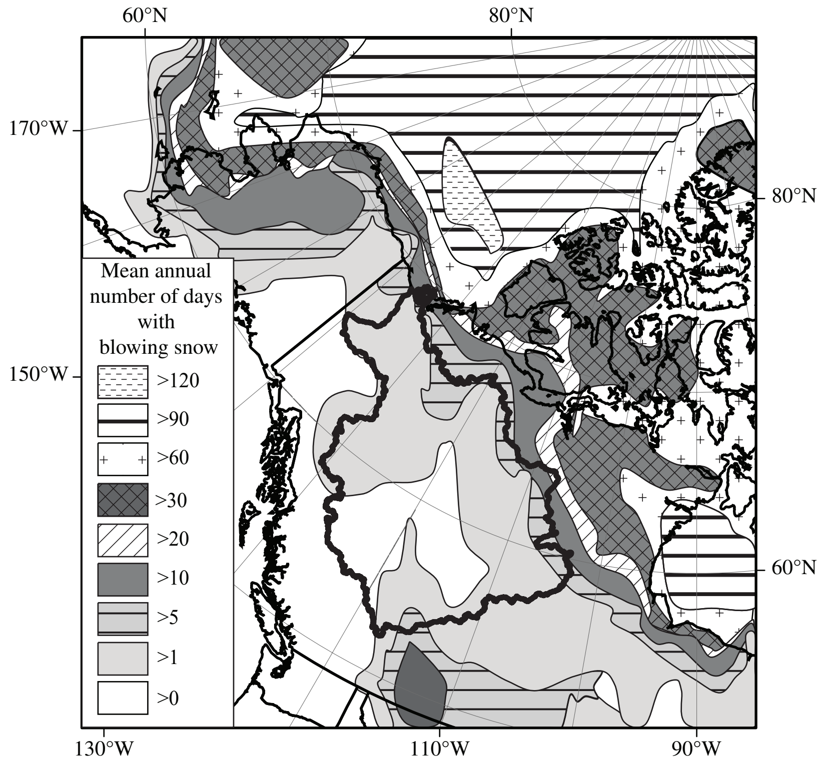

Fig. 1. Mean annual number of days with blowing snow events in Western Canada and Alaska for 1979–93. Boundary of Mackenzie River Basin is denoted by the thick line

An example of the zonally-averaged surface and blowing snow sublimation rates for 1979–99 is displayed in Fig.6. Note that the surface sublimation rates exhibit a seasonal cycle, with maximum values occurring in late spring and early fall and with lesser contributions during winter. Rare episodes of blowing snow sublimation are associated with strong wind events. During summer, there is no sublimation as snow is absent.

For the MRB as a whole, Déry and Yau (1999b) demonstrated that the divergence and sublimation of blowing snow may not contribute significantly to the large-scale surface mass balance since blowing snow events are rare in the largely forested Basin. On the other hand, significant surface sublimation is shown to occur across the entire Basin. This is at variance with the results of Walsh et al. (1994) who utilized 18 years of rawinsonde

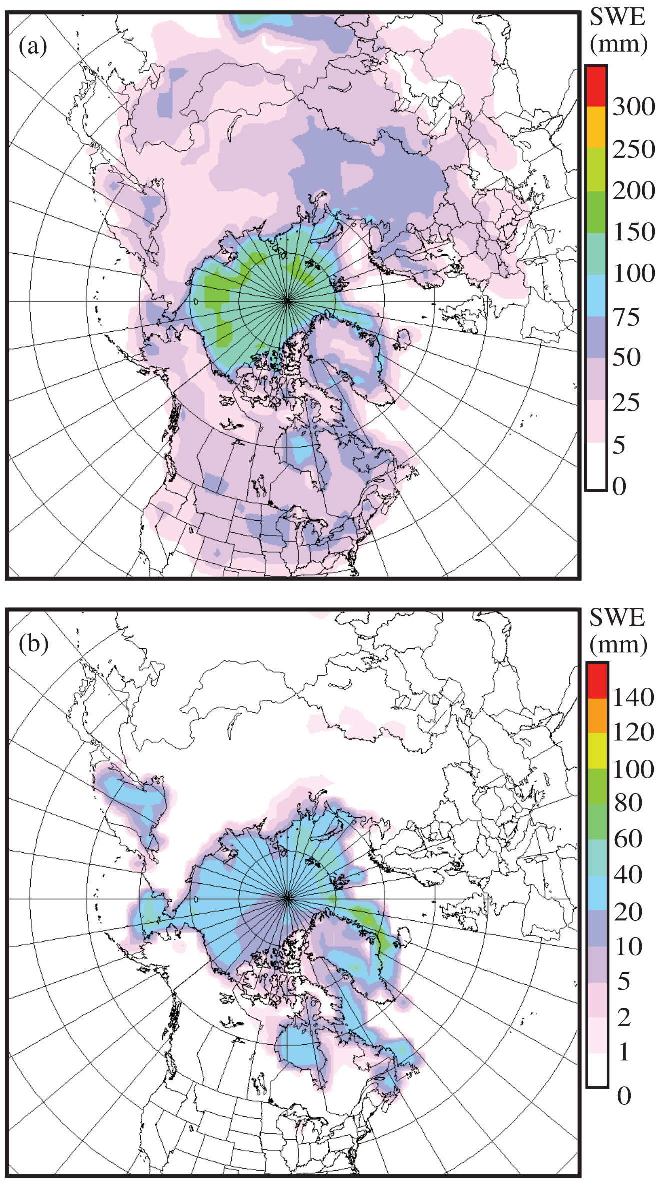

Fig.2. Mean annual (a) surface and (b) blowing snow sublimation rate (mm SWE) for the period 1979–93 in Northern Hemisphere

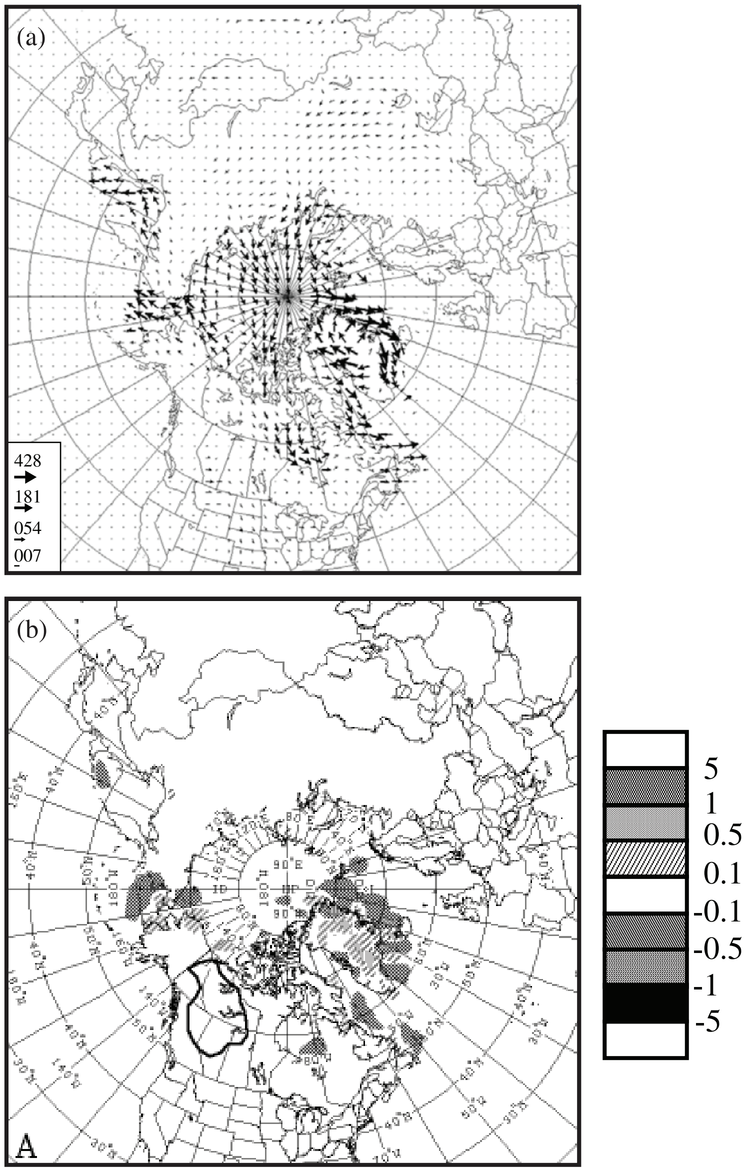

Fig. 3. Mean annual blowing snow (a) mass transport vectors ($Mg\ m^{-1}$) and (b) divergence rates (mm SWE) for the period 1979–93 in Northern Hemisphere

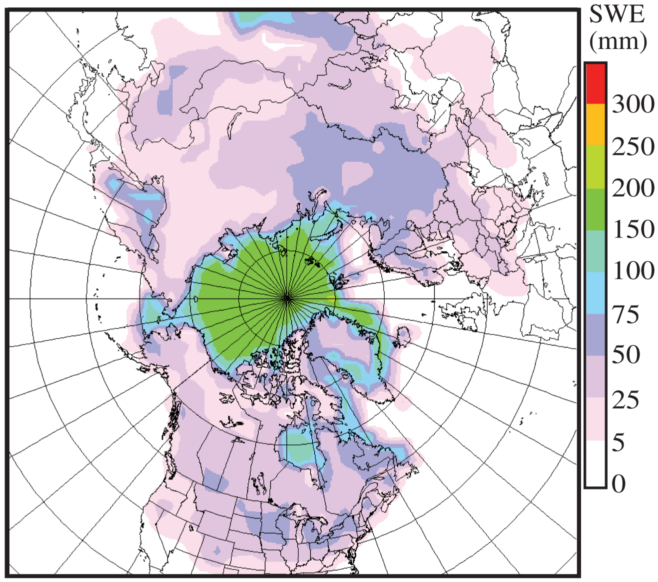

Fig. 4. Mean annual total surface sublimation, blowing snow sublimation, and divergence rates (mm SWE) for the period 1979–93 in Northern Hemisphere

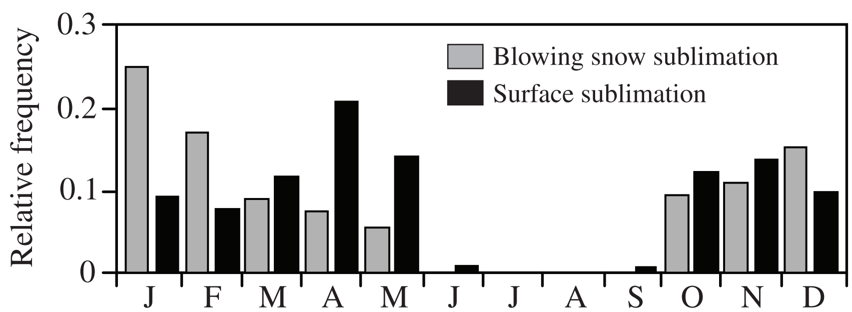

Fig. 5. Relative monthly contributions to the annual surface and blowing snow sublimation rates over the Mackenzie River Basin, 1979–93

data to obtain as a residual the evaporation over the MRB. They found negative monthly values of surface sublimation (i.e., deposition) approaching in some cases -10 mm SWE between October and April, and an annual total of about -45 mm SWE. In contrast, our estimates of surface sublimation for the same Basin are somewhat less than those of Betts and Viterbo (2000) who found the evaporation from the snow surface to average 92 mm yr^{-1} SWE over two years using the ECMWF forecast model. However, the authors noted a large positive bias (of about 60%) in the evaporation rate arising in the ECMWF model such that this value should in fact be closer to 37 mm yr^{-1} SWE. We determine that blowing snow and surface sublimation erode ≈ 29 mm yr^{-1} SWE over the entire MRB, a value more in line with the results of Betts and Viterbo (2000) than those of Walsh et al. (1994).

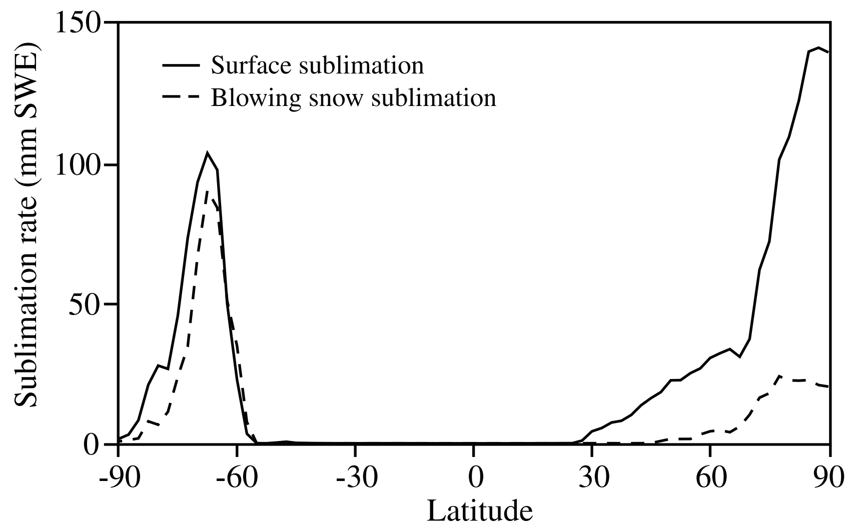

Fig. 6. Zonally-averaged surface and blowing snow sublimation rates (mm SWE) for 1979–99

5 Discussion

We have demonstrated that the ERA-15 dataset provides useful information for the detection of blowing snow and surface sublimation events and their impact on the large-scale surface mass balance. Nevertheless, some uncertainties exist in our results. The major factors of concern and their possible effects (listed in parentheses) are:

- the simple expression used to estimate wind speeds at which snow transport initiates or terminates (reductions or increases in Q_s, Q_t, and D)
- the usage of the blowing snow parameterizations at the scales of the ERA-15 dataset and that are not constrained by direct observations (changes to all blowing snow fluxes)

- the application of a parameterization for blowing snow sublimation based on the unsteady PIEKTUK-D model that generates relatively low sublimation rates of blowing snow compared to other snowdrift models, such as the Prairie Blowing Snow Model (Pomeroy et al. 1993), in part due to the appearance of negative thermodynamic feedbacks on the phase change process (a reduction of Q_s)
- the neglect of blowing snow in the ECMWF operational model and, hence, of the thermodynamic feedbacks associated with blowing snow sublimation in the ERA-15 dataset (an increase in Q_s and Q_{surf})
- the exclusion of surface sublimation during blowing snow events (a decrease in Q_{surf})
- the assumption of homogeneous surfaces and the neglect of blowing snow interception by vegetation (an increase in Q_t)
- the omission of surface sublimation from a forest canopy and when air temperatures exceed the freezing point (a decrease in Q_{surf})
- the sensitivity of surface and blowing snow sublimation rates to the ambient humidity (underestimation or overestimation of Q_s and Q_{surf}, depending on the accuracy of RH_i values)
- the assumption of neutral stability in the ABL (changes to Q_{surf})
- the uncertainty in the effects of entrainment and/or advection of dry air on sublimation processes (an underestimation of Q_s and Q_{surf}).

This study presents the first comprehensive attempt at establishing the large-scale impact of blowing snow and surface sublimation on the large-scale surface mass balance. As such, it provides possibly the best assessment yet of the significance of each of the relevant variables. However, we note that large uncertainties remain in closing the water budget at high latitudes, and more research is required to ascertain (or to improve) the results presented. A priority for future work, therefore, is the determination of scaling relations between station-based and large-scale gridded data.

6 Conclusion

A surface mass balance study of the MRB was conducted using the ERA-15 data for the years 1979–93 inclusive at a horizontal resolution of 2.5°. Emphasis was placed on surface sublimation, blowing snow sublimation and divergence. The results of the water budget computation indicate that this Basin loses mass on the order of 29 mm yr^{-1} SWE through these processes, disposing ≈7% of the total annual precipitation over the MRB. However, the importance of blowing snow sublimation and divergence varies widely within the Basin, with the greatest effects over the Arctic tundra.

As demonstrated by our underlying assumptions, the results provide only a first-order estimate of the contributions of these terms to the MRB surface mass balance. Blowing snow transport, for instance, is known to displace locally significant amounts of mass into large snowdrifts whereas other areas become nearly devoid of snow (e.g., Sturm et al. 2001; Woo et al. 1983). Nonetheless, as we approach a full closure of high latitude water budgets, this contribution is a first step in establishing the large-scale climatological impacts of surface sublimation, blowing snow sublimation and divergence on the MRB surface mass balance.

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List of Symbols

$a_0 - a_9$	empirical coefficients
B	empirical coefficient [= 2.2×10^{-6} kg $m^{-5.04}$ $s^{-3.04}$]
C	empirical coefficient [dimensionless]
D	horizontal divergence of mass through wind redistribution [mm yr^{-1} SWE; or m s^{-1} SWE]
E	evaporation rate [mm yr^{-1} SWE]
F_k, F_d	conductivity and diffusion terms [m s kg^{-1}]
M	divergence of water after melt and runoff [mm yr^{-1} SWE]
P	precipitation rate [mm yr^{-1} SWE]
Q_t	rate of blowing snow transport [kg m^{-1} s^{-1}]
Q_{surf}	surface sublimation rate [mm yr^{-1} SWE]
Q_s	sublimation of blowing snow [mm yr^{-1} SWE; or mm d^{-1} SWE]
q_{si}	saturation mixing ratio with respect to ice near the surface [kg kg^{-1}]
q_*	humidity scale [kg kg^{-1}]
RH	relative humidity [kg kg^{-1}]
S	change in storage or accumulation of snow at the surface [mm yr^{-1} SWE]
T_a	air temperature [°C]
U	wind speed at a height z [m] above the surface [m s^{-1}]
U_t	10-m wind speed threshold for initiation of snow transport [m s^{-1}]
U_{t0}	minimum value of the threshold 10-m wind speed [m s^{-1}]
U'	factor to remove dependence on the saltation mixing ratio [dimensionless]
u_*	friction velocity [m s^{-1}]
z_0	aerodynamic roughness length for momentum [m]
z_q	aerodynamic roughness length for moisture over snow [m]
κ	von Kármán constant [= 0.4]
ρ	density of air [kg m^{-3})]
ρ_{ice}	density of ice [kg m^{-3}]
ρ'	factor for conversion [to unit of mm yr^{-1} SWE]
ξ	a thermodynamic term [-1×10^{-12} m^2 s^{-1}]