A BULK BLOWING SNOW MODEL

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Abstract. We present in this paper a simple and computationally efficient numerical model that depicts a column of sublimating, blowing snow. This bulk model predicts the mixing ratio of suspended snow by solving an equation that considers the diffusion, settling and sublimation of blowing snow in a time-dependent mode. The bulk model results compare very well with those of a previous spectral version of the model, while increasing its computational efficiency by a factor of about one hundred. This will allow the use of the model to estimate the effects of blowing snow upon the atmospheric boundary layer and to the mass balance of such regions as the Mackenzie River Basin of Canada.

Keywords: Blowing snow, Bulk modelling, Mackenzie Basin, Sublimation, Suspension.

1. Introduction

In addition to its hazardous aspects, the transport of snow and its sublimation are being recognized as important factors in the water and energy budgets of windswept regions such as the Arctic tundra. Erosion and accumulation of snow by wind can lead to substantial heterogeneities in the snowcover with hydrometeorological implications. In addition to supplying moisture to the atmospheric boundary layer (ABL), sublimation of blowing snow acts as a further sink of mass from the surface that can lead to erroneous estimates of the snow depth in numerical models that neglect these processes.

This work is conducted within the context of the ongoing Mackenzie GEWEX Study (MAGS; Stewart et al., 1998) since part of the territory drained by the Mackenzie River is susceptible to frequent blowing snow events (Déry and Yau, 1999). One of our future goals is to assess the contribution of blowing snow transport and sublimation to the water budget in this large northern drainage basin. Accurate and efficient numerical modelling of blowing snow is thus necessary in this context.

We therefore introduce in this paper a bulk version of a blowing snow model that provides the thermodynamic feedbacks of sublimating, blowing snow in the ABL while keeping track of its effects on the snowcover. The numerical model discussed here is based on the PIEKTUK blowing snow model of Déry et al. (1998) and the initial work of Déry and Taylor (1996) on blowing snow. The novelty of the bulk version is that it is about 100 times faster that the spectral version but yields comparable results. Since the spectral PIEKTUK model is extensively described

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in Déry et al. (1998), we first present only a brief review of its formulation before examining steps that lead to a so-called "bulk" version of the model. Subsequently, results, some sensitivity tests and a concluding discussion will be presented.

2. Model Description

2.1. Spectral model – (PIEKTUK-S)

The PIEKTUK model of Déry et al. (1998) is spectral in nature in that it depicts explicitly, in either a time or fetch-dependent framework, a spectrum of blowing snow particles that are gamma-distributed at the lower model boundary and are suspended through diffusion from a saltation layer just above the snow-covered surface. The spectral number density of suspended particles $F(r, z, t)$ (m^{-4}) for particles of radius r (m) is taken to satisfy:

$$\frac{\partial F(r)}{\partial t} = \frac{\partial}{\partial z}\left(K(r)\frac{\partial F(r)}{\partial z} + v(r)F(r)\right) - \frac{\partial}{\partial r}\left(\dot{r}F(r)\right), \tag{1}$$

for the time-dependent, horizontally homogeneous case. Here, time is denoted by t (s), the vertical coordinate by z (m), and $\dot{r}$ denotes the rate of change of radius due to sublimation. We also abbreviated $F(r, z, t)$ by $F(r)$. Three active processes are depicted in the rhs of Equation (1): the vertical diffusion of blowing snow particles with eddy diffusivity $K(r)$ ($\mathrm{m}^2\ \mathrm{s}^{-1}$), the sedimentation of particles with a terminal velocity $v(r)$ ($\mathrm{m\,s}^{-1}$) obtained through a balance between the gravitational and drag forces (Déry et al., 1998), and the spectral shifting due to sublimation of blowing snow $\frac{\partial}{\partial r}(\dot{r}F(r))$ ($\mathrm{m}^{-4}\ \mathrm{s}^{-1}$).

Following the work of Rouault et al. (1991), the turbulent diffusion coefficient used by Déry et al. (1998) considered a reduction ζ from the eddy diffusivity for momentum K_m ($\mathrm{m}^2\ \mathrm{s}^{-1}$) due to the inertia of the particles such that:

$$K(r) = \zeta K_m. \tag{2}$$

(Note that a detailed discussion of the interaction between the spectrum of particles and turbulence can also be found in Lee (1975).)

Values of ζ are size-dependent and are discussed at length by Déry et al. (1998). In near-neutral conditions, we may assume:

$$K_m = u_* l\,, \tag{3}$$

with the mixing length l (m) given by (e.g., Stull, 1988):

$$l = \kappa(z + z_0)\left[1 + \kappa(z + z_0)/l_{max}\right]^{-1}. \tag{4}$$

In these equations, u_* (m s^{-1}) is the friction velocity, κ (= 0.4) is the von Kármán constant, and z_0 (m) the roughness length. Following Taylor (1969) and Mobbs and Dover (1993), an asymptotic value to l of 40 m is represented by l_{max}.

Neglecting for the moment the appropriate boundary conditions that are required to solve Equation (1), the above three relationships describe, in part, the suspension of blowing snow particles through the competing processes of diffusion, settling and sublimation in the spectral version of PIEKTUK (hereafter referred to as PIEKTUK-S). Déry et al. (1998) typically used 64-particle size bins to evaluate the blowing snow suspension and sublimation rates. For computational simplicity, we therefore investigate the possibility of using a bulk quantity, namely the blowing snow mixing ratio q_b (kg kg^{-1}), which is the ratio of the mass of suspended ice particles to that of dry air, to depict the amount of snow in suspension. The steps leading to this bulk approach are described in the following section.

2.2. Bulk model - (PIEKTUK-B)

2.2.1. *Formulation*

For the sake of computational efficiency, complex microphysical schemes have commonly employed a bulk method to derive certain hydrometeor species in atmospheric models (e.g., Kong and Yau, 1997). We apply here a similar technique for the computation of blowing snow suspension and sublimation.

Following Schmidt (1982) and others, we assume that blowing snow is composed of ice spheres such that the mixing ratio of blowing snow, q_b, can be related to the number density by

$$q_b = \frac{4\pi\rho_{ice}}{3\rho}\int_0^\infty r^3 F(r)\,\mathrm{d}r, \tag{5}$$

with ρ_{ice} (= 900 kg m^{-3}) and ρ (kg m^{-3}) denoting the constant densities of ice and air, respectively.

Multiplying Equation (1) by $(4\pi\rho_{ice}r^3/3\rho)$, followed by an integration with respect to r from 0 to ∞ and applying (5), we obtain

$$\frac{\partial q_b}{\partial t} = \frac{\partial}{\partial z}\left(K_b\frac{\partial q_b}{\partial z} + v_b q_b\right) + S_b, \tag{6}$$

where K_b (m^2 s^{-1}) is some bulk diffusion coefficient and v_b (m s^{-1}) is some bulk fall velocity. The sink in q_b due to the sublimation of blowing snow, S_b (kg kg^{-1} s^{-1}) is discussed in Section 2.2.2. To obtain v_b and K_b, we need to know the exact form of the number density function.

We begin by noting that Budd (1966) and Schmidt (1982) have showed that distributions of $F(r)$ typically follow those of a two-parameter gamma distribution such that:

$$F(r) = \frac{N r^{(\alpha-1)}\exp^{-r/\beta}}{\beta^\alpha\Gamma(\alpha)}, \tag{7}$$

with N (m^{-3}) being the total number concentration of particles, α (dimensionless) and β (m) the shape and scale parameters of the gamma distribution Γ.

Substituting Equation (7) into (5), integrating and solving for N, we get:

$$N = \frac{3\rho q_b \Gamma(\alpha)}{4\pi \rho_{ice} \Gamma(\alpha + 3)\beta^3}. \tag{8}$$

Several tests with PIEKTUK-S revealed that α varies little with height and is thus taken as constant (set to 2 following the analysis of King et al., 1996). Using α = 2 in (8), we can solve for β as

$$\beta = \frac{1}{2}\left[\frac{\rho q_b}{4\pi \rho_{ice} N}\right]^{1/3} . \tag{9}$$

We now approximate N in Equation (9) by a special solution N_s of Equation (1), as follows. For a steady-state, saturated environment (i.e., no sublimation), we may write:

$$K(r)\frac{\partial F(r)}{\partial z} = -v(r)F(r). \tag{10}$$

Integrating this equation from the top of the saltation layer z_s, assuming $l \approx \kappa(z + z_0)$ and neglecting any inertial effects, we retrieve the classical equation for suspended particle concentrations:

$$F(r, z) = F(r, z_s)\left[\frac{(z + z_0)}{(z_s + z_0)}\right]^{-v(r)/\kappa u_*} . \tag{11}$$

Integrating (11) from r = 0 to ∞ and assuming that $F(r)$ is given by a gamma distribution, we obtain

$$N_s = \int_0^\infty F(r, z_s)\left[\frac{(z + z_0)}{(z_s + z_0)}\right]^{-v(r)/\kappa u_*} \mathrm{d}r . \tag{12}$$

We now set $N = kN_s$ in (9). It is found that k = 3 gives the best agreement between the solutions of the spectral and the bulk models.

As a final measure, we need to specify a bulk terminal velocity that characterizes the particle distribution and which will vary with height. Applying the method of Kong and Yau (1997), we get:

$$v_b = \frac{\int_0^\infty v(r) r^n F(r)\,\mathrm{d}r}{\int_0^\infty r^n F(r)\,\mathrm{d}r}, \tag{13}$$

where n is a moment of the gamma distribution. As in Kong and Yau (1997), we tried setting n equal to 3 such that v_b represents the mass-weighted terminal velocity of the ice particle distribution. However, this approach predicts q_b profiles that are consistently too high compared to the results of PIEKTUK-S. After a number of tests, we find that the fifth moment of the distribution yields better approximations for q_b in PIEKTUK-B and, for this reason, set $n = 5$. These initial steps, which use information on the assumed gamma-distributed spectra of blowing snow particles, now allow us to proceed with the discussion of a bulk blowing snow model.

2.2.2. *Diffusion and Sublimation*

As mentioned previously, an alternative to the representation of the amount of suspended snow is the bulk quantity q_b, the blowing snow mixing ratio, governed by Equation (6). Note that although Equation (2) includes a reduction in the eddy diffusivity of the particles due to their inertia, we assume for the moment that ζ is unity such that $K_b = K_m$ as in Bintanja (1998a) and depicts the eddy diffusivity for q_b.

To conserve heat and moisture in the column of sublimating, blowing snow, we introduce two additional prognostic equations in the model for the ambient air temperature T_a (K) and water vapour mixing ratio q_v (kg kg^{-1}), which satisfy:

$$\frac{\partial T_a}{\partial t} = \frac{\partial}{\partial z}\left(K_h \frac{\partial T_a}{\partial z}\right) + Q \tag{14}$$

and

$$\frac{\partial q_v}{\partial t} = \frac{\partial}{\partial z}\left(K_v \frac{\partial q_v}{\partial z}\right) + E, \tag{15}$$

with K_h and K_v being the heat and moisture eddy diffusivities, taken as equal to that for momentum (Equation (3)). The source term for water vapour, E (kg kg^{-1} s^{-1}), is influenced by the sublimation process only, and therefore is set equal to $-S_b$. The heating rate (negative here) due to sublimation is represented by Q (K s^{-1}) in Equation (14) and is computed from:

$$Q = \frac{S_b L_s}{c_p} \tag{16}$$

with the latent heat of sublimation and heat capacity for air denoted by L_s (J kg^{-1}) and c_p (J kg^{-1} K^{-1}), respectively. We neglect here additional heat from the particles as this component contributes negligibly to the phase change. Note however that both T_a and q_v are not subject to settling as is q_b such that we expect greater vertical redistribution of heat and moisture compared to ice particles.

The sublimation term S_b (kg kg^{-1} s^{-1}) is derived as follows. Ignoring any radiation transferred to a particle, the change in mass m (kg) of a single ice sphere due to sublimation is obtained through (Thorpe and Mason, 1966):

$$\frac{\mathrm{d}m}{\mathrm{d}t} = \frac{2\pi r Nu (q_v/q_{is} - 1)}{(F_k + F_d)}, \tag{17}$$

where q_{is} (kg kg^{-1}) denotes the saturation water vapour mixing ratio with respect to ice, Nu represents the Nusselt number and where the conduction and diffusion terms involved in the phase change are respectively given by F_k and F_d (m s kg^{-1}) (Rogers and Yau, 1989).

To obtain the total sublimation rate for a spectrum of particles, we multiply Equation (17) with the particle number concentration and perform an integration over all radii, i.e.

$$S_b = \frac{1}{\rho}\int_0^\infty F(r)\,\frac{\mathrm{d}m}{\mathrm{d}t}\,\mathrm{d}r. \tag{18}$$

Assuming once again that the particles follow a gamma distribution (see Equation (7)), the integral yields the bulk sublimation rate:

$$S_b = \frac{q_b Nu(q_v/q_{is} - 1)}{2\rho_{ice} r_m^2 (F_k + F_d)}, \tag{19}$$

where the mean radius of the particle distribution, r_m (m), is defined as:

$$r_m = \frac{\int_0^\infty r F(r)\,\mathrm{d}r}{\int_0^\infty F(r)\,\mathrm{d}r} = \alpha\beta. \tag{20}$$

Ventilation effects due to the settling of suspended particles are in effect introduced by Nu, which is dependent on the Reynolds number Re through (Lee, 1975):

$$Nu = 1.79 + 0.606 Re^{0.5}, \tag{21}$$

where in still air

$$Re = \frac{2 r_m v_b}{\nu}, \tag{22}$$

with ν being the kinematic viscosity of air (1.53×10^{-5} m^2 s^{-1}). Thus to compute the diffusion and sublimation of blowing snow, it is clear that both r_m and v_b must be known quantities.

2.2.3. *Boundary and Initial Conditions*

Setting proper initial and spatial boundary conditions is crucial to the modelling of blowing snow. Of particular difficulty here is assigning the lower boundary conditions on humidity and the blowing snow mixing ratio and hence, the particle distribution, (note that the boundary conditions discussed here are applied only when snow transport is predicted to occur). At the present time, we take the lower boundary height z_{lb} to be at the snow surface and the model lid z_{ub} to be 1 km above the surface where fluxes of particles, temperature and moisture are assumed zero.

At the lower boundary, we take the air to be saturated with respect to ice for the control case. However, supplementary experiments will be conducted to test the sensitivity of this critical assumption. Given a value of relative humidity with respect to ice RH_i at a certain level and neglecting stability effects, the vertical variation of humidity is deduced from a logarithmic profile (e.g., Garratt, 1992) as:

$$q_v = q_{is} + \frac{q_*}{\kappa} \ln\left[\frac{(z + z_0)}{z_0}\right] \tag{23}$$

where q_* (kg kg^{-1}) represents the humidity scale. For temperature, however, we assume initially no variation with height and that there is no heat flux at z_{lb}.

At the onset of blowing snow, we assume that the saltation layer instantaneously develops, but that no particles are in suspension above z = 0.1 m, the first level above z_s. Thus we set a constant value for the saltation blowing snow mixing ratio $q_{b_{salt}}$ (kg kg^{-1}) in the saltation layer (Pomeroy et al., 1993):

$$q_{b_{salt}} = 0.385(1 - U_t/U_{10})^{2.59}/u_*, \tag{24}$$

where U_{10} and U_t are respectively the 10-m wind speed and its value at the cessation of blowing snow, in m s^{-1}. This quantity is then extrapolated and fixed at z = 0.1 m from $z = z_s$ based on the analytical profiles for $F(r)$ of Equation (11). Note that in our initial steps, we assess the vertical distribution of r_m by taking this parameter to be 100 μm at z_s following Pomeroy et al. (1993), although recent observations in Antarctica suggest r_m at z_s may be closer to 75 μm (King et al., 1996).

2.2.4. *Other Modifications*

Following the detailed sensitivity tests of Déry et al. (1998) with the PIEKTUK-S model, several other modifications have been incorporated into PIEKTUK-B. For instance, they find little difference in assuming that the particles are at the ambient air temperature instead of the ice bulb temperature, and hence we take T_a as the particle temperature.

The threshold velocity for transport in PIEKTUK-B is estimated following the study of Li and Pomeroy (1997) such that:

$$U_t = U_{t_0} + 0.0033(T_a + 27.27)^2 \tag{25}$$

where T_a is in degrees Celsius and the minimum value of the threshold 10-m wind speed, U_{t_0}, is equal to 6.975 m s^{-1} and is reached at about T_a = −27°C. Equation (25) shows increasing resistance to transport at temperatures near freezing and at very low temperatures.

In addition, we do not assume at t = 0 that RH_i = 1.0 within the saltation layer. With our lower boundary set to the snow-covered surface, we can expect that this layer will contribute to the sublimation process and add water vapour to the ABL.

Given that the calculation of particle suspension is no longer constrained by the diffusion of small particles encountered in the spectral model, we are able to reduce significantly the vertical grid resolution while increasing the timestep. For the results presented in the following section, 24 vertical levels equidistant on a logarithmic scale and a timestep of 2 s are used in PIEKTUK-B. These changes, in addition to the elimination of the 64-particle size bins, augment the efficiency of PIEKTUK-B by a factor of about 100 over PIEKTUK-S.

3. Results

In the previous section, we described a simplified bulk algorithm for the depiction of the blowing snow process. We now perform a few tests to evaluate the model output of the bulk version of PIEKTUK in comparison to its spectral formulation. Both versions of the model have been modified following the discussion in Section 2.2.4, with the exception of the vertical and temporal resolutions in PIEKTUK-S, which maintain those used in Déry et al. (1998). For our control experiment, we take U_{10} = 15 m s^{-1}, and initially set $T_a = -10$°C, as background environmental conditions for a blowing snow period of 10 min. The initial humidity profile is obtained following Equation (23) by taking RH$_i$ at z = 100 m to be 0.7 and constant above that level. Other parameters required for the integration, such as u_* and z_0, are calculated following Déry et al. (1998). The vertical model boundaries are fixed at z_{lb} = 0 and z_{ub} = 1 km, respectively.

To evaluate the ability of PIEKTUK-B to reproduce the results of PIEKTUK-S, we first examine in Figure 1 the profiles of blowing snow mixing ratio in the control experiment as predicted by both models 10 minutes after the initiation of snow transport. The analytical profile of q_b that arises through a steady-state balance between diffusion and settling only is also shown in Figure 1. We see clearly that little accuracy is lost in using the bulk model to predict the variation of q_b with height.

For further tests, we examine the blowing snow sublimation and transport fluxes that affect the surface mass balance. The vertically integrated sublimation rate Q_s in units of mm h^{-1} snow water equivalent (swe) for a column of blowing snow is obtained from:

$$Q_s = -\rho' \int_{z_{lb}}^{z_{ub}} S_b \, \mathrm{d}z, \tag{26}$$

where ρ' is the conversion factor from m s^{-1} to mm h^{-1} swe. For convenience, we introduce a negative sign in Equation (26) to report the sublimation rate as a positive quantity. The transport rate of blowing snow Q_t (kg m^{-1} s^{-1}) is given by

$$Q_t = \rho \int_{z_{lb}}^{z_{ub}} U q_b \mathrm{d}z \,, \tag{27}$$

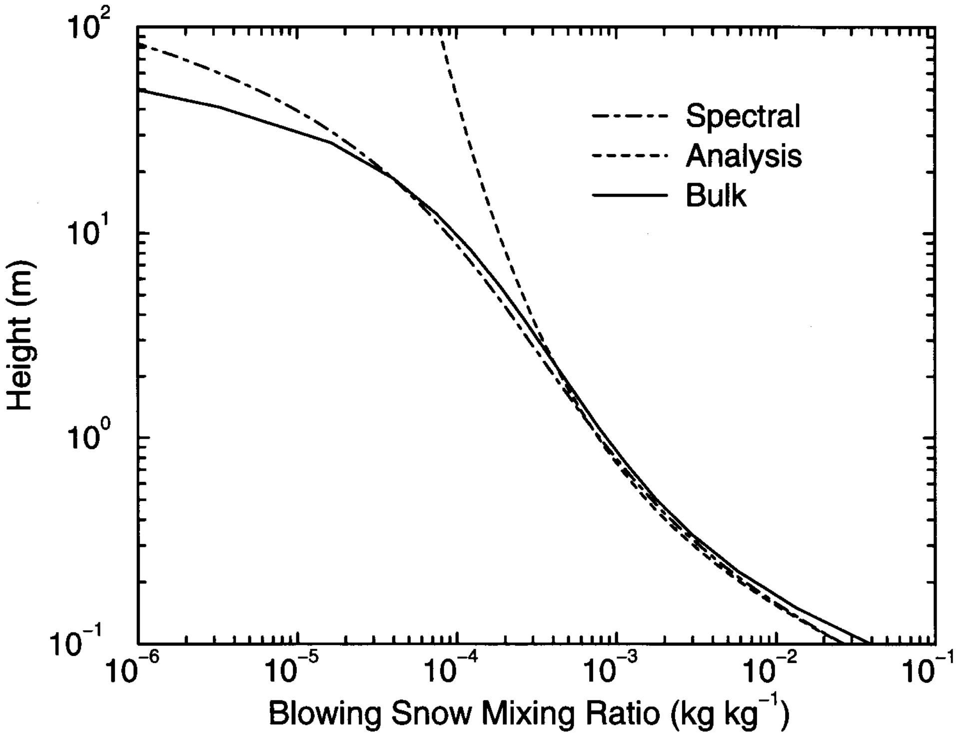

Figure 1. The profiles of blowing snow mixing ratio as predicted by the bulk and spectral versions of the PIEKTUK model 10 minutes after blowing snow initiation for the control experiment. The analytical result without sublimation ("Analysis") is also shown.

where U (m s^{-1}) is the wind speed. The evolution in time of Q_s and Q_t is shown in Figure 2 and, again, we see that for both cases good agreement between the bulk and spectral versions of PIEKTUK is found. Sublimation of blowing snow, as predicted by Déry et al. (1998), reaches a maximum within a few minutes of initiation, and then slowly decreases in time. This is related to the "self-limiting" property of blowing snow sublimation discussed by the authors. On the other hand, Q_t increases continually in time as higher humidities and diffusion act to augment the amount of suspended ice particles in a column of blowing snow.

Values of the blowing snow transport and sublimation rates predicted by both versions of PIEKTUK are shown in Table I for three values of the 10-m wind speed and two integration periods. The bulk model forecasts of the sublimation and transport rates match relatively well those of PIEKTUK-S, particularly at high wind speeds. For the control experiment, we find differences of approximately 5 and 3% between the integrated values of the sublimation and transport rates, respectively, as predicted by the bulk and spectral versions of PIEKTUK. For a one-hour period of blowing snow with U_{10} = 15 m s^{-1}, the cumulative sublimation rate leads to a depletion of 0.09 mm h^{-1} swe from the surface, equivalent to the

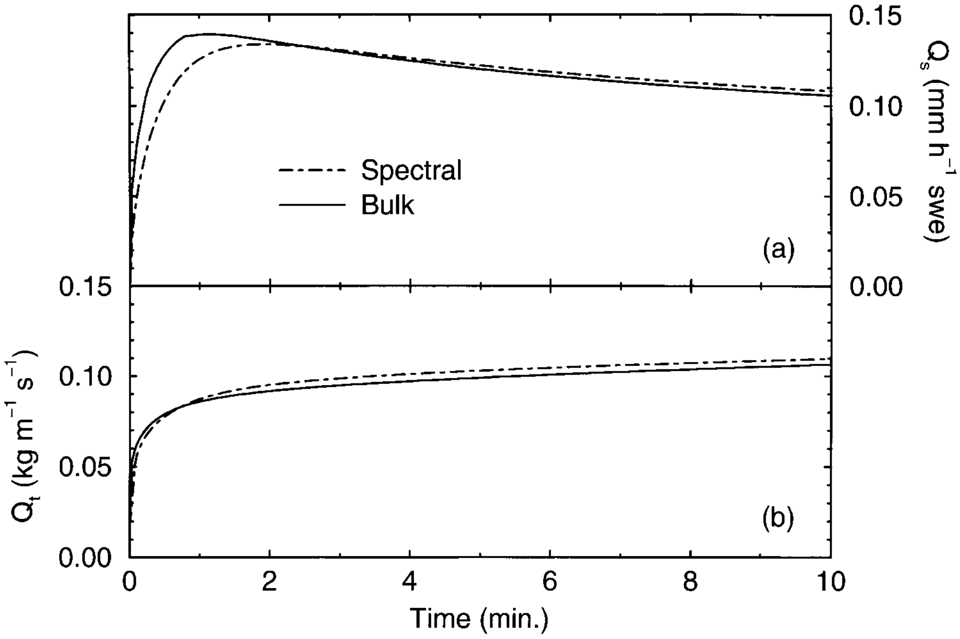

Figure 2. The temporal evolution of (a) the sublimation rate (Q_s) and (b) transport rate (Q_t) of blowing snow, predicted by the bulk and spectral versions of PIEKTUK for the control experiment.

removal of $\approx$ 2 mm swe per day. This is very similar to the sublimation rates reported by Schmidt (1982), King et al. (1996), and Bintanja (1998b).

In Figure 3, the thermodynamic profiles are shown, 10 minutes after the initiation of blowing snow, for the control experiment. These also show good correlation between the two versions of the model. Note how the sublimation of blowing snow leads to a weak temperature inversion in the ABL and a deviation from the logarithmic profile in humidity similar to the one proposed by Schmidt (1972) during blowing snow. Thus the temperature and humidity tendencies resulting from blowing snow sublimation predicted by PIEKTUK-B match those of PIEKTUK-S very well.

4. Sensitivity Tests

As discussed previously, the results presented in the previous section are highly dependent on the lower boundary conditions imposed on humidity. Maximum values of q_b are found in the saltation layer, which is usually a layer several centimetres thick just above the snow surface. Observations by Schmidt (1982) suggest that RH_i approaches 1.0 in this region but diffusion of moisture outside of the saltation layer may promote further sublimation near the surface. We therefore conducted

TABLE I

The sublimation rate (Q_s), in mm h^{-1} snow water equivalent (swe), and transport rate (Q_t) of blowing snow, for three wind speeds and two time integrations forecast by the spectral (S) and bulk (B) versions of PIEKTUK. Time-integrated values of sublimation (QT_s) and transport (QT_t) of blowing snow are also listed.

U_{10} (m s^{-1})	Time (min.)	Model Version	Q_s (mm h^{-1} swe)	QT_s (mm swe)	Q_t (kg m^{-1} s^{-1})	QT_t (kg m^{-1})
10	10	S	0.02747	0.004726	0.01592	9.296
		B	0.03185	0.005570	0.01716	9.939
	60	S	0.02223	0.02491	0.01692	58.83
		B	0.02535	0.02877	0.01892	64.50
15	10	S	0.1081	0.01973	0.1097	59.89
		B	0.1045	0.01999	0.1065	58.09
	60	S	0.06874	0.08882	0.1390	439.7
		B	0.06511	0.08644	0.1421	438.1
20	10	S	0.2041	0.03972	0.4903	255.1
		B	0.1897	0.03951	0.4619	243.3
	60	S	0.1052	0.1550	0.6863	2071
		B	0.09207	0.1434	0.6510	1962

several additional experiments to test the sensitivity of the results to this parameter by modifying the lower boundary condition imposed on humidity in PIEKTUK-B.

In our previous integrations, we assumed saturation with respect to ice at the surface. In the first test, we fix RH_i at 0.95 at the surface during the integration. As expected, Q_s is higher than for the control run, with an increase of about 14% in QT_s for a 10-min period of blowing snow (Figure 4). In two other experiments, we allow the lower boundary condition on humidity to vary in time (i.e., an “open” boundary condition) with limitation brought about by saturation with respect to ice. In one case, the initial humidity profile is as in the control run, while in the other one, we let $RH_i = 0.7$ initially throughout the column of air. We see that both show large increases in Q_s in comparison to the control experiment in the first few minutes that follow the initiation of blowing snow, but that both slowly tend towards the results of our control run in time as sublimation leads to increased moisture in the ABL. Differences in QT_s in these additional experiments are of the order of 20 to 30% higher than the control run.

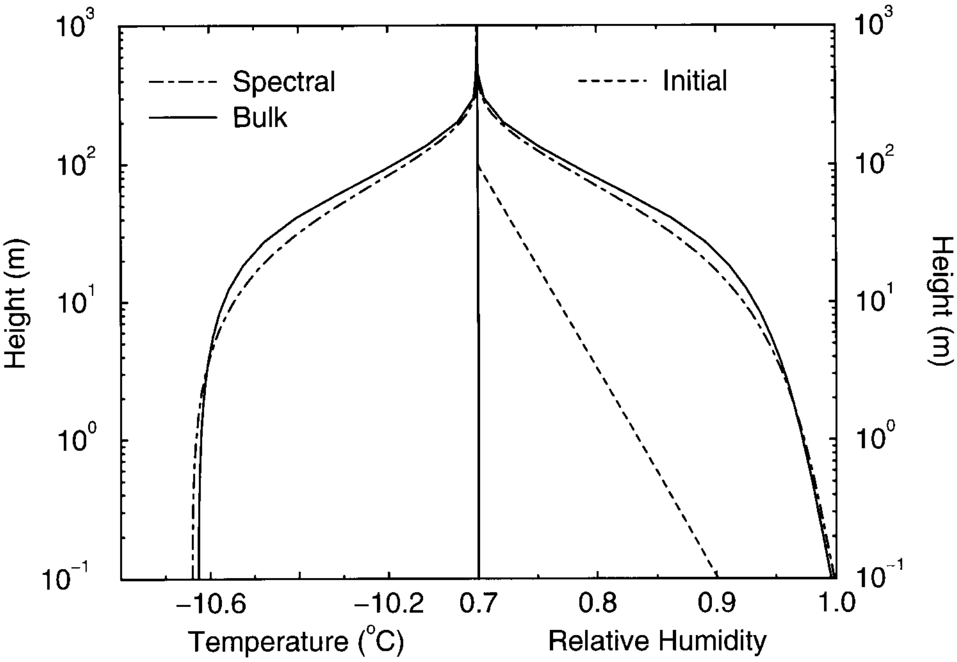

Figure 3. The profiles of ambient air temperature and relative humidity with respect to ice predicted by the bulk and spectral versions of the PIEKTUK model 10 minutes after blowing snow initiation for the control experiment.

5. Concluding Discussion

The water budget of a snow-covered surface may be affected by blowing snow through the redistribution of snow by wind and the concurrent sublimation of blowing snow. A number of studies have evaluated the contribution of these terms to the surface mass balance with notable variation on the significance of the sublimation component (e.g., Pomeroy et al., 1993, 1997; King et al., 1996; Bintanja, 1998b; Liston and Sturm, 1998). For instance, sublimation of blowing snow is evaluated to erode from a few millimetres swe at Halley, Antarctica over 6 months (King et al., 1996) to 37 mm swe during winter in a high-Arctic basin (Pomeroy et al., 1997). Considering the results presented in Table I for our control experiment in which $Q_s \approx 2$ mm d^{-1} swe, 17 days with continuous blowing snow would be required to erode the surface of the amount reported by Pomeroy et al. (1997). The climatology of cold-season processes compiled by Déry and Yau (1999) shows an annual average $\geq$ 30 blowing snow events for this region, thus potentially leading to the snow removal rates assessed by Pomeroy et al. (1997).

As discussed by Tabler and Schmidt (1972), Tabler (1975), and others, the sublimation rate will tend to increase with fetch or time if environmental conditions remain unchanged. However, in their idealized experiments, Déry et al. (1998), as well as the results in this study, show that the thermodynamic feedbacks of blowing

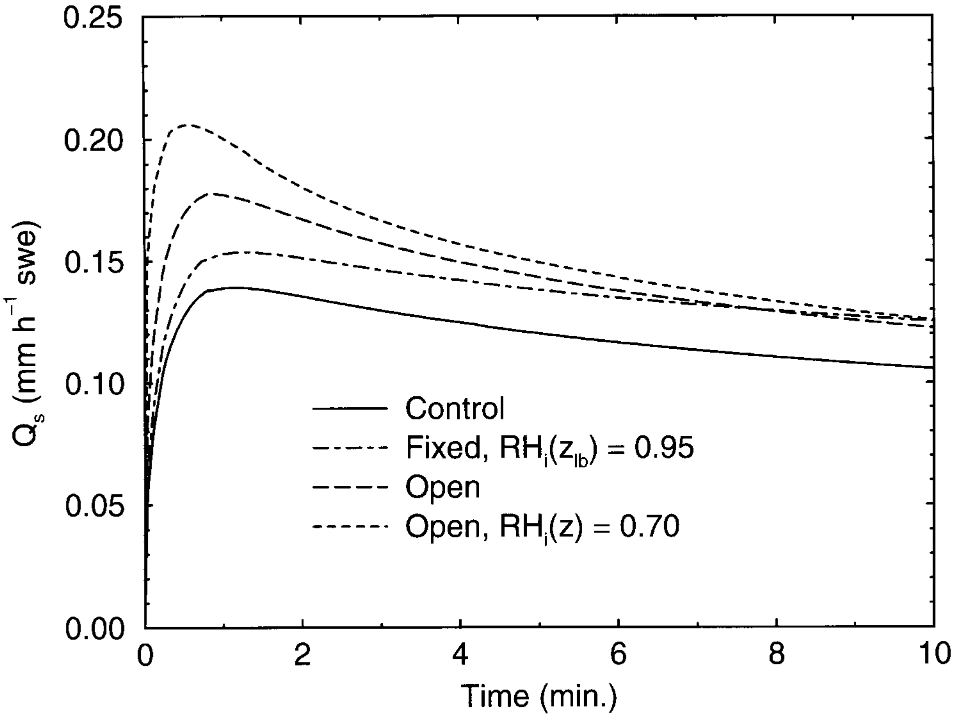

Figure 4. The temporal evolution of the sublimation rate Q_S of blowing snow for the control experiment and three sensitivity tests on the lower boundary conditions for humidity, (see text for a description of these tests.)

snow sublimation potentially lead to decreases in Q_S with time or fetch. As noted by King et al. (1996), Déry et al. (1998) and others, however, the modelling of blowing snow depends critically on the lower boundary conditions imposed on humidity. We conducted several tests that showed an increase of 14 to 30% in the accumulated sublimation rate from the control run when varying the initial and boundary conditions on humidity. Mixing of dry air from aloft (> 1 km) with air in the ABL may also promote sublimation of blowing snow. It remains clear that further investigation of the blowing snow process, including the measurement of humidity and temperature in near-surface air, is required to assess more accurately the contribution of blowing snow sublimation and transport to the mass balance of snow-covered surfaces.

To summarize, we have presented in this paper a brief outline of a simple and efficient algorithm of sublimating, blowing snow. This blowing snow model, named PIEKTUK, uses a bulk approach to predict the temporal evolution of the blowing snow mixing ratio, temperature and moisture profiles, as well as the interactive feedbacks between these, for a column of air in the atmospheric boundary layer. In comparison with a previous spectral version of PIEKTUK, the bulk model successfully forecasts the evolution of the sublimation rate of blowing snow and its mixing ratio profiles with significant savings in computer time (by a factor of about one hundred).

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