Large-scale mass balance effects of blowing snow and surface sublimation

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Received 27 August 2001; revised 22 March 2002; accepted 30 March 2002; published 3 December 2002.

[1] This study examines the effects of surface sublimation and blowing snow on the surface mass balance on a global and basin scale using the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA15) data at a resolution of 2.5° that span the years 1979–1993. The combined processes of surface and blowing snow sublimation are estimated to remove 29 mm yr^{-1} snow-water equivalent (swe) over Antarctica, disposing about 17 to 20% of its annual precipitation. In the Northern Hemisphere, these processes are generally less important in continental areas than over the frozen Arctic Ocean, where surface and blowing snow sublimation deplete upward of 100 mm yr^{-1} swe. Areas with frequent blowing snow episodes, such as the coastal regions of Antarctica and the Arctic Ocean, are prone to a mass transport >100 Mg m^{-1} yr^{-1}. Although important locally, values of the divergence of mass through wind redistribution are generally 2 orders of magnitude less than surface and blowing snow sublimation when evaluated over large areas. For the entire Mackenzie River Basin of Canada, surface sublimation remains the dominant sink of mass as it removes 29 mm yr^{-1} swe, or about 7% of the watershed's annual precipitation. Although the first of its kind, this study provides only a first-order estimate of the contribution of surface sublimation and blowing snow to the surface mass balance because of limitations with the data set and some uncertainties in the blowing snow process. *INDEX TERMS:* 1833 Hydrology: Hydroclimatology; 1863 Hydrology: Snow and ice (1827); 3349 Meteorology and Atmospheric Dynamics: Polar meteorology; *KEYWORDS:* blowing snow, Mackenzie Basin, mass balance, snow, sublimation, water budget

Citation: Déry, S. J., and M. K. Yau, Large-scale mass balance effects of blowing snow and surface sublimation, *J. Geophys. Res.*, *107*(D23), 4679, doi:10.1029/2001JD001251, 2002.

1. Introduction

[2] In cold climate regimes, precipitation tends to fall in the form of snow that accumulates at the surface. A nival regime constitutes an important form of storage for water that may be released during melt to the environment or that may be accumulated into an ice sheet if subfreezing conditions persist.

[3] Movement of air above a snowpack impacts the surface mass balance in several ways. For instance, winds may remove snow from easily erodible surfaces to relocate it in accumulation areas such as depressions and bushes, leading to substantial heterogeneities in the snow cover with hydrometeorological implications. In subsaturated air with respect to ice, the sublimation (phase change of ice to water vapor) of suspended blowing snow particles may represent an additional sink of mass from the surface. In a process referred here to as surface sublimation, supplemental water vapor may be directly transferred between the atmospheric boundary layer (ABL) and the snowpack, even when wind speeds do not exceed the threshold for transport of snow.

[4] Considering that high-latitude areas are currently undergoing significant changes in their environment [e.g., *Serreze et al.*, 2000], there is an increasing demand for accurate information on the hydrological cycle in polar regions. As outlined in several works [see, e.g., *Cullather et al.*, 2000], there are still some large uncertainties in the surface mass balance of the polar regions. The sparse observational network at high latitudes and complications arising from measurement taking in harsh climate conditions have contributed to serious deficiencies in the available meteorological data there [e.g., *Déry and Stieglitz*, 2002]. Thus, we are often required to seek alternative data sets to those derived from observations, especially for large-scale mass balance studies in such remote areas. Hence, we make use here of the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA15) product [*Gibson et al.*, 1997] for the years 1979 to 1993 to gauge the effects of blowing snow and surface sublimation to the global water budget. The ERA15 has been selected since several research groups, including *Cullather et al.* [2000], *Serreze and Hurst* [2000] and *Hagemann and Dümenil Gates* [2001], have shown that the ERA15 represents with accuracy the hydrological cycle at high latitudes.

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0148-0227/02/2001JD001251$09.00

[5] On a regional scale, a location where the components of the surface mass balance are still not well understood is the Mackenzie River Basin (MRB). This large watershed covers $\approx 1.8 \times 10^6$ km^2 in northwestern Canada and constitutes one of the most important sources of freshwater to the Arctic Ocean [*Stewart et al.*, 1998]. To study the energy and water cycles of the MRB, an intensive field campaign titled the Mackenzie GEWEX Study (MAGS) [*Stewart et al.*, 1998; *Rouse*, 2000] was launched in the early 1990s as Canada's contribution to the Global Energy and Water Cycle Experiment (GEWEX). This has provided a better understanding on some aspects of the MRB's water budget including its sources and sinks of water [e.g., *Lackmann et al.*, 1998; *Smirnov and Moore*, 1999; *Misra et al.*, 2000]. However, few of these studies have focused on cold season processes. Considering that parts of the MRB are blanketed by snow up to 200 days annually [*Phillips*, 1990], it becomes obvious that snow and ice processes play a critical role on its surface water budget [*Stewart et al.*, 2000].

[6] The objective of this study is to determine the large-scale impacts of blowing snow and surface sublimation in the global water budget, with specific reference to the MRB. To achieve this goal, we use the global ERA15 data set and hence, provide some much needed information on these processes in both northern and southern polar regions. The paper begins with some background information on the processes affecting the surface mass balance of cold regions, with special attention given to the surface sublimation and blowing snow components. Subsequently, the methodology is described followed by some results, sensitivity tests, and then comparisons with other studies. A discussion of the uncertainties in our results ensues and the paper closes with a summary of our findings.

2. Background

2.1. Mass Balance

[7] Following *King et al.* [1996], the annual surface mass balance for a nival regime may be expressed as:

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

where S is the storage or accumulation of snow at the surface, P is the precipitation rate, E is the evaporation rate which 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 of snow by wind transport and the sublimation of blowing snow, respectively. All terms in equation (1) are expressed in units of mm yr^{-1} snow-water equivalent (swe) unless stated otherwise.

[8] There has been recently substantial effort in the literature to determine the amount of precipitation less evaporation ($P - E$) in the polar regions, notably for Antarctica [e.g., *Giovinetto et al.*, 1992; *Connolley and King*, 1996; *Cullather et al.*, 1998; *Turner et al.*, 1999] as well as for the ice-covered Arctic Ocean [*Walsh et al.*, 1994, 1998; *Cullather et al.*, 2000]. In most cases, these studies have neglected terms linked to blowing snow as they were considered negligible, at least on a continental scale [*Connolley and King*, 1996; *Turner et al.*, 1999]. Nevertheless, *Giovinetto et al.* [1992] estimate a net transport of about 120×10^{12} kg yr^{-1} of mass northward across 70°S through blowing snow, equivalent to the removal of 7.7 mm yr^{-1} swe over the area south of that latitude. Other authors have suggested that the sublimation of blowing snow may also erode substantial amounts of mass from the Antarctic continent [e.g., *Bintanja*, 1998; *Gallée*, 1998; *Gallée et al.*, 2001] or the Arctic tundra [*Pomeroy et al.*, 1997; *Essery et al.*, 1999]. Few studies have examined the individual contribution of surface sublimation to the water budget of polar regions on a continental or basin scale, with perhaps the exception of *van den Broeke* [1997] and *Genthon and Krinner* [2001].

[9] 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 recently 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; *Smirnov and Moore*, 1999; *Misra et al.*, 2000], none of these authors have attempted to evaluate the impact of blowing snow to the surface mass balance of the MRB as a whole. Although *Walsh et al.* [1994] and *Betts and Viterbo* [2000] examine the role of surface sublimation in the MRB, the authors come to contradictory conclusions, with the former yielding significant deposition during wintertime over the MRB 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 climate regions including the MRB warrants this study.

2.2. Humidity

[10] Before we proceed, a brief discussion on the issue of ambient humidity is required since sublimation processes depend in part on the moisture content of the ABL. To derive the relative humidity with respect to ice RH$_i$ from the dew-point temperature, we may use the Magnus equation to compute the saturation mixing ratio over ice q_{si} (kg kg^{-1}) [*Kong and Yau*, 1997; *Pruppacher and Klett*, 1997]:

$$q_{si} = \frac{3.8}{P_s}\exp\left(\frac{21.87(T_a - T_0)}{(T_a - 7.66)}\right), \tag{2}$$

where P_s is the surface atmospheric pressure (hPa), T_a the air temperature (K) and T_0 a constant with a value of 273.16 K. Similarly, the saturation mixing ratio over water q_s (kg kg^{-1}) is obtained by [*Kong and Yau*, 1997]:

$$q_s = \frac{3.8}{P_s}\exp\left(\frac{17.27(T_a - T_0)}{(T_a - 35.86)}\right). \tag{3}$$

Given the dew-point temperature T_d (K), we may compute the specific humidity q_v (kg kg^{-1}) in the air by substituting T_a by T_d in equation (3). Then the relative humidity with respect to ice is simply RH$_i$ = q_v/q_{si} while the relative humidity with respect to water RH$_w$ = q_v/q_s. Since at subfreezing temperatures $q_{si} \leq q_s$ at all times, it is possible that air subsaturated with respect to water may be supersaturated with respect to ice, thereby promoting deposition rather than sublimation (see discussion below).

2.3. Surface Sublimation

[11] 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). Here, we follow the methodology of *van den Broeke* [1997] by estimating Q_{surf} from

$$Q_{surf} = \rho' u_* q_*, \tag{4}$$

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]:

$$u_* = \frac{\kappa U}{\ln\left(\frac{(z+z_0)}{z_0}\right)}, \tag{5}$$

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. In a similar fashion, 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]:

$$q_* = \frac{\kappa q_{si}(\mathrm{RH}_i - 1)}{\ln\left(\frac{(z+z_q)}{z_q}\right)}, \tag{6}$$

in which 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 $\mathrm{RH}_i > 1.0$ 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 [*Schmidt*, 1991; *Pomeroy and Dion*, 1996].

2.4. Blowing Snow Sublimation

[12] 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 subsaturated (with respect to ice) ABL, they sublimate at relatively fast rates despite exhibiting certain self-limiting properties [*Déry et al.*, 1998; *Déry and Yau*, 1999b; 2001a]. 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 equation 1) to the water budget of snow-covered regions [e.g., *Bintanja*, 1998; *Essery et al.*, 1999; *Xiao et al.*, 2000; *Déry and Yau*, 2001b]. However, these models cannot be applied directly to long-term, global data sets 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

$$Q_s = \left(a_0 + a_1\xi + a_2\xi^2 + a_3\xi^3 + a_4U_{10} + a_5\xi U_{10} + a_6\xi^2 U_{10} + a_7U_{10}^2 + a_8\xi U_{10}^2 + a_9U_{10}^3\right)/U', \tag{7}$$

provided good estimates (R^2 = 0.95) of Q_s (here in units of mm d^{-1} swe) at a Canadian Arctic location. Equation (7)

Table 1. Coefficients for Equation (7)

Coefficient	Value
a_0	3.78407×10^{-1}
a_1	-8.64089×10^{-2}
a_2	-1.60570×10^{-2}
a_3	7.25516×10^{-4}
a_4	-1.25650×10^{-1}
a_5	2.48430×10^{-2}
a_6	-9.56871×10^{-4}
a_7	1.24600×10^{-2}
a_8	1.56862×10^{-3}
a_9	-2.93002×10^{-4}

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:

$$\xi = \frac{(\mathrm{RH}_i - 1)}{2\rho_{ice}(F_k + F_d)} \tag{8}$$

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 (7) is normalized by a factor U' (dimensionless) to remove a dependence on the saltation mixing ratio. Values for the coefficients $a_0 - a_9$ are provided in Table 1.

[13] Although developed with meteorological data collected at a site in the Canadian Arctic, equation (7) is applicable to other regions of the world where the blowing snow process deviates little from its manifestation in the Canadian Arctic. For instance, vertical profiles of suspended blowing snow particles reported by *Mann et al.* [2000] at Halley, Antarctica, closely resemble those of *Pomeroy and Male* [1992] in the Canadian prairies. The development of the PIEKTUK-D model and the blowing snow sublimation parameterization are based in large part on the blowing snow observations of *Pomeroy and Male* [1992] and *Pomeroy and Gray* [1990] that ultimately resulted in their Prairie Blowing Snow Model (PBSM) [*Pomeroy et al.*, 1993]. Thus, we are confident that the application of equation (7) will yield reasonable approximations of the blowing snow sublimation rates at various locations worldwide. Some possible exceptions include the blue ice areas of Antarctica where, according to *Bintanja* [2001a], the snow drifting rates differ somewhat from those over its snow-covered fields. However, blue ice regions cover only a small fraction of the total area of the Antarctic continent and will not impact considerably the results of this large-scale mass balance study.

2.5. Blowing Snow Transport

[14] 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 which extend into the southern sections of the MRB, we follow their results which express Q_t (kg m^{-1} s^{-1}) as:

$$Q_t = BU_{10}^C, \tag{9}$$

Table 2. Observed and ERA15 Mean Annual Conditions of Air Temperature, Dew-Point Temperature, Relative Humidity With Respect to Ice and 10-m Wind Speed at Four Canadian Meteorological Stations During 1993[a]

Station	Latitude, Longitude	T_a, °C		T_d, °C		RH_i		U_{10}, m s^{-1}	
		OBS	ERA15	OBS	ERA15	OBS	ERA15	OBS	ERA15
Fort McMurray	56.7°N, 111.3°W	1.6	−3.0	−4.4	−7.1	0.81	0.98	2.5	2.8
Inuvik	68.3°N, 133.5°W	−5.6	−7.7	−10.8	−11.0	0.83	1.09	2.4	3.0
Norman Wells	65.3°N, 126.8°W	−3.3	−6.5	−8.6	−10.6	0.84	1.01	2.6	2.7
Yellowknife	62.5°N, 114.4°W	−3.5	−8.4	−9.3	−13.8	0.82	1.08	4.1	3.1

[a] OBS, observed; T_a, air temperature; T_d, dew-point temperature; RH_i, relative humidity with respect to ice; and U_{10}, 10-m wind speed.

where $B = 2.2 \times 10^{-6}$ 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

$$D = -\frac{\rho'}{\rho}\nabla . Q_t, \tag{10}$$

with ρ (kg m^{-3}) being the air density. To obtain D, therefore, the mass transport vectors at four neighboring grid points are required. Consequently, this allows the transfer of mass to adjacent grid cells. 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 of the surface mass balance will decrease in importance for constant values of Q_t and increasing fetches for snow transport.

3. Methodology

3.1. Data and Events

[15] The gridded data used in this study are the 6-hourly European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA15) data on a 2.5° latitude/longitude grid covering the period from 1979 to 1993 inclusive [*Gibson et al.*, 1997]. Before applying the relationships for the computations of blowing snow transport and sublimation, a "blowing snow event" must be inferred from the ERA15 data and here we follow the criteria of *Déry and Yau* [1999a]. Briefly, they define a blowing snow event to occur whenever the near-surface air temperature is subfreezing and a threshold wind speed for transport, computed following *Li and Pomeroy* [1997], are satisfied in a grid cell where the surface is snow-covered land or sea ice (in concentration of more than 50%). A "surface sublimation event" occurs when all the criteria for a blowing event are met, except that winds must be weaker than the specified threshold for blowing snow transport. During an episode of blowing snow, we therefore assume that RH_i quickly approaches saturation with respect to ice in a shallow layer near the surface such that Q_{surf} becomes minimal.

[16] Once a surface sublimation or a blowing snow event has been detected, Q_{surf}, Q_s and Q_t are computed following the equations reported in the previous section. These rates are assumed constant for the 6-hourly periods of the ERA15 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 ERA15 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 ERA15 data, we present our results using a polar stereographic projection since blowing snow and sublimation events occur most frequently in polar regions. The new grids are composed of 50 × 50 points centered over the poles and have a horizontal resolution of 250 km true at 60°N or S.

3.2. Evaluation of Data and Methods

[17] A few tests are conducted to evaluate the data and methodology we propose for the mass balance study. For that purpose, we have selected four meteorological stations within the MRB (see Table 2). The 6-hourly observations of T_a, T_d and U_{10} were thus obtained for Fort McMurray, Alberta and Inuvik, Norman Wells and Yellowknife, Northwest Territories during 1993. Values of RH_i were subsequently computed following the equations reported in section 2.2. Note that these humidity measurements do not appear to suffer from the instrument errors induced by icing observed at other locations in the Canadian Arctic [*Déry and Stieglitz*, 2002]. Similar time series were then generated by interpolating the ERA15 meteorological data onto the four stations using a technique similar to that of *Reiff et al.* [1984].

[18] As an example, time series of the observed and ERA15 meteorological data at Inuvik, Northwest Territories, during 1993 are shown in Figure 1. This clearly demonstrates the effectiveness of the ERA15 data to represent actual conditions observed in the MRB. Nevertheless, as stated by *Gibson et al.* [1997], the ERA15 exhibits certain systematic biases. A cold temperature bias, associated with the improper treatment of some land surface processes in wintertime [*Hagemann and Dümenil Gates*, 2001], is evident in Figure 1 and exists throughout high-latitude regions in the ERA15. Even though the two time series of dew-point temperatures appear to be highly correlated ($R^2 = 0.92$), values of RH_i deduced from the ERA15 are systematically too high during wintertime due to the presence of cold temperature and dew-point biases this data set exhibits. These biases will not affect much the calculations of Q_t but may reduce Q_{surf} and Q_s substantially as both depend on RH_i. Wind speeds deduced from the ERA15 are actually greater than observed at Inuvik although *Déry and Yau* [1999a] generally find the opposite trend at various high-latitude locations. A reduction in ERA15 wind speeds is most notable in katabatic wind regions [see *Déry and Yau*, 1999a] and will tend to reduce the overall blowing snow transport and sublimation rates. Finally, a combination of the temperature and wind speed biases will affect the determination of the blowing snow events through the threshold wind speed for transport.

[19] Table 2 also provides the annual climatological values of T_a, T_d, RH_i and U_{10} at the four chosen stations during

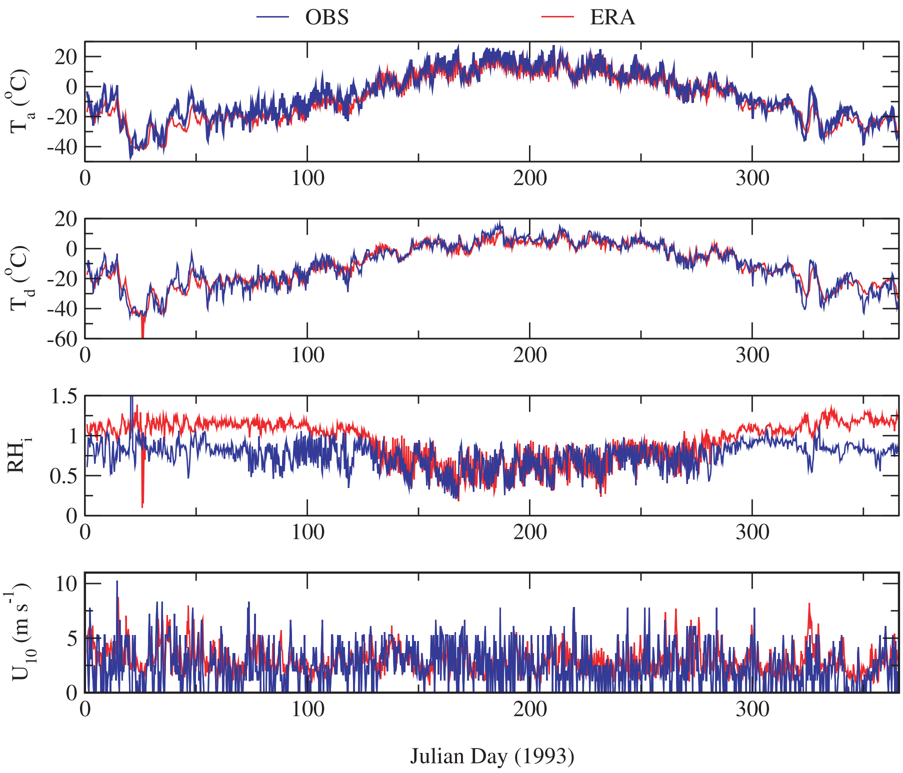

Figure 1. Six-hourly values of the observed (OBS) and ERA15 air temperature (T_a), dew-point temperature (T_d), relative humidity with respect to ice (RH_i) and 10-m wind speed (U_{10}) at Inuvik, Northwest Territories during 1993.

1993. Although the systematic errors in the ERA15 values of T_a and T_d are, once again, apparent, the overall climatology for these four stations is well depicted in 1993. This certainly gives us initial confidence in applying this data set for our global mass balance study. Of concern, however, are the elevated values of RH_i in association with the cold temperature and dew-point biases. Figure 2 demonstrates that values of the ERA15 RH_i are generally greater than their

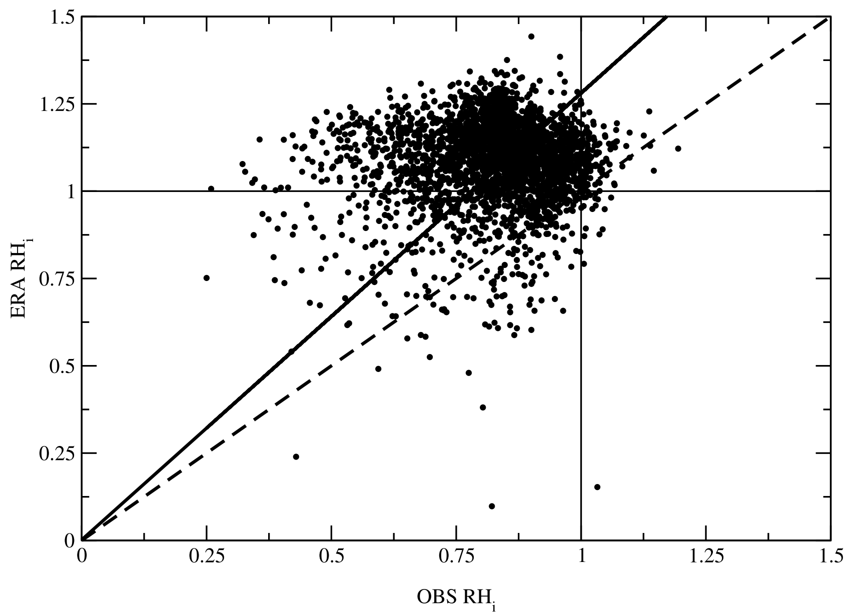

Figure 2. The observed (OBS) and ERA15 relative humidities with respect to ice (RH_i) during 1993 at Inuvik, Norman Wells and Yellowknife, Northwest Territories, and Fort McMurray, Alberta. The thick solid line represents the linear least-squares approximation to the data ($Y = 1.28X$) passing through the origin, and the thick dashed line represents the 1:1 line. Note that values of RH_i are computed only when both the observed and ERA15 temperatures are <0°C.

Table 3. Total Annual Surface Sublimation, Blowing Snow Sublimation and Transport Rates Based on Observational, the Original ERA15 and Bias-Corrected ERA15 Data at Four Canadian Meteorological Stations During 1993[a]

	Q_{surf}, mm			Q_s, mm			Q_t, Mg m^{-1}		
Station	OBS	ERA15	BIAS	OBS	ERA15	BIAS	OBS	ERA15	BIAS
Fort McMurray	33.0	17.5	55.5	0.884	0.0	0.0	2.61	0.0	0.0
Inuvik	39.0	−1.68	40.9	0.382	0.0249	0.217	1.41	1.02	1.02
Norman Wells	28.9	1.42	39.0	4.29	3.73	5.74	16.3	15.8	15.8
Yellowknife	50.5	7.02	46.1	7.49	0.106	0.317	30.0	1.38	1.38

[a] Q_{surf}, total annual surface sublimation rate; Q_s blowing snow sublimation rate; Q_t, transport rate; OBS, observational data; and BIAS, bias-corrected ERA15 data.

observed counterparts by a factor of 1.28. Since these high-latitude cold temperature and dew-point biases are present and spatially uniform within the entire ERA15 data set, we have decided to divide their values of RH$_i$ by this factor throughout our study. For consistency with our blowing snow climatology [*Déry and Yau*, 1999a], we choose here to apply a correction only to RH$_i$ and not to T_a and T_d. We justify our choice as follows. We note that the threshold wind speed for blowing snow is a parabolic function of T_a with a minimum value of about 7 m s^{-1} at −27°C [*Li and Pomeroy*, 1997]. Since most blowing snow events occur at temperatures warmer than −27°C, correcting the cold temperature bias would yield a higher threshold wind speed which can be met by correcting the generally weaker wind bias in the ERA15. In other words, correcting both the cold temperature and the weaker wind speed biases in the ERA15 would not change significantly the frequency of blowing snow events relative to the situation when both T_a and U_{10} are left uncorrected. Without a bias correction, however, the application of the surface and blowing snow sublimation parameterizations to the original ERA15 data yields much lower values of these components of the surface water budget. Indeed, using the equations reported in the previous section for the specified events, we show in Table 3 that estimates of Q_{surf} and Q_s based on the ERA15 data are much improved compared to those based on observational data when the 1.28 factor is applied to these RH$_i$ data. Note however that the comparisons are less favorable for Yellowknife where the ERA15 underestimates the mean annual wind speed by 1 m s^{-1}, thereby reducing significantly the blowing snow fluxes. Finally, observe that values of Q_t for each site are not affected by this humidity correction.

3.3. Limitations

[20] Despite the accuracy of the ERA15 to reproduce well the large-scale atmospheric environment, using these near-surface data at a resolution of 2.5° and at 6 h intervals imposes certain limitations on the veracity of the results presented in the following section. Since the surface sublimation, blowing snow sublimation and transport rates vary nonlinearly with the ambient conditions of wind speed, temperature and humidity, the temporally and spatially averaged meteorological fields may mask natural variability and extreme events. For instance, katabatic winds over the Greenland and Antarctic ice fields are not necessarily well reproduced within the ERA15 and may lead to a possible underestimation of blowing snow fluxes in these regions [*Déry and Yau*, 1999a].

[21] To address these scaling issues, another series of tests are performed utilizing the PIEKTUK-D model in a steady-state mode over a fetch of 250 km, a distance typical of an ERA15 grid cell. The model is first run with initial conditions of T_a = −10°C, RH$_i$ = 70% and U_{10} = 15 m s^{-1} in our baseline simulation. The surface is assumed to be entirely snow-covered and of sufficient depth to provide a continuous source of blowing snow particles along the fetch. The daily surface sublimation, blowing snow sublimation and transport rates for this baseline case are listed in Table 4. Next, we conducted several additional experiments whereby one of the initial meteorological conditions was slightly modified. This includes changes of ±5°C in T_a, ± 10% in RH$_i$, and ±5 m s^{-1} in U_{10}, leading to 3^3 = 27 possible combinations in the initial meteorological conditions. The intent here is to represent some of the possible spatial and/or temporal variability in meteorological conditions (excluding surface properties) that would occur in nature within an ERA15 grid cell at a given time. The ensemble run, representing the average of the 27 simulations having mean values of the meteorological conditions stipulated in the baseline case, yields remarkably similar values of the surface and blowing snow sublimation rates and, to a lesser degree, the transport rate of blowing snow (Table 4). An order of magnitude separates the largest from the smallest rates, explaining the high variance associated with the ensemble runs.

3.4. Other Considerations and Uncertainties

[22] As discussed thoroughly by *Liston and Sturm* [1998], there exist many factors that require some consideration in the study of the blowing snow phenomenon. These include, for instance, conditions of the snowpack and subgrid-scale variations in the snow cover and the surface. Deviations in the standard logarithmic wind profiles may also occur over real topography. In land areas, vegetation is also likely to intercept drifting snow while promoting sublimation from snow-covered canopies that may enhance the total latent heat fluxes over forests (see section 2.3).

Table 4. Daily Surface Sublimation, Blowing Snow Sublimation and Transport Rates for a Single Baseline Experiment Compared to Values From an Ensemble of 27 Experiments[a]

Test	Q_{surf}, mm d^{-1}	Q_s, mm d^{-1}	Q_t, 1ckg m^{-1} s^{-1}
Baseline	4.82	1.84	0.252
Ensemble	5.09	1.82	0.390
Range	1.39–13.0	0.396–4.87	0.0770–1.05
σ	2.72	1.22	0.310

[a] The range presents the minimum and maximum values, and σ presents the standard deviation for each variable in the 27 ensemble runs. Q_{surf}, daily surface sublimation rate; Q_s, blowing snow sublimation rate; and Q_t, transport rate.

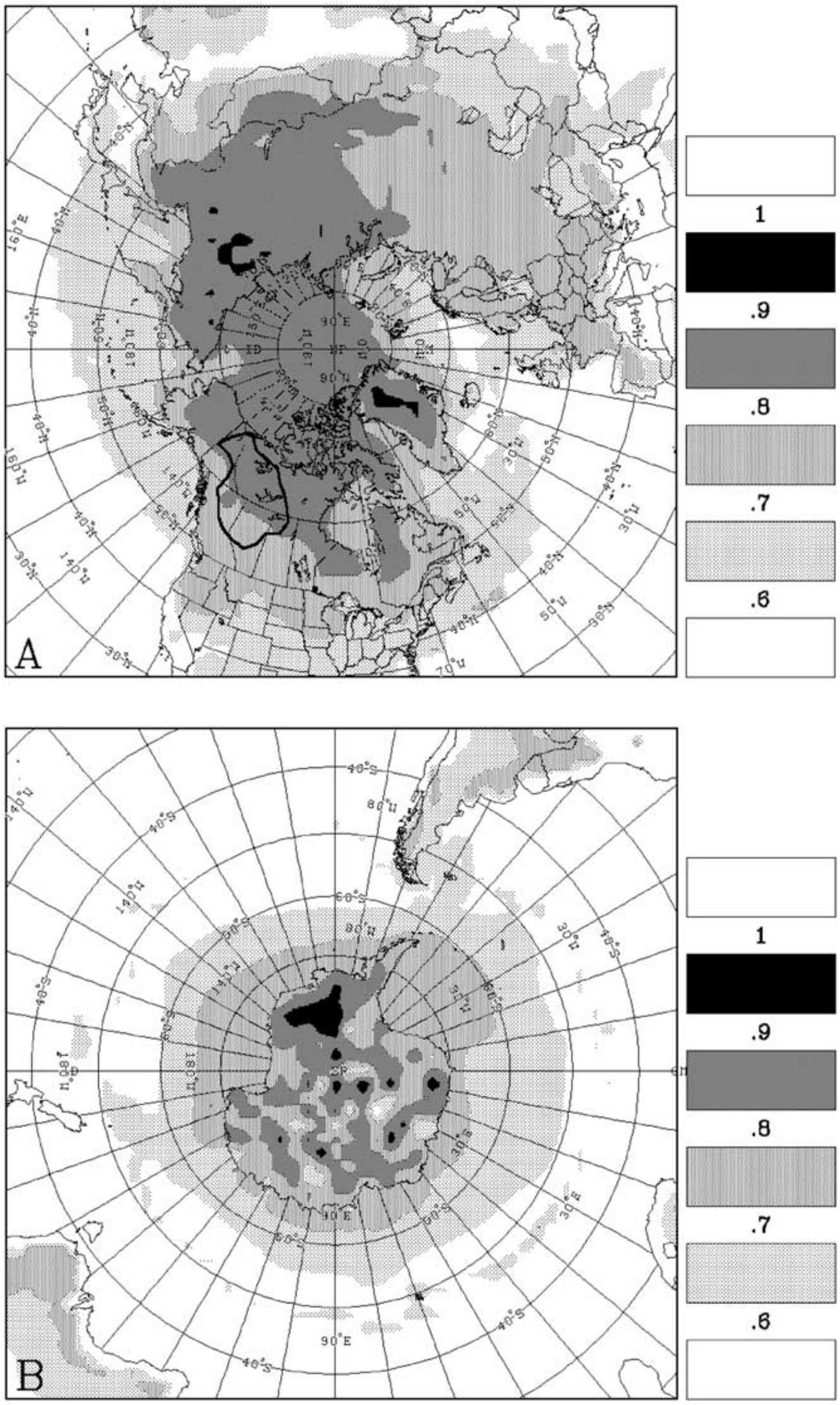

Figure 3. The mean relative humidity with respect to ice for the period 1979–1993 when T_a < 0°C (a) in the Northern Hemisphere and (b) in the Southern Hemisphere. The bold outline in Figure 3a represents the Mackenzie River Basin.

These processes thus introduce additional uncertainties in our results that require some discussion in section 7.

4. Results

4.1. Global Results

[23] As was stated earlier, both blowing snow and surface sublimation depend highly on conditions of RH_i near the surface. We first present, therefore, average values of RH_i when T_a < 0°C (and thus for varying time periods) in both hemispheres (Figure 3). Recall that these values are adjusted by a factor of 1.28 from the original ERA15 data set. This illustrates that, on average for the period 1979 to 1993, high values of RH_i are found mainly over the Antarctic and Greenland ice sheets as well as in some sections of Siberia. Due to the atmosphere's decreasing capacity to hold water vapor in the very cold temperatures experienced in these

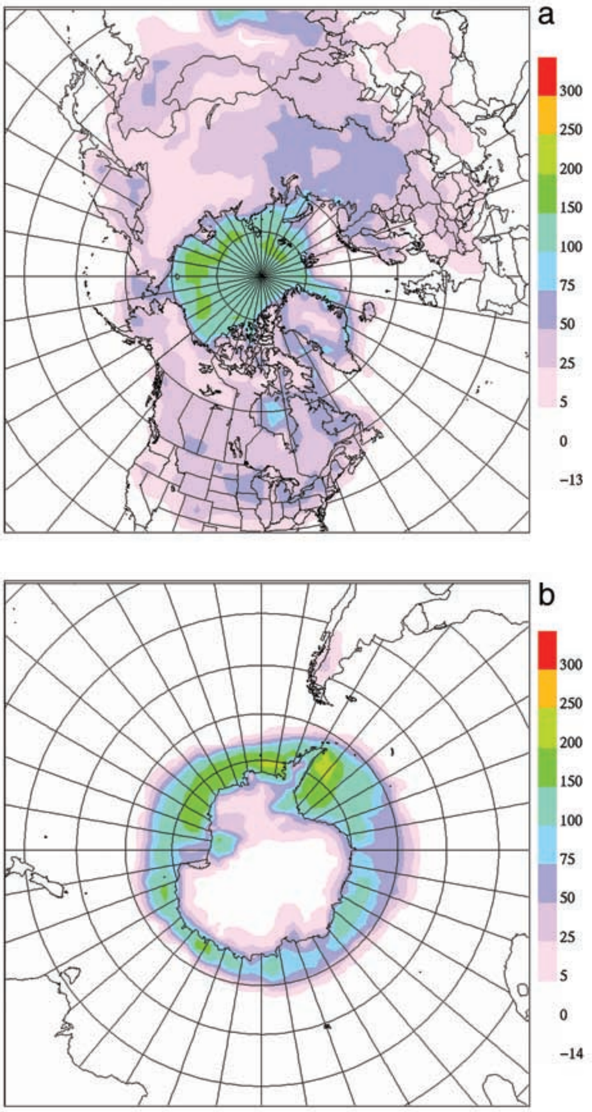

Figure 4. The mean annual surface sublimation rate (mm swe) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere.

areas, values of RH_i approaching the ice saturation point are readily attained. Relatively drier conditions exist over southern portions of North America and Eurasia, whereas the Arctic Ocean experiences a mean value of RH_i above 80%.

[24] Figure 4 presents the contribution of surface sublimation to the surface mass balance of the Northern and Southern Hemispheres. We readily observe the maximum values of surface sublimation achieved over snow-covered Arctic ($Q_{surf} > 150$ mm yr^{-1} swe) and Antarctic ($Q_{surf} > 200$ mm yr^{-1} swe) sea ice where strong winds enhance evaporative fluxes in these regions. Over land, local maxima are also found in Scandinavia and western Russia as well as in southern Canada and the northern United States (Figure 4a). On the other hand, low sublimation rates are inferred over a large portion of the Antarctic continent, Siberia and the Greenland ice sheet where bitterly cold temperatures and high RH_i values constrain latent heat fluxes.

[25] The sublimation of blowing snow computed following equation (7) is shown in Figure 5 to contribute significantly to the water budget of snow-covered sea ice in the Southern Hemisphere and, to a lesser extent, in the Northern Hemisphere. The Greenland ice sheet is also susceptible to frequent blowing snow sublimation events and some of its coastal areas experience the erosion of >20 mm yr^{-1} swe through this process. Continental regions sustain relatively less blowing snow sublimation ($Q_s < 2$ mm yr^{-1} swe) since wind speeds do not reach sufficient strengths to generate considerable episodes of blowing snow there [*Déry and Yau*, 1999a].

[26] Since the criteria for a blowing snow event must be met to apply the transport calculations, we find that, as expected, areas with the most blowing snow events [see *Déry and Yau*, 1999a] coincide with locations of large mass transport (Figure 6). Some sections of Greenland, the frozen Arctic Ocean and Antarctica are subject to the redistribution

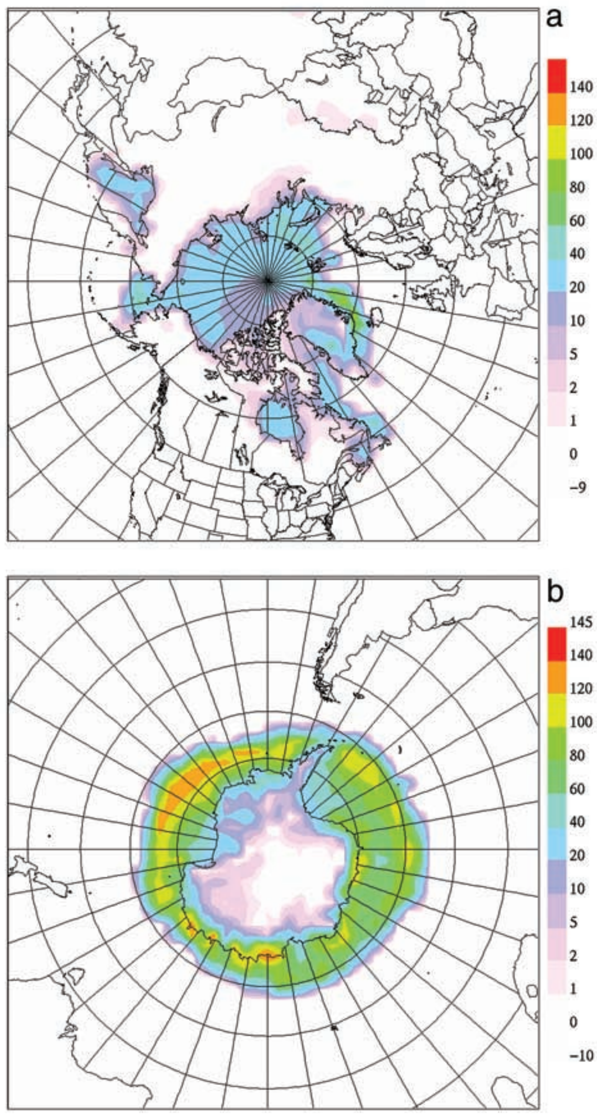

Figure 5. The mean annual blowing snow sublimation rate (mm swe) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere.

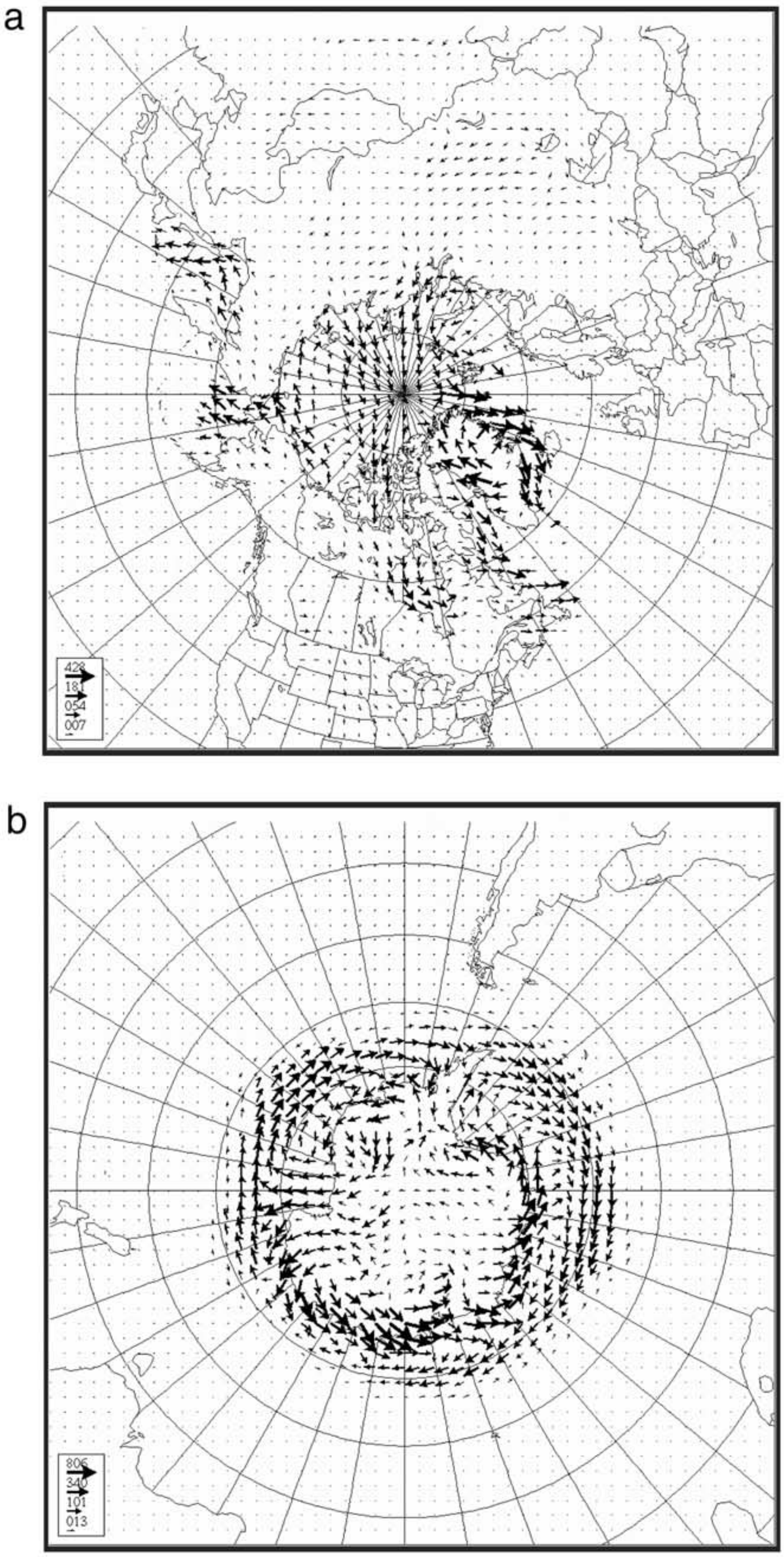

Figure 6. The mean annual blowing snow mass transport vectors (Mg m^{-1}) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere. Note the varying scales in each plot.

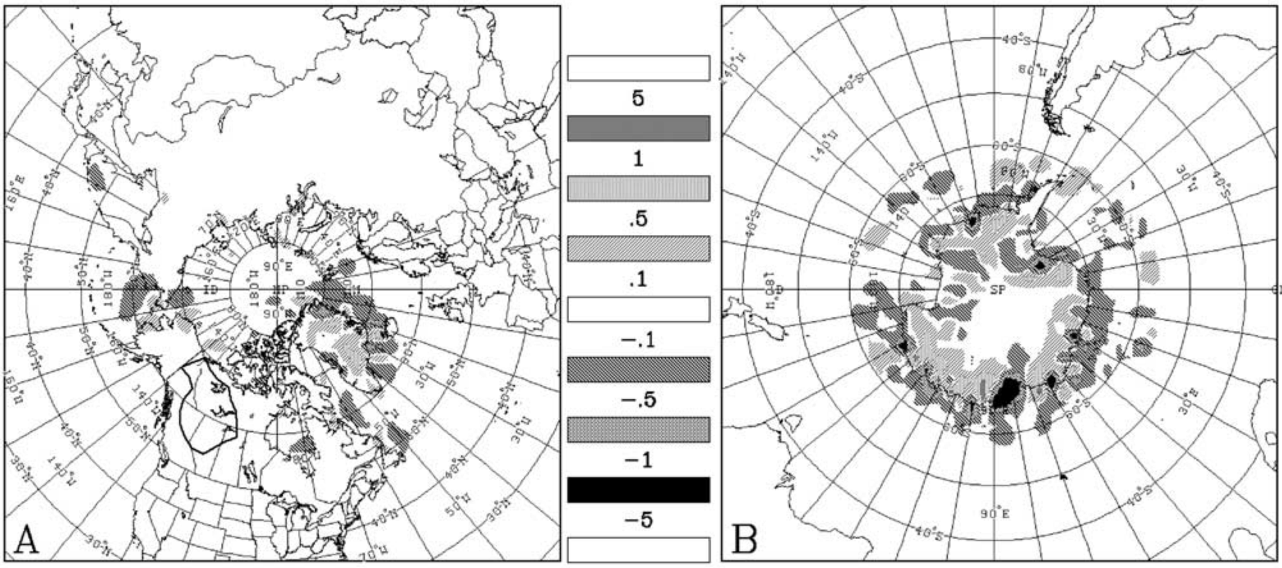

Figure 7. The mean annual blowing snow divergence rate (mm swe) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere. The bold outline in Figure 7a represents the Mackenzie River Basin.

100 Mg m^{-1} yr^{-1} or more. The mass transport vectors also reveal some of the persistent climatological features in both hemispheres. For instance, on the eastern shores of Greenland, strong, northeasterly winds driven by the climatological Icelandic low force massive amounts of snow over the open waters of the North Atlantic Ocean. In western Greenland, winds have a strong southerly component and displace snow in a poleward direction. On the other hand, the appearance of the circumpolar trough and the dominance of high pressure systems in the interior of Antarctica clearly emerges in the Southern Hemisphere [*King and Turner*, 1997]. The mass transport vectors are generally higher in magnitude in the Southern Hemisphere than in the Northern Hemisphere, especially in the windy coastal areas of Antarctica (Figure 6b).

[27] Computing the mass divergence from these transport vectors, we may then evaluate the net contribution of snow transport to the local water budget (term D in equation 1). Figure 7 demonstrates that in the Northern Hemisphere, a single prominent zone of mass divergence is inferred over central Greenland with the loss of up to 1 mm yr^{-1} swe due to blowing snow. Over Antarctica, katabatic winds tend to remove mass from the interior regions of the continent and displace it to coastal areas or over sea ice.

[28] Figure 8 illustrates the combined large-scale effects of surface sublimation and blowing snow to the global surface mass balance. In both hemispheres, these terms have the largest impact over sea ice and the nearby coastal regions where their sum attains values of 150 mm yr^{-1} swe or more. In comparison, continental areas experience lesser effects from these cold climate processes with typical contributions of blowing snow and surface sublimation combining to erode 50 mm yr^{-1} swe at most. A summary of the contribution of the surface sublimation and blowing snow processes to the large-scale surface mass balance is presented in Table 5 for selected areas of interest. Areally averaged over zonal bands of 10°, we find strong latitudinal variations in Q_{surf} and Q_s, with these terms combining to dispose about 134 mm yr^{-1} swe between 60 and 70°S (Table 5). In the Northern Hemisphere, surface sublimation is the dominating process removing mass near the pole as winds are generally lower here than in Antarctic coastal areas. From a global perspective, about 70% of the annual exchange of water between the surface in the ice phase and the atmosphere in the gas phase is attributed to surface sublimation with the remainder comprising of blowing snow sublimation.

[29] Figure 9 reveals that, when zonally-averaged, Q_{surf} and Q_s both approach peak values of ≈100 mm yr^{-1} swe at 67.5°S in the Southern Hemisphere. In the Northern Hemisphere, on the other hand, Q_{surf} and Q_s attain their zonally-averaged maximum value of 142 mm yr^{-1} swe at 87.5°N and of 24 mm yr^{-1} swe at 77.5°N, respectively. The divergence of mass through wind transport is generally 2 orders of magnitude less than both Q_{surf} and Q_s in all regions (Table 5) and is therefore not included in Figure 9.

4.2. Results for the Mackenzie River Basin

[30] Given that this study is motivated by MAGS, we now emphasize our results for the MRB of Canada (outlined in bold in Figure 3a). Of the three processes examined in this study, only surface sublimation contributes significantly to the overall water budget of the MRB (Figure 4a). Blowing snow sublimation reaches values of no more than a few millimeters per year swe along the Arctic coastline and remains negligible throughout most of the basin (Figure 5a). Minimal values of the blowing snow transport and divergence rates also prevail within the boreal forest in association with the reduced wind speeds there (Figures 6a and 7a). 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. Note also that this value is comparable with the sublimation rates reported at

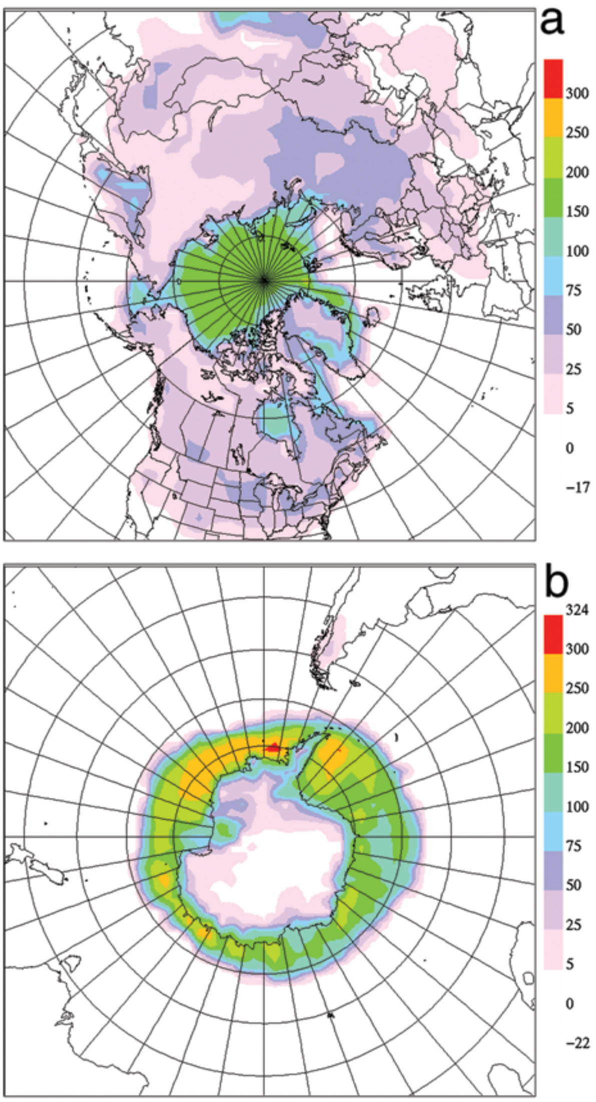

Figure 8. The mean annual total surface sublimation, blowing snow sublimation and divergence rates (mm swe) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere.

four individual meteorological stations within the MRB (see Table 3).

4.3. Temporal Trends

[31] Figure 10 depicts the relative monthly contributions to the annual rates of blowing snow and surface sublimation for Antarctica, the MRB and Greenland. Over the Antarctic and Greenland ice sheets, we observe a strong seasonality in the surface sublimation rates, with summertime fluxes exceeding at times by a factor of 10 those inferred during the winter. In contrast, blowing snow sublimation displays little temporal variation over these ice sheets. In the MRB, both surface and blowing snow sublimation approach zero during summertime as the basin remains nearly void of snow at this time. Peak values of blowing snow (surface) sublimation are achieved during the month of January (April).

[32] An example of the daily surface and blowing snow sublimation rates during one year at a single location is displayed in Figure 11. We have chosen Inuvik, Northwest Territories as our site of interest to allow comparisons with the meteorological time series shown in Figure 1. Observe how the surface sublimation rates exhibit a seasonal cycle representative of the entire MRB (see Figure 10), with maximum values occurring in late spring and early fall and with lesser contributions during winter. During summer, sublimation remains zero as the snowpack ablates completely near Julian day 140. The rare episodes of blowing snow sublimation are, as expected, associated with strong wind events (compare Figure 1).

5. Sensitivity Tests

[33] In this section, we perform two additional experiments in an attempt to gauge the sensitivity of our results to values of RH_i. In section 3.2, we determined that it was necessary to reduce values of RH_i inferred from the ERA15 by 28% for them to match reasonably well the climatological values observed at four sites in the MRB. The bias-corrected ERA15 humidity data were then applied in section 4 to obtain our so-called "control" experiment. In this first sensitivity experiment (labeled ST1), we therefore return to the original ERA15 data set and repeat our surface mass balance study with the uncorrected humidity data. Since blowing snow transport is not affected by these changes, we present here results solely for the surface and blowing snow sublimation rates. Figure 12 shows that the ERA15's cold temperature and dew-point biases lead to a significant reduction in the overall surface sublimation. In fact, an increase of 28% in values of RH_i leads to a decrease of 79% in Q_{surf} over the globe. Hence, the contribution of surface sublimation to the global water budget is reduced from about 11.0 to 2.3 mm yr^{-1} swe. Blowing snow sublimation experiences similar trends when derived from the original

Table 5. Areally Averaged Contribution of Surface Sublimation, Blowing Snow Sublimation and Divergence to the Surface Mass Balance Within Latitudinal Bands of 10° in the Southern and Northern Hemispheres, as Well as for Antarctica, the Mackenzie River Basin, Greenland and the Globe[a]

Region	Q_{surf}	Q_s	D	Sum
		Southern Hemisphere		
80°–90°	19.41	5.498	-1.956×10^{-2}	24.89
70°–80°	59.40	33.46	5.486×10^{-2}	92.91
60°–70°	70.21	63.86	-3.855×10^{-2}	134.0
50°–60°	4.700	7.56	1.953×10^{-2}	12.28
ANT	14.05	15.32	4.294×10^{-2}	29.41
		Northern Hemisphere		
80°–90°	123.3	22.42	3.187×10^{-3}	145.7
70°–80°	70.80	17.21	5.222×10^{-3}	88.04
60°–70°	32.83	5.609	1.621×10^{-3}	38.44
50°–60°	25.35	2.346	-4.456×10^{-3}	27.69
MRB	28.54	0.2318	1.530×10^{-3}	28.77
GRE	50.32	15.53	0.1218	65.97
		Globe		
Globe	10.95	4.405	8.342×10^{-4}	15.36

[a]All units are in mm yr^{-1} swe, with positive (negative) numbers indicating a sink (source) term in the surface mass balance. Q_{surf}, surface sublimation; Q_s, blowing snow sublimation; D, divergence; ANT, Antarctica; MRB, Mackenzie River Basin; and GRE, Greenland.

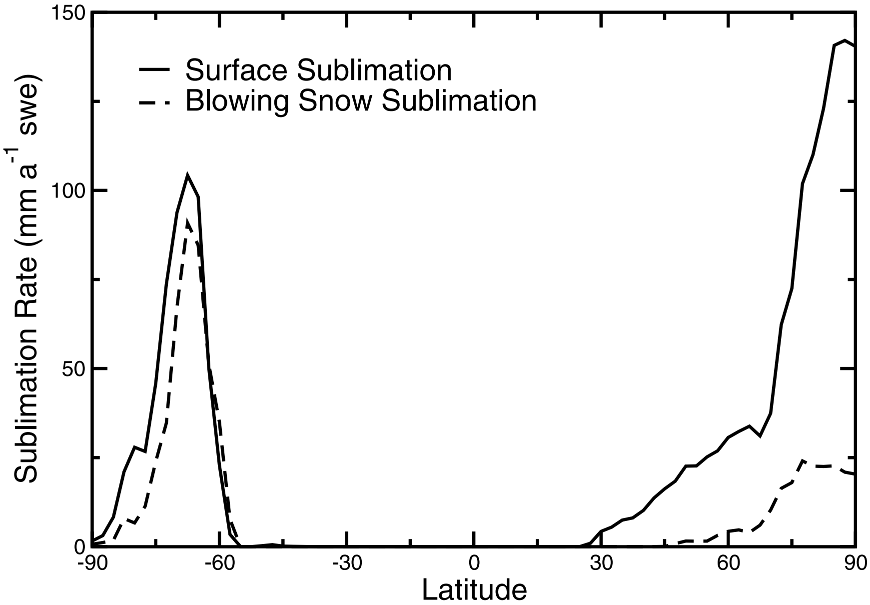

Figure 9. The zonally-averaged surface and blowing snow sublimation rates (mm swe) for the period 1979–1993.

ERA15 data set with a decrease of 64% in the global value of Q_s (Figure 13).

[34] Recent studies by *Bintanja* [2001b] and *Déry and Yau* [2001b] suggest that blowing snow sublimation rates may be enhanced by as much as 80% due to the vertical entrainment and/or horizontal advection of dry air into a column of blowing snow. Although these processes are not known to operate in all environments, we apply here this result in order to obtain a possible upper bound for both surface and blowing snow sublimation. In this second sensitivity test (ST2), the results for Q_{surf} and Q_s shown in Figures 4 and 5, respectively, are simply multiplied by a factor of 1.8 and lead to increased sublimation rates everywhere (not shown).

[35] A summary of the impact of these sensitivity tests on Q_{surf} and Q_s is presented in Table 6 for Antarctica, Greenland, the MRB and the globe. We clearly see that the control experiment provides nearly mid-range values of Q_{surf} and Q_s in all regions. In accordance with the recent findings of *King et al.* [2001], the sublimation rates are extremely

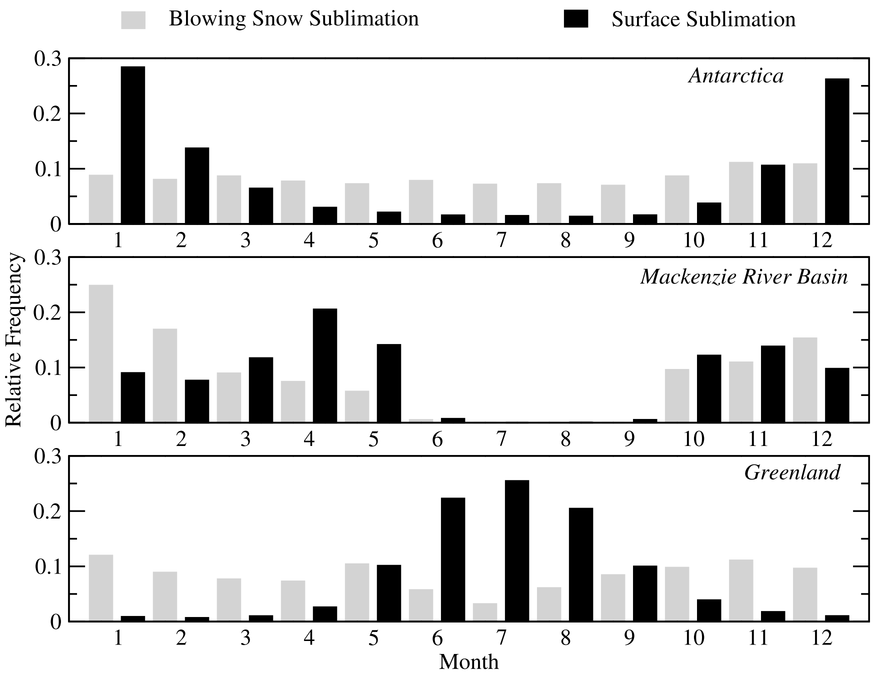

Figure 10. The relative monthly contributions to the annual surface and blowing snow sublimation rates over Antarctica, the Mackenzie River Basin and Greenland for the period 1979–1993.

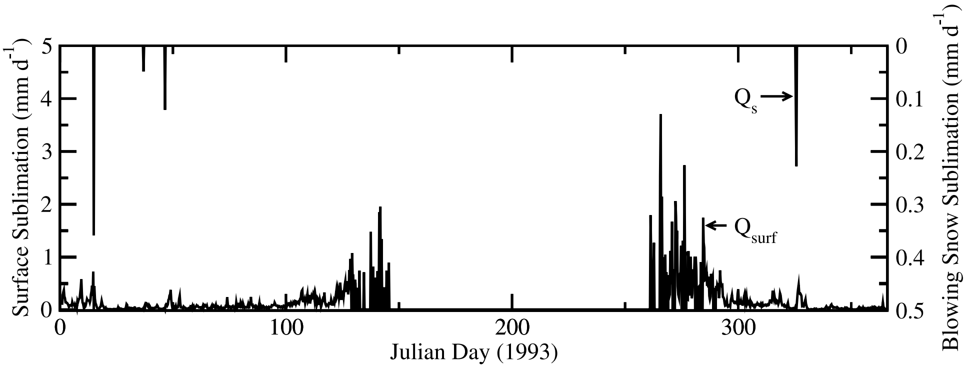

Figure 11. The daily surface (Q_{surf}) and blowing snow (Q_s) sublimation rates inferred from the ERA15 during 1993 at Inuvik, Northwest Territories.

sensitive to the ambient humidity. For instance, increasing values of RH_i by 28% (ST1) induces a decline of 104% in surface sublimation rates over Antarctica.

6. Comparison With Other Studies

[36] In this section, we compare our findings with those from previously published studies. Note that unless stated otherwise, we discuss here our results from the control simulation (section 4), but that additional comparisons with the two sensitivity tests are conducted in the summary tables.

6.1. Southern Hemisphere

[37] The results for surface sublimation over Antarctica presented in Figure 4 qualitatively resemble those of Figure 9 obtained by *van den Broeke* [1997] with a GCM. For instance, both studies display subdued surface sublimation rates over the interior of Antarctica, with increasing values of this process toward the Antarctic coastline and over sea ice. As in *van den Broeke*'s [1997] work, we also capture the very high surface sublimation rates inferred over the Ross Ice Shelf and the katabatic wind regions of eastern Antarctica. Areally-averaged over the continent, we report in Table 7 that the total surface sublimation rate for Antarctica is about 14 mm yr^{-1} swe (in the CTL experiment), a value somewhat less than the 23 to 34 mm yr^{-1} swe estimate of *van den Broeke* [1997]. This discrepancy can be explained in part by *van den Broeke*'s [1997] methodology that also considers evaporation from the surface even at temperatures above 0°C whereas we completely neglect this process. Recall from section 2 that we set Q_{surf} to zero during snow drifting events such that the remainder of this difference can be attributed to the sublimation of blowing snow. According to our estimates, this process removes an additional 15 mm yr^{-1} swe from the surface mass balance of Antarctica. Combined, we find that these two processes erode about 29 mm yr^{-1} swe from the Antarctic continent, a value well within the range reported by *van den Broeke* [1997] (compare his Figure 9 with our Figure 8). Note also that our estimate of blowing snow sublimation for Antarctica matches very well the 17 mm yr^{-1} swe documented by *Bintanja* [1998], although it is significantly less than the value of 38 mm yr^{-1} swe achieved by *Gallée* [1998].

[38] Furthermore, the total contribution of surface sublimation and blowing snow to the surface mass balance inferred from the original ERA15 (ST1 simulation) in this study are less than the evaporation (essentially the sublimation) rates found by *Bromwich et al.* [1998], *Turner et al.* [1999] and *Genthon and Krinner* [2001] using the same data set for Antarctica. Specifically, they report the erosion of 21, 8 and 10 mm yr^{-1} swe respectively for Antarctica, whereas our calculations show that surface and blowing snow sublimation volatilize about 6 mm yr^{-1} swe over the same area (Table 7). This is in contrast to our estimates of these terms when the bias-corrected ERA15 measurements are applied since we then obtain a total value of 30 mm yr^{-1} swe for Antarctic surface and blowing snow sublimation. Using the ERA15 precipitation totals reported by *Bromwich et al.* [1998], *Turner et al.* [1999], and *Genthon and Krinner* [2001], we then estimate that blowing snow and surface sublimation dispose between 17 to 20% of the annual precipitation over Antarctica. Note that the disparities for values of P and Q_{surf} derived from the same data set arise in part due to the definition of the Antarctic land mask (with or without the grounded ice shelves) and the methodology adopted to obtain these estimates. In any case, the ERA15's high RH_i bias combined with the neglect of blowing snow processes can lead to significant underestimates of evaporative fluxes at continental scales.

[39] We have shown in addition that, as expected, regions subject to frequent blowing snow events are also susceptible to substantial mass transport. *Giovinetto et al.* [1992] compute a net transport of about 120 $\times$ 10^{12} kg yr^{-1} of mass northward across 70°S through blowing snow. We estimate from our calculations the displacement of about 1 $\times$ 10^{12} kg yr^{-1} across 70°S, about 2 orders of magnitude lower than that provided by *Giovinetto et al.* [1992]. These differences can be explained in part by their use of a katabatic wind model to drive their transport fluxes versus our use of the ERA15 data set that exhibits lower wind speeds than observed in Antarctica [*Déry and Yau*, 1999a]. In addition, the high dependence of the mass transport equation on the wind speed means that slight disparities in the wind speed can yield substantial differences in the transport estimates (equation (9)). Finally, *Giovinetto et al.* [1992] assume that all mass transport at 70°S is equatorward although our results show that the mass transport vectors contain a large zonal component at this latitude (see Figure 6b). Our divergence computations for the Antarctic continent show the removal of no more than 0.005 mm yr^{-1}

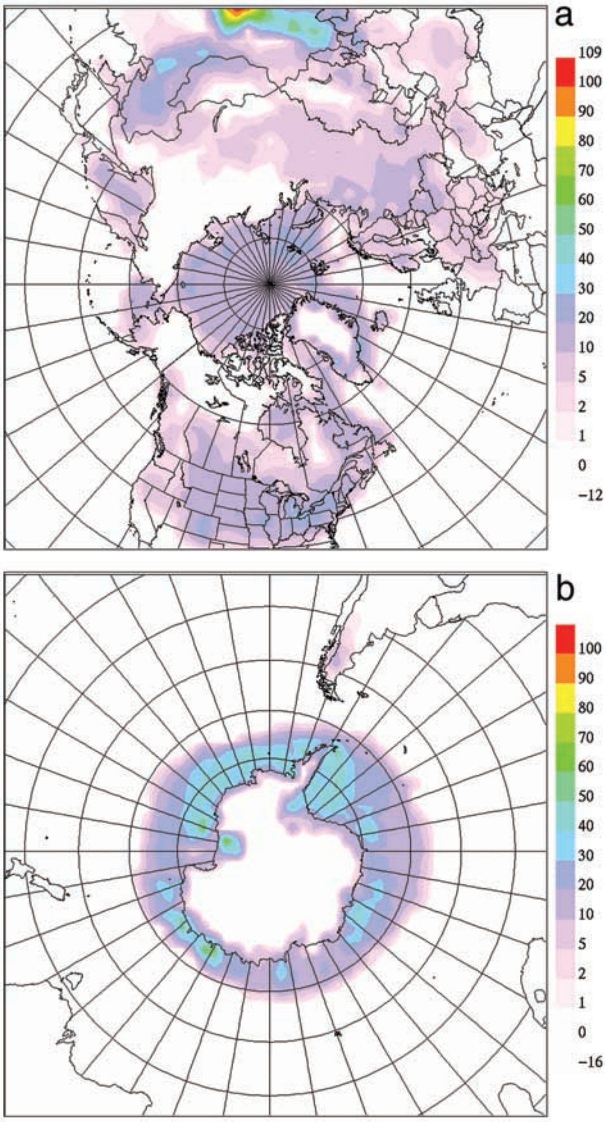

Figure 12. The mean annual surface sublimation rate (mm swe) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere as inferred from the original ERA15 data set.

swe off the continent through this process. Averaging values of D over wide areas tends to reduce the impact of blowing snow transport to the mass balance as regions of mass convergence will oppose trends for divergent areas of mass. As stated by several other authors [e.g., *Connolley and King*, 1996; *Turner et al.*, 1999], mass divergence through wind transport can be neglected for large-scale mass balance budgets of Antarctica.

6.2. Northern Hemisphere

[40] In the Northern Hemisphere, there are relatively fewer large-scale, cold season mass balance studies to validate our own research. Fortunately, the works recently published by *Walsh et al.* [1994, 1998] and *Cullather et al.* [2000] have begun to fill this gap in the literature. However, to our knowledge, none have examined the large-scale impacts of blowing snow to the surface mass balance in the boreal polar region. One possible exception is a monograph by *Mikhel' et al.* [1971] that presents volume transport estimates of snow in Russia. As opposed to our findings in Antarctica, our calculations of blowing snow transport are approximately 1 order of magnitude too large when compared to this Russian study. Furthermore, the mass balance studies of *Walsh et al.* [1994, 1998], *Serreze and Hurst* [2000] and *Cullather et al.* [2000] provide some detailed information about the spatial distribution of precipitation and evaporation at high northern latitudes, but lack explicit information on the individual roles of surface sublimation or blowing snow in the water budget of these areas.

[41] As part of the Surface Heat Budget of the Arctic Ocean (SHEBA) experiment [*Uttal et al.*, 2002], extensive meteorological measurements were conducted over the frozen Arctic Ocean. *Andreas et al.* [2002] analyzed the

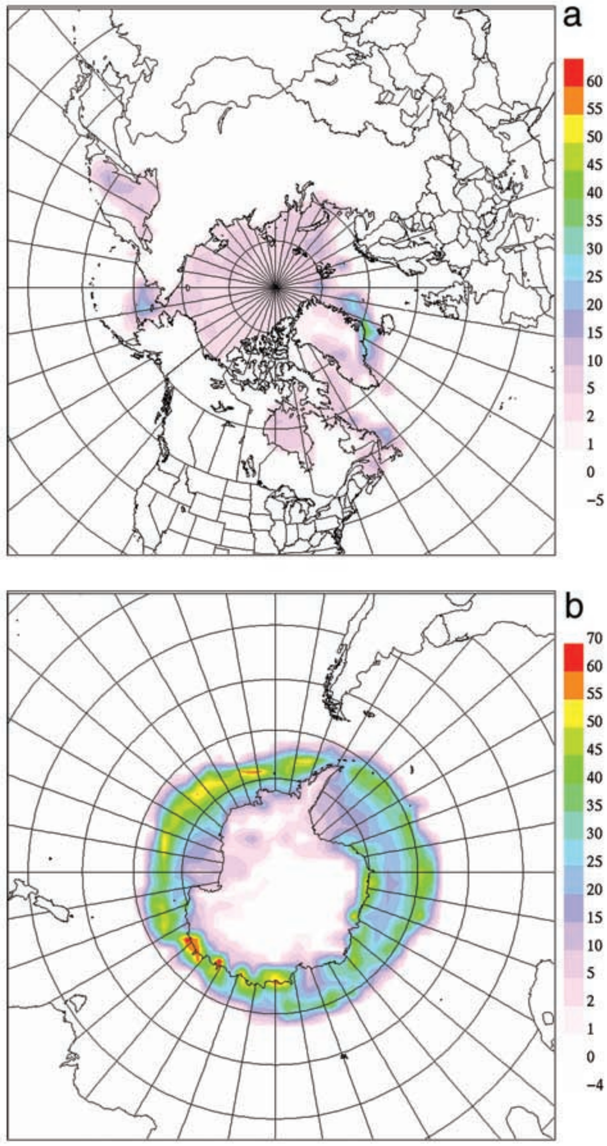

Figure 13. The mean annual blowing snow sublimation rate (mm swe) for the period 1979–1993 (a) in the Northern Hemisphere and (b) in the Southern Hemisphere as inferred from the original ERA15 data set.

Table 6. Values of the Surface and Blowing Snow Sublimation Rates Areally Averaged Over Antarctica, the Mackenzie River Basin, Greenland, and the Globe for the Control Experiment and Two Sensitivity Tests[a]

Location	CTL	ST1	ST2
		Surface Sublimation	
ANT	14.05	−0.510 (−104%)	25.29 (+80%)
MRB	28.54	0.400 (−99%)	51.37 (+80%)
GRE	50.32	4.903 (−90%)	90.58 (+80%)
Globe	10.95	2.298 (−79%)	19.71 (+80%)
		Blowing Snow Sublimation	
ANT	15.32	5.911 (−61%)	27.58 (+80%)
MRB	0.2318	0.06656 (−71%)	0.4172 (+80%)
GRE	15.53	4.006 (−74%)	27.95 (+80%)
Globe	4.405	1.570 (−64%)	7.929 (+80%)

[a] Values of surface and blowing snow sublimation rates are given in mm yr^{-1} swe. Departures (expressed as percentages) from the control experiment are listed in parentheses. CTL, control experiment; ST1 and ST2, the two sensitivity tests, discussed in the text; ANT, Antarctica; MRB, Mackenzie River Basin; GRE, Greenland.

humidity data collected during SHEBA and determined that RH_i values approach consistently the ice saturation point. *Andreas et al.* [2002] suggest that these conditions arise due to the presence of leads in the Arctic Ocean. These small areas of open waters promote much larger latent heat fluxes than over snow-covered sea ice (by 2 orders of magnitude), thereby saturating the low-level Arctic air. Given that leads are subgrid-scale phenomena in the ERA15 data set, it is not surprising that our results show lesser values of moisture than those observations collected during SHEBA.

6.3. Mackenzie River Basin

[42] For the MRB as a whole, we 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 [*Déry and Yau*, 1999a]. Only sections of the MRB lying in the Arctic tundra are susceptible to frequent blowing snow events and hence may lose significant mass through this process. In contrast, surface sublimation is found to be quite important across the entire basin. This is in opposition to the research conducted by *Walsh et al.* [1994] who utilized 18 years of rawinsonde observations 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 in the MRB for the period October to April for a total of about −45 mm swe (Table 8). In contrast, our estimates of surface sublimation for the same basin are somewhat less than those of *Betts and Viterbo* [2000] who establish the evaporation from the snow surface to be on average 92 mm yr^{-1} swe for the same area over two years using the ECMWF forecast model. However, the authors state that their study suffers from 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. In the CTL experiment, we determine that blowing snow and surface sublimation erode $\approx$29 mm yr^{-1} swe over the entire MRB, a value obviously more in line with the results of *Betts and Viterbo* [2000] than those of *Walsh et al.* [1994].

7. Discussion

[43] We have demonstrated in section 4 that the ERA15 data set, corrected for a high relative humidity bias, provides useful information for the detection of blowing snow and surface sublimation events and their impact on the large-scale surface mass balance. Nevertheless, we remind the reader some of the uncertainties (previously alluded to in section 3) that exist in our results. The principal factors of concern (and their possible impact on our current findings, listed in parentheses) are as follows: (1) 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); (2) the usage of the blowing snow parameterizations at the scales of the ERA15 data set and that are not constrained by direct observations (changes to all blowing snow fluxes); (3) 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 (e.g., the PBSM), in part due to the appearance of negative thermodynamic feedbacks on the phase change process (a reduction of

Table 7. Values of the Surface and Blowing Snow Sublimation Rates Over Antarctica From Recent Studies Compared to Our Results for the Control Experiment and Two Sensitivity Tests[a]

Study	Method[b]	Period	Q_{surf}	Q_s
Bintanja [1998]	BSM + OBS	1987	. . .	17
Bromwich et al. [1998]	ERA	1979–1993	8	. . .
Gallée [1998]	BSM + MAM	3 days	. . .	38
Genthon and Krinner [2001]	EOA	5/1985–4/1991	16	. . .
	ERA	1979–1993	10	. . .
	GCM	1979–1988	12–68	. . .
Turner et al. [1999]	ERA	1979–1993	21	. . .
van den Broeke [1997]	GCM	5 years	23–34	. . .
This study (CTL)	ERA + BSM	1979–1993	14.1	15.3
This study (ST1)	ERA + BSM	1979–1993	−0.5	5.9
This study (ST2)	ERA + BSM	1979–1993	25.3	27.6

[a] Values of surface and blowing snow sublimation rates are given in mm yr^{-1} swe. Q_{surf}, surface sublimation rate; Q_s, blowing snow sublimation rate; CTL, control experiment; and ST1 and ST2, two sensitivity tests, described in the text.

[b] The following abbreviations are used for the methods: BSM, blowing snow model; OBS, meteorological observations; ERA, European Centre for Medium-Range Weather Forecasts Reanalysis; EOA, ECMWF operational forecasts; MAM, mesoscale atmospheric model; and GCM, general circulation model.

Table 8. Values of the Surface and Blowing Snow Sublimation Rates Over the Mackenzie River Basin From Recent Studies Compared to Our Results for the Control Experiment and Two Sensitivity Tests[a]

Study	Method[b]	Period	Q_{surf}	Q_s
Betts and Viterbo [2000][c]	EOA	1996–1998	37	...
Walsh et al. [1994]	OBS	1973–1990	−45	...
This study (CTL)	ERA + BSM	1979–1993	28.5	0.2
This study (ST1)	ERA + BSM	1979–1993	0.4	0.1
This study (ST2)	ERA + BSM	1979–1993	51.4	0.4

[a] Values of surface and blowing snow sublimation rates are given in mm yr^{-1} swe. Q_{surf}, surface sublimation rate; Q_s, blowing snow sublimation rate; CTL, control experiment; and ST1 and ST2, two sensitivity tests, described in the text.

[b] The following abbreviations are used for the methods: BSM, blowing snow model; OBS, meteorological observations; ERA, European Centre for Medium-Range Weather Forecasts Reanalysis; and EOA, ECMWF operational forecasts.

[c] Reduced by 60% from its original value to account for a bias in their model.

Q_s); (4) the neglect of blowing snow in the ECMWF operational model and, hence, of the thermodynamic feedbacks associated with blowing snow sublimation in the ERA15 data set (an increase in Q_s and Q_{surf}); (5) the exclusion of surface sublimation during blowing snow events (a decrease in Q_{surf}); (6) the assumption of homogeneous surfaces and the neglect of blowing snow interception by vegetation (an increase in Q_t); (7) the omission of surface sublimation from a forest canopy and when air temperatures exceed the freezing point (a decrease in Q_{surf}); (8) the sensitivity of surface and blowing snow sublimation rates on the ambient humidity (underestimation or overestimation of Q_s and Q_{surf}, depending on the accuracy of RH_i values); (9) the assumption of neutral stability in the atmospheric boundary layer (changes to Q_{surf}); (10) the uncertainty in the effects of entrainment and/or advection of dry air on sublimation processes (an underestimation of Q_s and Q_{surf}).

[44] Considering that this is the first comprehensive attempt at establishing the large-scale impact of blowing snow and surface sublimation to the large-scale surface mass balance, it is possibly the best estimate yet available on the importance of these terms. The points raised above demonstrate, however, that large uncertainties remain in closing the water budget at high latitudes and that more research is required to ascertain (or improve) the results presented herein.

8. Summary

[45] We have conducted a surface mass balance study of the polar regions with an emphasis on several cold season processes. The water budgets were computed using the ERA15 data for the years 1979–1993 inclusive at a horizontal resolution of 2.5°. The results demonstrate that both surface (11 mm yr^{-1} swe) and blowing snow sublimation (4 mm yr^{-1} swe) are two important processes on global scales. On smaller scales, the Antarctic continent and the MRB lose mass on the order of 29 mm yr^{-1} swe through these processes, disposing 17–20% and 7% respectively of the total annual precipitation over these regions. However, the importance of surface and blowing snow sublimation vary significantly from region to region. The divergence of mass associated with blowing snow transport, on the other hand, contributes negligibly to the large-scale water budget of most areas.

[46] As demonstrated by two sensitivity tests and some of our underlying assumptions, the results provide only a first-order estimate of the contributions of these terms to the surface mass balance. Blowing snow transport, for instance, is known to displace locally significant amounts of mass into large snowdrifts whereas other areas remain nearly void of snow [e.g., *Prasad et al.*, 2001; *Sturm et al.*, 2001]. Nonetheless, as we approach the full closure of high latitude water budgets, this contribution provides a first step in establishing the large-scale climatological impacts of surface sublimation, blowing snow sublimation and divergence to the mass balance of polar regions.

[47] **Acknowledgments.** We wish to acknowledge the contributions of Graham Mann (University of Leeds), John King (British Antarctic Survey), Michiel van den Broeke (University of Utrecht), David Bromwich (Ohio State University), Edgar Andreas (CRREL, Hanover) and Marc Stieglitz (Columbia University) during the preparation of this work. Comments by Matthew Sturm (CRREL, Fort Wainwright) and two anonymous referees also greatly improved the text. This research was supported by the Natural Sciences and Engineering Research Council of Canada through a GEWEX collaborative research network grant.

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