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Modelling snowdrift sublimation and its effect on the moisture budget of the atmospheric boundary layer

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ABSTRACT

A one-dimensional atmospheric surface layer model including turbulent diffusion and gravitational settling of suspended snow particles is used to simulate vertical profiles of snowdrift sublimation rates and the associated effects on the humidity and temperature profiles in the lowest 10 m. The simulations show that the thermodynamic feedback effects associated with snowdrift sublimation, i.e., strong increases in humidity and cooling, can significantly reduce the snowdrift sublimation rate, in particular in strong winds when large numbers of particles are being suspended. This negative feedback occurs because snowdrift sublimation depends on the undersaturation and temperature. Mechanisms that take away moisture from the surface layer, such as entrainment or horizontal advection of dry air, tend to weaken this feedback and enhance modelled snowdrift sublimation as the air generally remains undersaturated. Near the surface, however, the thermodynamic feedbacks dominate in strong winds, reducing the upward moisture flux from the surface. Then, snowdrift sublimation is the main contributor to the upward moisture flux at 10 m. Interestingly, in strong winds, the simulated total upward moisture flux in snowdrifting conditions is less than that in similar non-drifting conditions. Hence, the model results indicate that occurrence of snowdrift sublimation may, counterintuitively, eventually lead to a reduction of the surface-atmosphere moisture transport.

1. Introduction

The process of drifting snow is of glaciological and hydrological importance, as vast amounts of snow can be transported horizontally under strong winds. This constitutes a significant redistribution of the precipitated snow from one region to another with potentially large consequences for the local surface mass balance (Radok, 1970) and river runoff (Liston and Sturm, 1998) in snow-covered regions. Another important effect of snowdrift is that suspended particles experience a continuous sublimation and thereby induce a moisture flux from the particle to the ambient undersaturated atmosphere, as first noted by Dyunin (1967). The sublimation of snowdrifting particles influences the surface ablation as particles originating from the surface are being sublimed continuously, which represents an effective moisture flux from the surface to the atmosphere. As a result, the moisture budget of the boundary layer will be affected, most notably by increases in moisture content. Similar mechanisms occur in the sublimation of sea spray, which can drastically affect the sea–atmosphere moisture fluxes over the oceans (Rouault et al., 1991; Makin, 1997).

The first attempts to quantify snowdrift sublimation were done by Schmidt (1972) and Lee (1975); these were followed by calculations of snowdrift sublimation over seasonal snowcovers in North America by Schmidt (1982) and Pomeroy et al. (1993). Later it was realized that the feedbacks associated with the heat and moisture budgets of the boundary layer are essential in

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evaluating the snowdrift sublimation rate (Dover, 1993; Smith, 1995; Mann, 1998; Déry et al., 1998; Bintanja, 2000a; Xiao et al., 2000). In these studies the sublimation of a spectrum of suspended particles is calculated, as well as its effect on the heat and moisture budgets of the ambient air and on the particle spectrum (subliming particles decrease in size).

The main thermodynamic feedbacks associated with snowdrift sublimation are: (1) the increase in the moisture content of the ambient air caused by snowdrift sublimation which, in turn, decreases since it is proportional to the undersaturation of the air, and (2) the latent heat flux from particle to the surrounding air is compensated by a sensible heat flux in the opposite direction, which cools the air–particle mixture; this, in turn, reduces the saturation vapour pressure and hence sublimation rates. Both effects constitute negative feedbacks, demonstrating the inherent self-limiting nature of snowdrift sublimation. The general view is thus that once drifting snow enters into suspension and starts subliming, the moisture content of the air will increase and the air will cool, diminishing the undersaturation of the air and hence the sublimation rate. This goes on until, eventually, the air becomes saturated and sublimation vanishes. An example of the effectiveness of these mechanisms is given in Fig. 1, where relative humidity is plotted against wind speed for a weather station in Dronning Maud Land, Antarctica, during austral summer 1997–98. It clearly demonstrates that relative humidity increases once winds become stronger than the threshold for snowdrift initiation as a result of snowdrift sublimation. Observations made at the coastal station Halley (Antarctica) during the STABLE2 campaign in winter 1991 exhibit similar characteristics (Dover, 1993; Mann, 1998). These show a rapid formation of a saturated layer close to the surface as a result of snowdrift sublimation, which deepens as snowdrift continues. This behaviour is not a general feature, however, as there are indications that moisture contents in areas with seasonal snowcovers such as the North American prairies do not appear to increase much during snowdrift (Schmidt, 1982). To understand these differences in the behaviour of the moisture budget during snowdrift between various locations and the subsequent effects on the sublimation rate is the prime motivation behind this modelling study.

Pomeroy et al. (1993) use a Prairie Blowing Snow Model to calculate sublimation rates in Canadian Prairie environments from input of measured meteorological variables taken at one height, which were then extrapolated to the sur-

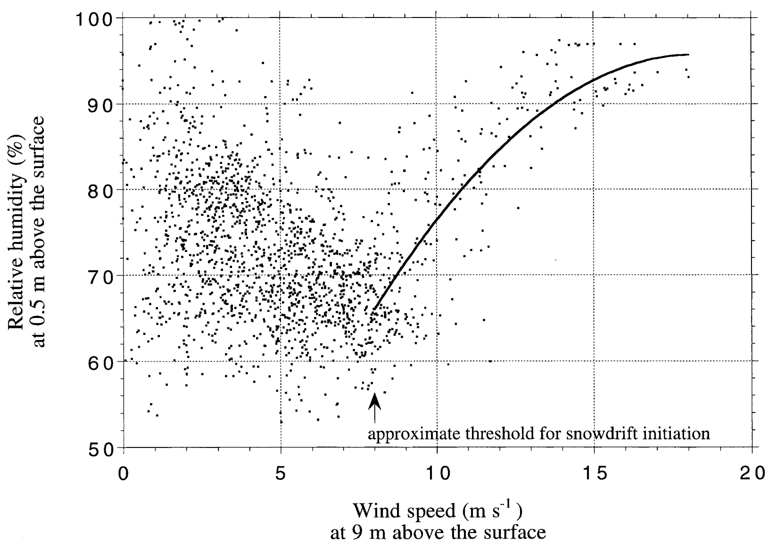

Fig. 1. Relative humidity (with respect to ice) as a function of wind speed as measured over a snow field in Dronning Maud Land, Antarctica, during austral summer 1997–98. Each dot represents a half-hourly mean value, while the line represents the best fit to the data in snowdrifting conditions.

face. Their analyses show that snowdrift sublimation can potentially have an important impact on the surface mass balance, with as much as 74% of the annual snowfall being eventually sublimed. However, it is unclear how the lower atmosphere remains undersaturated when such huge moisture fluxes are present. Smith (1995), Bintanja (1998b) and Gallée (1998) use a combination of numerical models and weather observations to evaluate the snowdrift sublimation rates over the Antarctic ice sheet. They all find that snowdrift sublimation rates are insignificant in the high interior where winds are weak and temperatures low. On the other hand, these studies agree that the relatively warm and windy coastal regions could experience strong snowdrift sublimation rates, which could significantly affect the surface mass balance. Even though the uncertainties in such estimates are still considerable (as will be demonstrated in this paper), it shows that snowdrift sublimation may be a non-negligible component of the surface mass balance of large ice sheets.

If all moisture originating from snowdrift sublimation would effectively be removed, that is, keeping undersaturation constant, snowdrift sublimation rates could become extremely high as demonstrated in Fig. 2. Under favourable conditions (warm, windy), equivalent latent heat fluxes could become as high as 1000–1500 W m^{-2} (equal to 30–50 mm water equivalent per day).

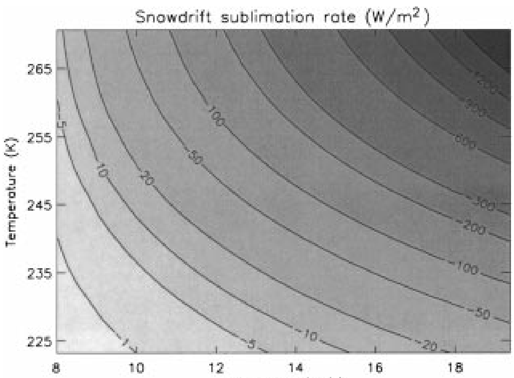

Fig. 2. Vertically-integrated snowdrift sublimation rates (in W m^{-2}) as a function of air temperature and wind speed at 10 m. Relative humidity was assumed constant at 70%. Note that a sublimation rate of 32.8 W m^{-2} is equivalent to a flux of 1 mm water equivalent per day, which equals 1.16×10^{-5} kg m^{-2} s^{-1}.

Obviously, these are enormous moisture fluxes with the potential to profoundly affect the surface mass balance of snow covered regions and the moisture budget of the overlying boundary layer; however, it is difficult to see how such gigantic fluxes could sustain without saturating the ambient air. To put these numbers in perspective, various observations indicate that surface sublimation rates over various snow surfaces rarely exceed 100 W m^{-2} under the most favourable conditions (Bintanja and Van den Broeke, 1995; King and Turner, 1997). The main reasons for the very high potential snowdrift sublimation rates are: (1) suspended snow particles are exposed on all sides, leaving a large surface area to mass ratio which is favourable for snowdrift sublimation (compared to surface sublimation), and (2) suspended particles are continuously ventilated as a constant upward drag is required to keep them at a certain average height (depending on their weight); the resulting mean vertical velocity difference equals the terminal fall velocity while they additionally experience turbulent fluctuations (ventilation significantly increases sublimation rates compared to still air conditions (Pruppacher and Klett, 1978)).

The foregoing indicates that mechanisms that are capable of effectively removing moisture from the atmospheric boundary layer may have profound effects on the magnitude of the snowdrift sublimation rate by keeping the air undersaturated. Examples of such mechanisms are horizontal or vertical advection of dry air and entrainment of dry air from aloft. If removal of the "additional" moisture during snowdrift events is somehow very efficient, the near-surface boundary layer will not become saturated and large snowdrift sublimation rates may occur for extended periods of time. If, on the other hand, such moisture-removal mechanisms are weak or absent, the negative thermodynamic feedbacks will quickly diminish snowdrift sublimation as well as surface sublimation (which depends on the vertical moisture gradient at the surface). Hence, paradoxically, the onset of snowdrift sublimation may reduce the total (i.e., snowdrift plus surface) sublimation and the associated moisture flux towards the atmosphere compared to a similar situation without snowdrift.

In this paper, we will include in a simple manner the mechanisms of downward entrainment of dry air into the boundary layer and of the horizontal advection of dry air to illustrate the potential effects of such processes on snowdrift sublimation rates. We will demonstrate the complex nature of the effects of snowdrift sublimation on the surface–atmosphere moisture fluxes. We will do this by applying a surface layer model, which includes all relevant thermodynamic effects of snowdrift sublimation (and vertical turbulent transports). All results presented in this paper are modelling results, which will be validated with data when these are applicable.

2. Model outline

The surface layer model (SNOWSTORM) used here to calculate vertical profiles of snowdrift-related quantities is described in detail by Bintanja (2000a). It will be briefly summarized here, for the convenience of the reader, in particular the aspects concerning snowdrift sublimation. The model centers around the atmospheric surface layer balance equations for momentum, heat, moisture and snowdrifting suspended particles:

$$\rho \frac{\partial u}{\partial t} = -\frac{\partial(\rho \overline{u'w'})}{\partial z}, \tag{1}$$

$$\rho c_p \frac{\partial \theta}{\partial t} = -\frac{\partial(\rho c_p \overline{w'\theta'})}{\partial z} - \lambda \bar{S}, \tag{2}$$

$$\frac{\partial q}{\partial t} = -\frac{\partial(\overline{w'q'})}{\partial z} + \frac{\bar{S}}{\rho_f} + E_H + A, \tag{3}$$

$$\rho \frac{\partial \eta_r}{\partial t} = -\frac{\partial}{\partial z}(\rho \overline{w'\eta_r'} - \rho V_r \eta_r) + \rho S_r, \tag{4}$$

where u, θ and q represent mean wind, potential temperature and specific humidity, respectively, $\overline{w'X'}$ is the vertical turbulent flux of quantity X, $\rho = \rho_f(1-C) + \rho_p C$ is the total density of the air–snow mixture with ρ_f and ρ_p the density of ice and air and $C = \eta/\rho_p$ the volume concentration of particles ($0 \leqslant C \leqslant 1$) with η the total particle mass concentration. c_p is the specific heat of air–snow mixture, λ the latent heat of sublimation, η_r, V_r and S_r the particle mass concentrations, terminal fall velocities and sublimation rates in particle size class with mid-radius r, $\bar{S}$ the sublimation rate summed over all particle size classes at each height above the surface (z). The first terms on the right-hand sides represent the vertical divergence of the various turbulent fluxes. The second term on the right-hand sides of (2) and (3) represents the cooling and the moisture input caused by snowdrift sublimation, respectively. Incidentally, these terms render the generally applied (in non-drifting conditions) Monin–Obukhov similarity theory to describe surface layer profiles invalid, as this theory requires that the turbulent fluxes are constant with height.

The terms E_H and A in eq. (3) represent entrainment of moisture at the top of the boundary layer and horizontal advection of moisture. In this study we are interested mainly in the effects of advection and entrainment on the moisture balance of boundary layer, and therefore similar terms in the other budget equations have been neglected for simplicity. Although we acknowledge that these may be relevant for a realistic description of the boundary layer response in general, their inclusion is not necessary for the qualitative analyses performed here. The second and third term on the right-hand side of (4) represent the gravitational settling of the particles and the effects of snowdrift sublimation, respectively. The particle equations (4) basically express a balance between gravitational settling, upward turbulent diffusion of suspended particles, with the sublimation term continuously reducing the size and mass of the suspended particles. In steady-state snowdrift, there is a net upward surface flux of suspended particles (erosion) balancing the mass loss by snowdrift sublimation. Hence, snowdrift sublimation induces a mass loss by the surface through net erosion.

The model uses 48 particle size classes with radii ranging from 5 to 475 μm and a bin size of 10 μm. The sublimation term in eq. (4) is expressed as:

$$S_r = -M_r \frac{\partial}{\partial r}\left(\frac{n_r}{4\pi r^2 \rho_p}\left(\frac{\partial m}{\partial t}\right)_r\right), \tag{5}$$

where M_r is the mass of particles and n_r is the number density of particles in size class r. $(\partial m/\partial r)_r$ is the rate of change of mass of a single subliming sphere of radius r (Thorpe and Mason, 1966):

$$\left(\frac{\partial m}{\partial t}\right)_r = \frac{2\pi r(q/q_s - 1)}{\dfrac{\lambda}{k_T T}\dfrac{1}{\mathrm{Nu}}\left(\dfrac{\lambda M_w}{RT}\right) + \dfrac{1}{k_v q \rho_f \mathrm{Sh}}}, \tag{6}$$

where q_s is the saturation specific humidity, k_T and k_v are the thermal diffusivity of air and the molecular diffusivity of water vapour in air, respectively, T is the absolute temperature, R the universal gas constant and M_w the molecular weight of water vapour. In this expression, Nu and Sh are the Nusselt and Sherwood numbers, respectively, which specify the enhancement of particle sublimation by ventilation over its value in still air. The ventilation rate of each particle depends on its mean terminal fall velocity as well as on the turbulent intensity of the relative velocity fluctuations (Lee, 1975; Dover, 1993). The terminal fall velocity depends on the weight of the particle and hence its radius (Mann, 1998). Heavy particles with high inertia will be unable to follow the high frequency turbulent fluctuations, which results in relatively high turbulent ventilation rates. Hence, turbulent ventilation depends on particle radius, but also on the turbulent intensity. Total particle sublimation rates depend on wind speed, particle radius, the undersaturation of the ambient air and the air temperature. Note that the snowdrifting particle spectrum changes continuously as particles steadily become smaller through sublimation (it is assumed that particles are spherical), reducing the sublimation rate.

Sublimation of snowdrift leads to cooling of the air through the final term in eq. (2). The cooling is largest where sublimation is most intense. This is generally near the surface, so that snowdrift sublimation leads to intensifying thermal stable stratification. This reduces the vertical turbulent fluxes by diminishing the turbulent fluctuations. Hence, the vertical moisture flux is reduced, intensifying the negative feedback between snowdrift sublimation and increases in moisture content of the ambient air.

The turbulent fluxes in eqs. (1)–(4) are closed using first-order closure and the modified E–ε closure scheme (Bintanja, 2000a), solving the prognostic equations for turbulent kinetic energy (E) and dissipation (ε). Bintanja (2000b) demonstrates that snowdrift suspension induces a stable density gradient, thereby decreasing turbulent kinetic energy and turbulent length scales. In the most extreme cases (strong winds, high amounts of particles), decreases in E can amount to 20% over its value without suspended particles, resulting in decreases in the eddy exchange coefficient of up to 40%. The most extreme effects of a two-phase flow on turbulence occur near the surface where the suspended load is most abundant. However, Bintanja's (1998a; 2000b) theoretical calculations demonstrate that also at higher levels in the atmospheric surface layer the effects of suspended snowdrift on the turbulent intensity of the flow can be significant, with high particle Richardson numbers. Xiao et al. (2000) show that the value of the eddy exchange coefficient is important for the evaluation of the sublimation rate, since a high turbulent moisture flux will transport the excess moisture more effectively upward enabling higher sublimation rates.

The saltation layer particle concentration (η_s) is used as lower boundary for eq. (4). For this we use the empirically derived relationship of Pomeroy and Gray (1990):

$$\eta_s = \frac{0.68\rho_f}{c_t u_* g h_s}(u_*^2 - u_{*t}^2) \qquad \text{for } u_* \geqslant u_{*t} \tag{7}$$

where u_* and u_{*t} are the friction velocity and the threshold friction velocity, respectively, c_t a parameter related to saltation efficiency, g the acceleration of gravity and h_s the height of the saltation layer taken proportional to $u_*^{1.27}$ (Pomeroy and Male, 1992). In the saltation layer, particles are distributed in the 48 size classes according to the well-known two-parameter gamma-distribution function (Budd, 1966) with $\alpha = 2$ and a mean radius of 75 μm. The saltation layer is regarded to be saturated at all times, so no snowdrift sublimation occurs there (Mann, 1998).

Fig. 3 shows a schematic picture of the simplified moisture budget in snowdrifting conditions. The moisture sources of snowdrift and surface sublimation add moisture to the boundary layer whereas turbulent diffusion, entrainment and horizontal advection act as moisture sinks. Entrainment of air from the free atmosphere into

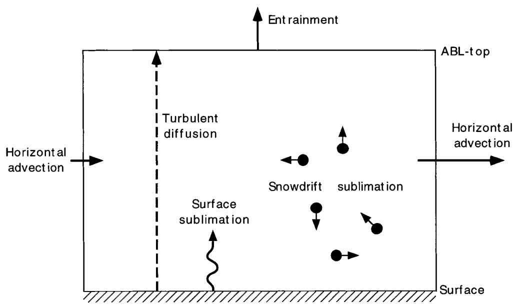

Fig. 3. Schematic representation of the moisture budget in the atmospheric boundary layer during snowdrifting.

the boundary layer is a widely occurring feature over ice sheets and glaciers (Fortuin and Oerlemans, 1993; Van den Broeke et al., 1994). In eq. (3), the entrainment moisture flux (E_H) applied to the top model layer is taken as:

$$E_H = w_e \left(\frac{\Delta q}{\Delta z}\right)_e, \tag{8}$$

where w_e is the entrainment velocity and $(\Delta q/\Delta z)_e$ is the (negative) vertical moisture gradient over the entrainment layer. We will assume that both parameters are independent of the prognostic boundary layer quantities, so that E_H represents a constant sink of moisture applied to the uppermost model layer ($z = H$) in all simulations. In other words, we assume that the free atmosphere overlying the boundary layer effectively removes the excess moisture from the boundary layer, enabling a constant humidity gradient over the entrainment layer. In the real atmosphere, the entrainment fluxes are strongly related to the wind shear and the temperature inversion over the entrainment layer.

Horizontal advection of moisture plays a rôle when for instance cold dry air flows down a slope while adiabatically heating such as in katabatic flows. The horizontal advection of moisture (A) in eq. (3) is represented as:

$$A = u \frac{\partial q}{\partial x}, \tag{9}$$

where $\partial q/\partial x$ represents the horizontal gradient in specific humidity (assumed to be independent of height and time) along a stream line of the flow. We will assume that this term is negative, as is found in descending, katabatic flows over Antarctica's steeply inclined surfaces (Van den Broeke, 1997), indicating advection of dry air. The vertically varying (through vertical variations in wind speed) advection term A is applied to each model layer. The magnitude of the horizontal advection term is constant in time during the model runs.

The model domain ranges from the focus height (2–10 cm), taken equal to 80% of the saltation layer height were the wind speed during snowdrift is assumed to remain constant, to a height of 100 m. We will analyze and present profiles only for the lowest 10 m, as this is the range of interest because the vast majority of the suspended particles generally remain below 10 m and it is the approximate range for which the surface layer equations (1)–(4) are valid. Their extension to higher levels is merely applied to reduce the unwanted effects of uncertain upper boundary conditions on the profiles below 10 m. For example, we assume a temporarily constant turbulent moisture flux at the upper model boundary equal to the initial surface latent heat flux (in fact, the initial vertical turbulent moisture flux is assumed constant throughout the model domain, representing the latent heat flux originating from the surface). Additional model runs performed with the upper boundary at 1000 m did not change the results in a qualitative sense. At the lower boundary the latent heat flux is free to change as it depends only on the simulated local vertical gradients in q and u. The surface temperature is calculated assuming a zero potential temperature gradient at the focus. As virtually nothing is known about the temperature and moisture fields in the saltation layer, this procedure to evaluate the surface latent heat flux can be regarded as a plausible approximation of this complex process. Each model run starts with logarithmic wind, potential temperature and specific humidity profiles (and the associated profiles of E and ε) and no suspended particles. Each simulation will be run for 83 min, during which most interesting features with regard to snowdrift sublimation take place. For comparison, Mann et al. (2000) mention that typical snowdrift events at Halley, Antarctica, last for about 36 hours.

3. Results of model simulations

Snowdrift sublimation rates depend crucially on wind speed and temperature (Fig. 2). In the present analyses we will assume constant initial temperatures (-10°C) throughout all model runs while we will vary the wind speed. In lower temperatures, the sublimation rates and the associated effects on the moisture budget are weaker as evident from Fig. 2. The initial vertical moisture gradient ($\partial q/\partial z$), which determines the initial surface latent heat flux, is assumed the same in each model run. Table 1 shows values of various model variables. Model results will be qualitatively compared to observations where possible. The observations of snowdrift and concurrent wind, temperature and humidity profiles at Halley,

Table 1. *Values of the various model variables*

Parameter	Value
threshold friction velocity u_{*t}	$0.25\ \mathrm{m\ s^{-1}}$
air density ρ_f	$1.34\ \mathrm{kg\ m^{-3}}$
snow particle density ρ_p	$900\ \mathrm{kg\ m^{-3}}$
c_t	2.8
initial potential temperature	263.15 K
initial humidity gradient q_* ($= \kappa\, \partial q/\partial \ln z$)	-1×10^{-5} kg/kg

Antarctica, during winter 1991 in the STABLE2 experiment is the best and most complete dataset available to date (Dover, 1993; Mann, 1998; Mann et al., 2000). Additionally, Schmidt (1982) describes measurements of simultaneous snowdrift and meteorological profiles on a winter snow cover in Wyoming, USA.

3.1. *Excluding entrainment and advection*

In these model simulations, the terms E_H and A in eq. (3) have been assumed zero to investigate the evolution of snowdrift sublimation in a surface layer where only turbulent diffusion can transport away the moisture generated by snowdrift sublimation. As this is not a very efficient transport mechanism, the thermodynamic feedback effects of the surface layer can dominate the moisture budget response. The majority of the thermodynamic snowdrift models have been applied in this manner (Mann, 1998; Déry et al., 1998; Bintanja, 2000a; Xiao et al., 2000). Note that the additional moisture (i.e., produced by snowdrift sublimation) must remain in the model domain since the upper-boundary moisture flux is fixed at its initial value; the additional moisture will therefore accumulate in the model domain. The model is initialized with logarithmic wind profiles using 10 m winds of 8.2, 9.8, 12.5, and 14.8 m s^{-1}, which all induce surface friction velocities higher than the threshold (i.e., 0.3, 0.4, 0.6 and 0.8 m s^{-1}) and cause snowdrift suspension and sublimation. These can be viewed upon as being forced externally by for instance varying synoptic pressure gradients.

Fig. 4 shows the modelled temporal evolution of relative humidity in the lowest 10 m for two cases. Obviously, relative humidity increases in both cases, the strongest increases occurring in the strongest winds. This is due entirely to the

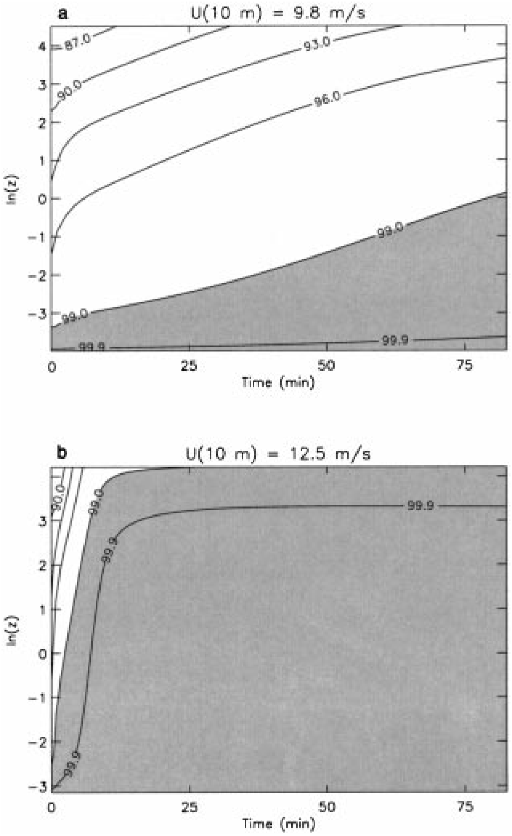

Fig. 4. Modelled temporal evolution of the relative humidity profile for 10 m wind speeds of (a) 9.8 m s^{-1} and (b) 12.5 m s^{-1}. The following contours are drawn: 87, 90, 93, 96, 99, 99.9.

injection of moisture by snowdrift sublimation. Under strong wind forcing the layers close to the surface approach saturation very quickly (within 20 min), effectively diminishing snowdrift and surface sublimation. The STABLE2 data confirm this, with rapid increases in humidity in the lowest layers immediately following onset of snowdrift, saturating an 11 m thick layer within about one hour (Mann et al., 2000). The observations show that the relative humidity at 11 m rose from 87% to 97% during this period, indicating that the simulated increases in humidity are realistic. Near-saturated conditions are reached fastest near the surface because (1) they are initially already close to saturation and (2) most particles susceptible to sublimation are close to the surface. Fig. 5 depicts the temporal variation of various quantities for

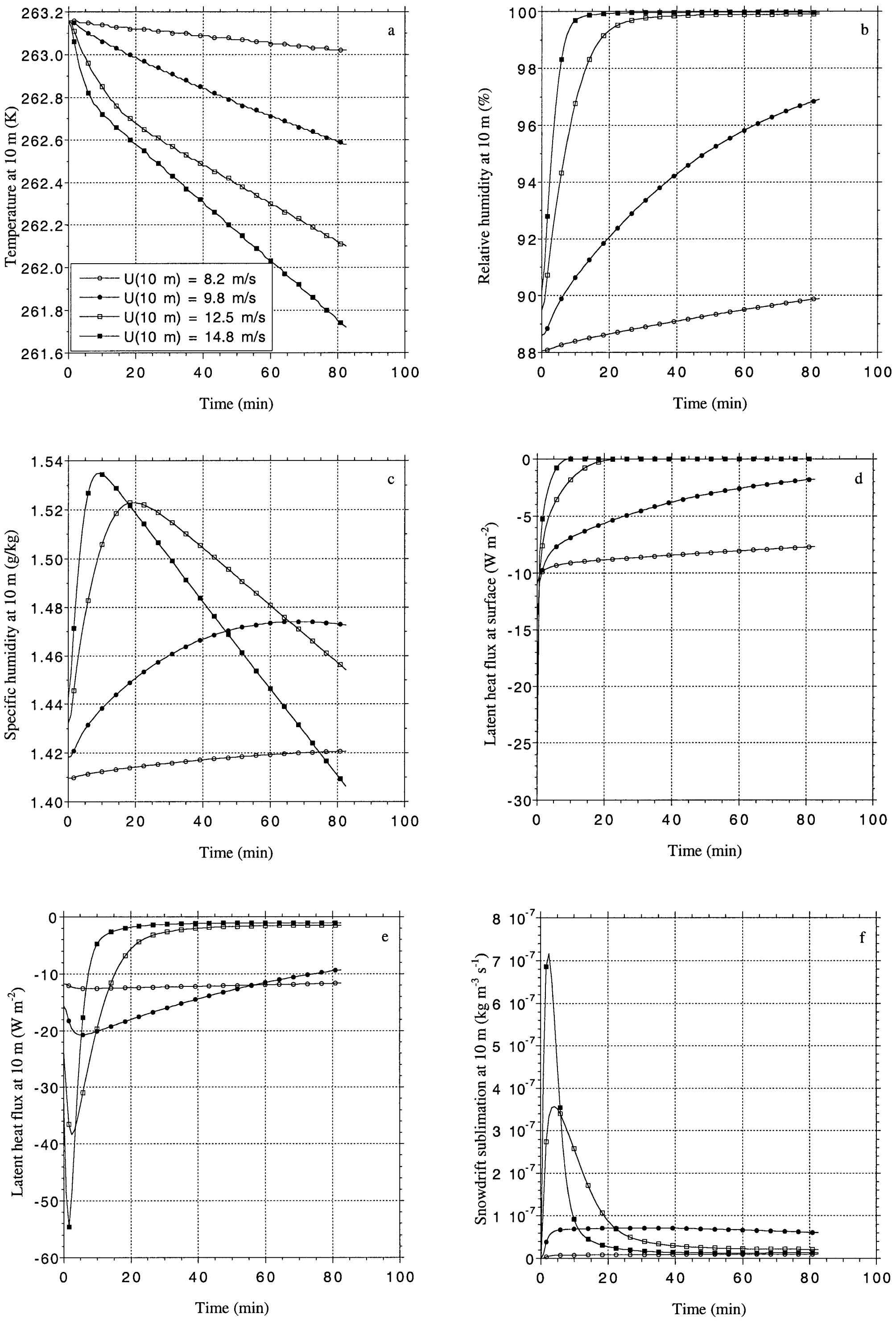

Fig. 5. Modelled temporal evolution of (a) 10 m temperature, (b) 10 m relative humidity, (c) 10 m specific humidity, (d) surface latent heat flux, (e) 10 m latent heat flux, and (f) 10 m snowdrift sublimation rate for 4 values of wind forcing. Negative latent heat fluxes indicate upward moisture transports.

all four cases. Increases in relative humidity are largest under the strongest winds, and similarly the associated temperature decreases are most pronounced in the strongest winds. The latter contributes significantly to the rapid saturation of the atmosphere as lower temperatures decrease the saturation water vapour pressure. This is why the specific humidity at 10 m decreases after an initial increase in the strongest forcings. Once saturation is reached, specific humidity depends only on temperature and must therefore decrease. For the two weakest forcings, relative humidity increases throughout the simulation and temperature decreases are smaller, resulting in increasing specific humidity values at 10 m.

The upward latent heat flux at the surface generally declines after snowdrift initiation, as near-surface humidities increase (Fig. 5d). For the strongest wind forcing, the simulations show that the surface latent heat flux reduces to zero within 20 min after onset. This fast response is in qualitative agreement with the STABLE2 observations (Mann et al., 2000), which exhibit similarly fast saturation of the near-surface layers in snowdrift. At 10 m, simulated latent heat fluxes exhibit peak values just a few min after onset of snowdrift, which compares well with simulations of Déry et al. (1998). This can be attributed to the contribution of the extra moisture generated by snowdrift sublimation below the 10 m level, increasing temporary the vertical moisture gradient. The extra moisture is then transported upward along with the moisture originating from the surface. After that, the latent heat fluxes decline due to the increasing moisture content and the decreasing vertical moisture gradient, similar to the near-surface decrease. Interestingly, the upward latent heat flux in the surface layer including snowdrift sublimation can become smaller than in similar non-drifting situations when only surface sublimation occurs. In fact, the simulations show that the upward moisture flux can even be reduced to zero under the strongest forcings! Paradoxically, this means that the addition of an apparent "extra" moisture source may effectively diminish the moisture flux from the surface to the atmosphere. This is a counterintuitive result, which can largely be attributed to the fact that snowdrift sublimation constitutes a huge potential moisture source (Fig. 2) that cannot be removed efficiently to higher levels by turbulent diffusion alone. Consequently, it accumulates in the surface layer where it leads to rapid saturation as a result of the temporary dominance of snowdrift sublimation in the moisture budget. Apparently, upward turbulent diffusion of moisture is not efficient enough to remove the additional moisture away from the lowest atmospheric layers. Moreover, the vertical turbulent transports close to the surface are hampered by the negative buoyancy induced by the stable particle and thermal stratification induced by snowdrift (Déry and Taylor 1996; Bintanja, 2000a, 2000b).

Initially, modelled sublimation rates are largest close to the surface, since the number of particles is largest there (Fig. 6). When these layers become saturated the maximum in snowdrift sublimation progresses upward. The sublimation rates at 10 m show peak values just after the "onset" of snowdrift. Then, the atmosphere is still relatively dry and the suspended particles undergo rapid sublimation. However, sublimation rates decline after this initial peak due to increasing humidity, especially for the strongest forcings. Sublimation rates at 10 m (and throughout the entire surface layer) diminish significantly in the first 10–20 min after snowdrift initiation for the highest wind speeds, remaining small thereafter. Fig. 7 depicts the vertical profile of simulated sublimation rate averaged over the integration period. The vertical gradient in the snowdrift sublimation rate decreases with increasing forcing. This is because under weak wind forcing there are relatively few particles at higher levels and the air remains undersaturated at all levels, resulting in a large vertical sublimation gradient. Under strong forcing, the air rapidly

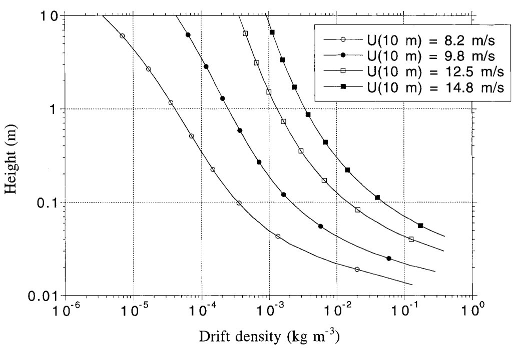

Fig. 6. Simulated vertical profiles of snowdrift density for 4 values of wind forcing.

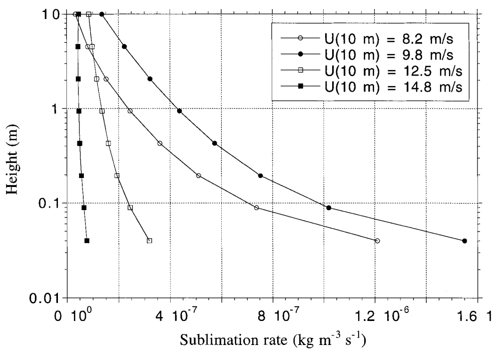

Fig. 7. Simulated vertical profile of integrated (over the 83 min model run) snowdrift sublimation rates for 4 values of wind forcing.

reaches saturation resulting in a weak vertical gradient.

A more interesting feature of Fig. 7 is that the absolute value of the vertically and temporally averaged sublimation rate ($\hat{S}$) peaks at intermediate values of the ambient wind speed. This is caused by two opposing effects: (1) the increase in $\hat{S}$ with wind speed is caused by the fact that in high winds more particles enter into suspension and that they are stronger ventilated, (2) under high sublimation rates the ambient air gets saturated quickly, reducing the temporally averaged sublimation rate. At low winds, the first effect dominates and the sublimation rate increases with wind speed. At high winds, the second effect dominates so that the temporarily integrated sublimation rate decreases with increasing wind speed irrespective of the higher number of suspended particles. This is exactly what is found in the STABLE2 data, which very clearly demonstrates that above a certain wind speed snowdrift sublimation starts to decline due to the increased humidity levels, irrespective of the further increase in the amount of suspended particles (Mann et al., 2000). This behaviour is expected to be influenced by other processes that affect the moisture budget, as will be investigated in the next two sections.

3.2. *The effects of entrainment*

2 additional simulations have been performed for a 10 m wind speed of $9.8\ \mathrm{m\ s^{-1}}$. In these simulations, dry air is entrained downward into the model domain (at $z = H$) with values of the term E_H in eq. (3) of $1.13 \times 10^{-6}\ \mathrm{s^{-1}}$ and $2.25 \times 10^{-6}\ \mathrm{s^{-1}}$, respectively. The magnitude of the entrainment flux is chosen so that its effect on the moisture budget is significant, as some observations of humidity during snowdrift over seasonal snowcovers indirectly imply. Fig. 8a, b show the simulated temporal variations in relative humidity in the lowest 10 m for the two cases with entrainment. While for $E_H = 0\ \mathrm{s^{-1}}$ (Fig. 4a) the relative humidity steadily increases at all levels, the inclusion of the above entrainment fluxes compensate for this. In each of the two cases with entrainment the additional upward moisture flux effectively removes the snowdrift sublimation-generated moisture. Hence, modelled relative humidities generally decline after an initial increase due to snowdrift sublimation. Entrainment takes away moisture at the upper part of the model, thereby increasing the vertical gradient of specific humidity and hence the upward turbulent transport of moisture. Apparently, the inclusion of significant entrainment fluxes is able to efficiently remove most snowdrift-generated moisture. Nonetheless, the simulations show that the layers close to the surface initially experience rapid moisture inputs from snowdrift sublimation before entrainment is able to remove this excess moisture. This is because entrainment takes place at the top of the model, and it takes time before it affects the layers close to the surface. As a result, the entrainment moisture flux cannot prevent the initial strong decrease in surface moisture flux.

This is apparent from Fig. 8c, d, which depict the latent heat flux. Note that the initial latent heat flux is $-15.5\ \mathrm{W\ m^{-2}}$. Apart from the initial decline close to the surface just after snowdrift initiation, the upward turbulent fluxes of moisture significantly increase throughout the entire domain when entrainment of dry air is included. It should be noted that the applied entrainment fluxes are probably unrealistically large (even though these may occur in some regions with descending, Föhn-type winds), which explains the vigorous drying. Fig. 9 shows the simulated temporal evolution of vertically-integrated snowdrift sublimation rate for the three cases. Inclusion of such high entrainment rates facilitates a continuous snowdrift sublimation rate, the magnitude of which depends crucially on the value of the entrainment flux. It can be concluded that an appreciable upward moisture transport from the

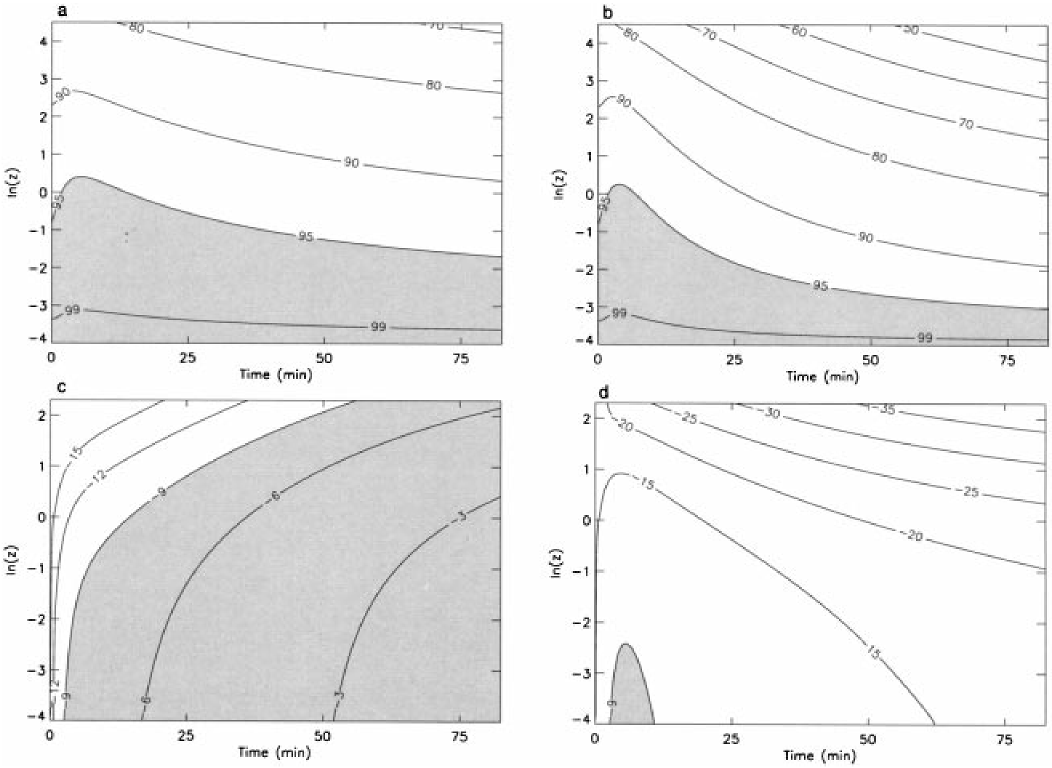

Fig. 8. Modelled temporal evolution of the relative humidity profile for the two cases including entrainment. (a) $E_H = 1.13 \times 10^{-6}\ s^{-1}$ and (b) $E_H = 2.25 \times 10^{-6}\ s^{-1}$. Temporal evolution of the latent heat flux profile for the cases (c) $E_H = 0\ s^{-1}$ and (d) $E_H = 1.13 \times 10^{-6}\ s^{-1}$.

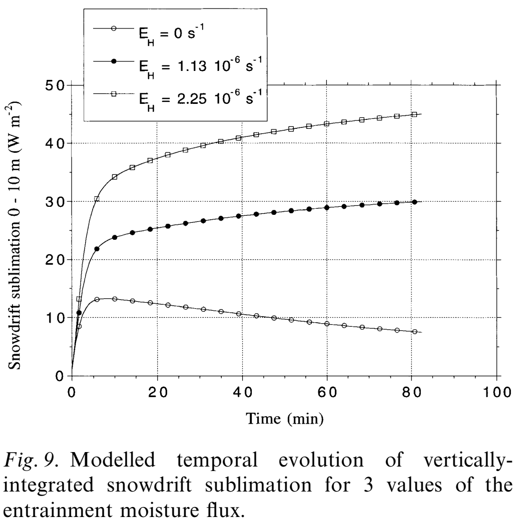

Fig. 9. Modelled temporal evolution of vertically-integrated snowdrift sublimation for 3 values of the entrainment moisture flux.

boundary layer to the free atmosphere aloft by entrainment is necessary to efficiently remove the 'extra' moisture and hence to facilitate snowdrift sublimation, thereby significantly increasing the surface–atmosphere moisture fluxes.

3.3. *The effects of horizontal advection*

3 additional simulations have been carried out to investigate the effect of horizontal advection of dry air, again for a wind forcing of 9.8 m s^{-1}. This mechanism is potentially important in descending, katabatic regions, in which cold and dry air flows down the slope (Van den Broeke, 1997). Vertically varying values of A (eq. (3)) of 4.4×10^{-8}, 8.8×10^{-8} and $1.76 \times 10^{-7}\ s^{-1}$ (vertical mean values) have been applied and compared to the model simulation without horizontal advection. Again, these values were chosen so that the effects on the moisture budget were significant. The mod-

elled temporal variation of the important variables is shown in Fig. 10. Obviously, the additional "source" of dry air reduces the relative humidity and enhances the snowdrift sublimation rate and the vertical turbulent latent heat flux. The main difference with the inclusion of entrainment is that the advection of dry air occurs throughout the entire vertical column, and can thus instantly affect the near-surface moisture budget. The simulations show that in the case with the largest advection the latent heat flux at the surface becomes larger than it initially was, after an initial decline due to snowdrift sublimation. For the weaker advection cases, the surface latent heat flux remains smaller than its initial value, indicating that in these cases horizontal advection of dry air is not efficient enough to remove all additional moisture injected by snowdrift sublimation near the surface. At 10 m, the upward turbulent flux of moisture increases in each of the three cases with advection, which is caused by the enhanced vertical moisture gradient invoked by (1) the increase in snowdrift sublimation-generated moisture content at lower levels and (2) the general decline in moisture content imposed by the advection term. Modelled snowdrift sublimation rates at 10 m exhibit their usual temporal variation pattern: a sharp increase just after onset of snowdrift, followed by a slight decline rapidly thereafter caused by the increase in humidity (remember that the air near the surface is close to saturation, so that an injection of moisture caused a rapid increase in relative humidity); finally, sublimation rates tend to increase again (for the cases with strong advection of dry air) because of the removal of moist air. In intermediate winds, the increase in moisture content due to snowdrift sublimation is larger than the decrease caused by advection, leading to a decrease in snowdrift sublimation rates with time.

4. Discussion

The results of the various simulations presented in the previous sections have shown that snowdrift sublimation and its associated effects on the moisture budget of the boundary layer may depend strongly on the ability of other boundary layer processes to remove the inherent moisture input. Hence, the relative effects of moisture removal by mixing or advection compared to moisture input by sublimation determines whether moisture can accumulate in the boundary layer. If the removal mechanisms are weak or absent, the thermodynamic effects associated with snowdrift sublimation (increased humidity, reduced temperature) quickly diminish snowdrift sublimation and surface sublimation, the effects being strongest for the highest winds. As a consequence, the vertical turbulent flux of moisture from the surface to the atmosphere may actually be reduced in the presence of snowdrift (compared to a similar situation without snowdrift).

Of particular interest is the effect of the snowdrift sublimation on the moisture budget and the *surface* sublimation rate. Because snowdrift sublimation will tend to saturate the layers closest to the surface first as these are initially already close to saturation, the surface sublimation (which depends on the vertical moisture gradient at the surface) will be significantly reduced. Results show that the importance of surface sublimation declines sharply with increasing wind speed owing mainly to the reduction of the specific humidity gradient near the surface in strong winds. Calculations using the STABLE2 data have revealed that 70–100% of the total sublimation is due to snowdrift sublimation during winter (Smith, 1995). This was established by comparing the measured moisture flux at 11 m with the calculated (from data) snowdrift sublimation rates. Although the STABLE2 measurements were taken during winter when temperatures were lower than those imposed here, it nevertheless demonstrates that snowdrift sublimation may be the only moisture source left during snowdrifting events. The simulations show that increasing levels of moisture removal by either entrainment or horizontal advection apparently tends to increase the importance of surface sublimation, a somewhat surprising result as one would expect the more efficient moisture producing mechanism of snowdrift sublimation to increase its dominance in undersaturated environments. In the cases with strongest moisture removal, surface sublimation accounts for 40–50% of the total moisture transport at the end of the integration (compared to 20% in the case without removal).

In order to infer the dependence of the moisture fluxes on the wind forcing, Fig. 11a, b depict the modelled upward moisture fluxes at the surface

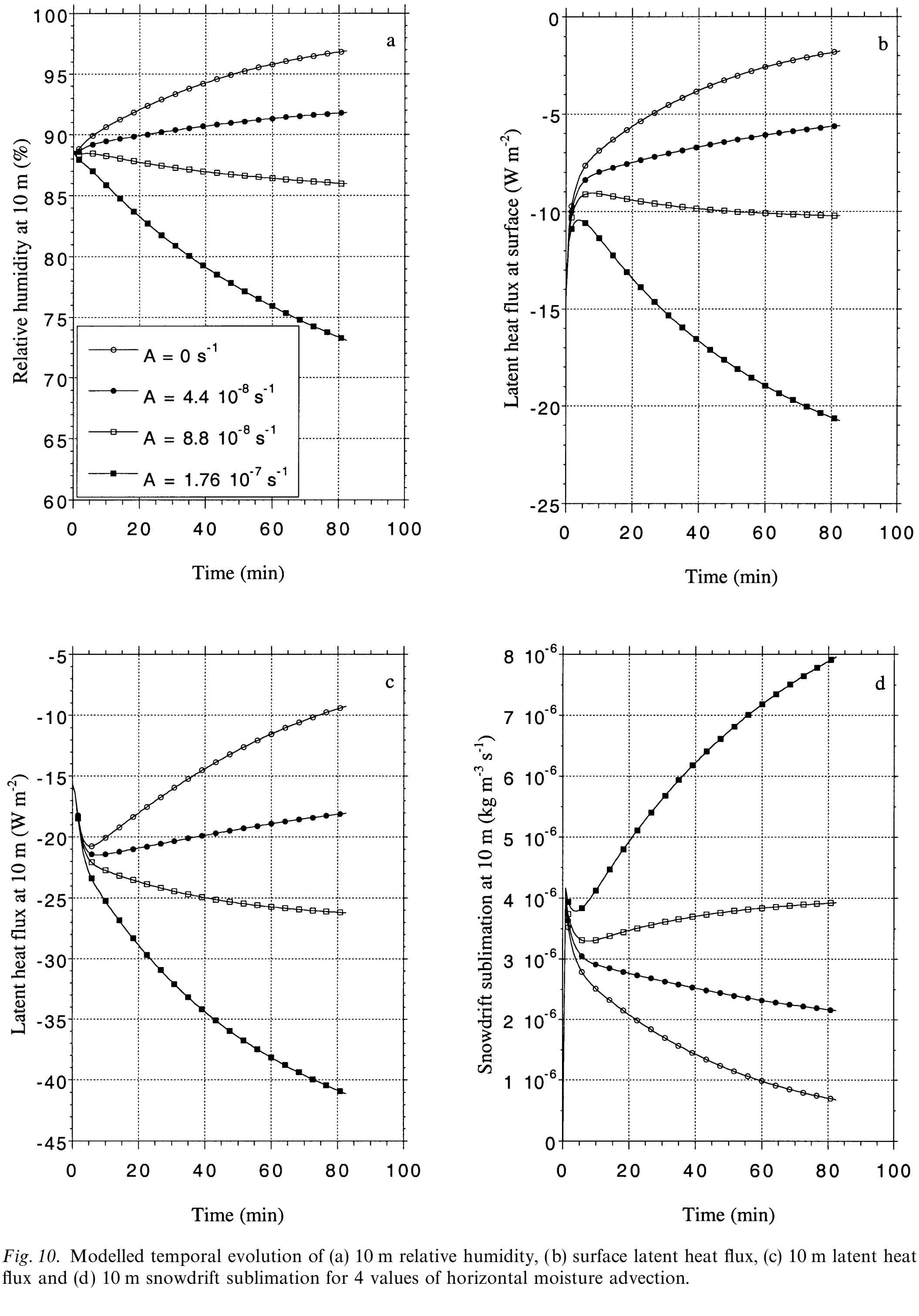

Fig. 10. Modelled temporal evolution of (a) 10 m relative humidity, (b) surface latent heat flux, (c) 10 m latent heat flux and (d) 10 m snowdrift sublimation for 4 values of horizontal moisture advection.

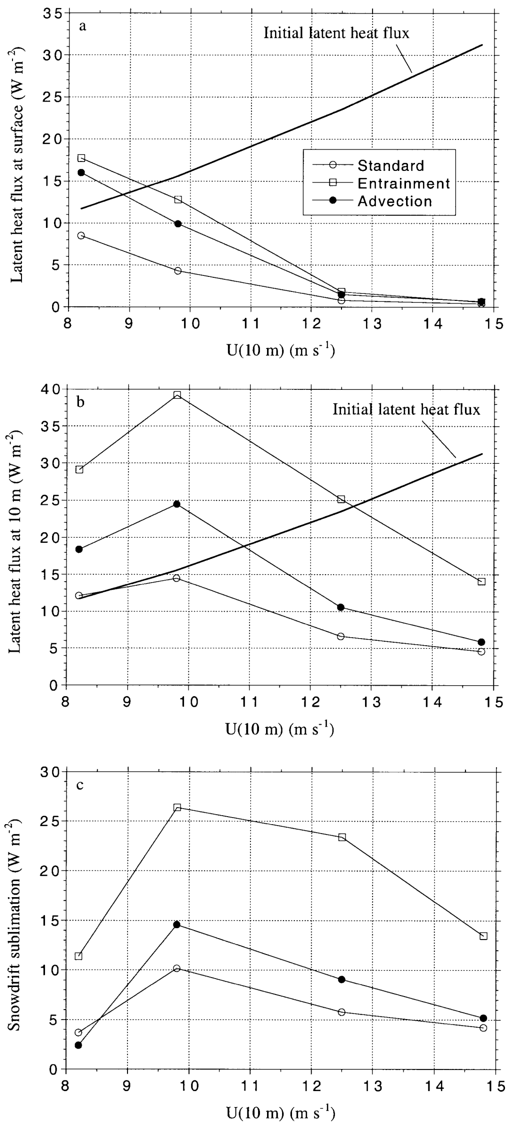

Fig. 11. Simulated variation of (a) temporally averaged surface latent heat flux, (b) temporally averaged 10 m latent heat flux, and (c) temporally and vertically averaged snowdrift sublimation rate with 10 m wind speed. The cases without advection and entrainment (standard), with $E_H = 1.13 \times 10^{-6}\ s^{-1}$ (entrainment) and with $A = 8.8 \times 10^{-8}\ s^{-1}$ (advection) are shown. In (a) and (b) the thick lines denote the initial latent heat flux, which are similar to the latent heat flux profiles in the case without snowdrift.

and at 10 m, respectively, averaged over the integration period. In interpreting this figure one should bear in mind that the choice of the length integration period is arbitrary. For instance, if the integration period were longer, the latent heat fluxes at 10 m under the strong wind forcings with either entrainment or advection would be higher. Nevertheless, it nicely demonstrates the potential effect of the processes that remove moisture from the boundary layer on the vertical moisture fluxes. Without advection or entrainment, the simulated upward moisture fluxes rapidly decrease with increasing wind speed and are always smaller than in the case without snowdrift. This is because initially strong snowdrift sublimation rapidly saturates the surface layers resulting in low values of surface and snowdrift sublimation rates (Fig. 11c). When moisture removal is efficient, though, the air remains undersaturated and snowdrift sublimation can continue, enhancing the upward moisture flux at 10 m. In weak wind forcing, the modelled moisture fluxes then become larger compared to the case without snowdrift. Under stronger wind forcing, however, the effect of the huge moisture input by snowdrift sublimation is still dominant, reducing the upward moisture flux compared to the case without snowdrift. Then, most of the moisture flux at 10 m is due to snowdrift sublimation, which qualitatively agrees with the observed moisture budget at Halley during winter (Smith, 1995). In particular the surface moisture flux remains small in all cases under strong winds, indicating that surface sublimation will vanish during snowdrift irrespective of the presence of moisture removal mechanisms. This clearly demonstrates the dominant effect of snowdrift sublimation on the moisture budget near the surface.

The upward moisture flux at 10 m peaks at intermediate values of wind forcing in all cases. This can be attributed to the maximum values of $\hat{S}$, as indicated in Fig. 11c. Imposing efficient moisture removal mechanisms can significantly enhance snowdrift sublimation rates, but the strongest increases occur at intermediate values of wind forcing. This is because at stronger winds the surface layer gets saturated more quickly, even with entrainment or horizontal advection included. Interestingly, in all cases considered here the modelled snowdrift sublimation rate and upward total moisture flux peak at intermediate values of wind strength. This again demonstrates how effective snowdrift sublimation can saturate the atmosphere in strong winds, thereby reducing upward moisture fluxes. King et al. (1996) used the STABLE2 data to show that the winter snowdrift sublimation rate was about 0.7 mm w.e. per month, which translates to a latent heat flux of about 0.8 W m^{-2}. This is a very small fraction of annual accumulation, which clearly contrasts with the finding of Pomeroy et al. (1993) that up to 74% of accumulation ($\sim$70 mm w.e.) is removed through snowdrift sublimation on the Canadian Prairies. Our computational results seem to imply generally low snowdrift and especially surface sublimation rates during snowdrift events.

Vertical profiles of relative humidity, latent heat flux and sublimation rate at the end of the integrations are presented in Fig. 12. The inclusion of additional moisture removing mechanisms cause lower humidity values, higher snowdrift and surface sublimation rates and larger upward moisture fluxes throughout the surface layer. Their presence becomes increasingly more important for longer snowdrifting events. Even in a short run as carried out here entrainment induces almost an order-of-magnitude increase in sublimation rate compared to the standard case. This, obviously, depends strongly on the magnitude of the applied entrainment and horizontal advection fluxes.

The next logical question thus is how realistic are the values used for entrainment and horizontal advection of dry air? Unfortunately, there are not many studies over snow in which these terms are properly quantified. If we regard a surface layer of 10 m depth, an entrainment flux of 2.25×10^{-6} s^{-1} corresponds to a latent heat flux of 75 W m^{-2}. This is an unrealistically high value compared to the initial surface latent heat fluxes (31 W m^{-2} in the strongest wind forcing). As it was chosen so that the effect on snowdrift sublimation would be noticeable, this indicates that realistic variations in moisture entrainment will probably have negligible effect on snowdrift sublimation. A horizontal advection of dry air of 8.8×10^{-8} s^{-1} is equivalent with a latent heat flux of 2.9 W m^{-2}. This is much lower than the applied entrainment fluxes. However, since the advection immediately applies to the entire column (unlike entrainment), its effect is about equally strong on the moisture budget near the surface where the largest snowdrift sublimation rates initially occur. Using ECHAM4 model output, Van den Broeke

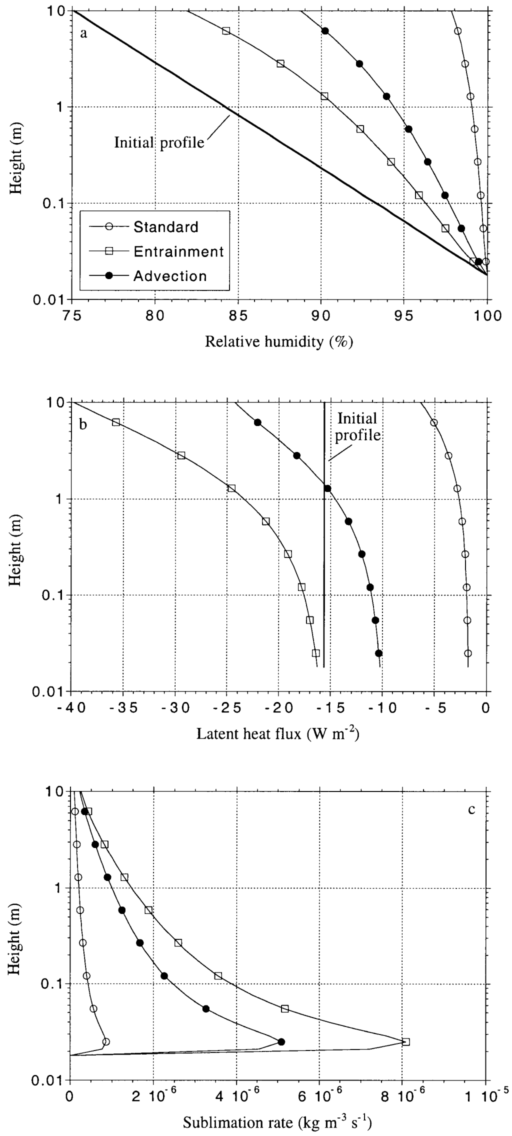

Fig. 12. Simulated vertical profiles of relative humidity, latent heat flux and snowdrift sublimation rate at the end of the 83 min integration for the same cases as in Fig. 11. The thick lines show the initial profiles, which also represent non-snowdrifting conditions.

(1997) finds that maximum values of horizontal advection of moisture in Antarctica are equivalent to about 10 W m^{-2} for a 1000 m thick boundary layer in non-drifting conditions.

Another possible mechanism to obtain lower relative humidities and higher sublimation rates is advection of warm air. Relatively warm air can contain more moisture, as the undersaturation of the air is higher. Especially in Föhn winds or katabatic flows the adiabatic heating is generally largest in the strongest winds, which could thus effectively increase snowdrift sublimation rates. At Halley, the STABLE2 data shows that stronger winds are generally accompanied by higher air temperatures at the surface (Mann et al., 2000), facilitating higher snowdrift sublimation rates. Furthermore, they note that a decrease in fetch, for instance induced by a succession of snow fields and vegetation elements, may reduce the rate at which the air gets saturated. This may also explain why the near-surface air in seasonal snowcovers generally remains undersaturated during snowdrift events (Schmidt, 1982).

It is evident that a parameterization of snowdrift sublimation rate in terms of just the snowdrift transport rate as presented by Schmidt (1982) is an oversimplification of the complex nature of snowdrift sublimation, and therefore cannot be applied generally. This study demonstrates that atmospheric processes affecting the moisture content of the lowest atmospheric layers may have significant effects on the snowdrift sublimation rate. Also the thermodynamic feedback effects associated with snowdrift sublimation and the inherent stratification effects in two-phase flow suggest that a simple parameterization may not be feasible. Mann et al. (2000) calculate snowdrift sublimation rates from the STABLE2 data and present these in a similar way as in Fig. 2. While they show that "observed" snowdrift sublimation rates increase with temperature and wind speed in a manner roughly similar to that in Fig. 2, maximum values are only about 23 W m^{-2} at a temperature of about -10°C and a wind speed of about 14 m s^{-1}. Evidently, this is much smaller than the values given in Fig. 2, which clearly demonstrates how effective the negative feedback effects involved in snowdrift sublimation can be. On the other hand, measurements of latent heat fluxes over seasonal snowcovers are usually larger (Pomeroy and Essery, 1999), indicating that moisture-removal mechanisms may be more important there.

5. Concluding remarks

This paper addresses the sublimation of suspended snow particles and their effect on the moisture budget of the atmospheric surface layer. Calculations have been performed with a one-dimensional surface layer model including snowdrift turbulent diffusion, settling and sublimation. The mechanism of snowdrift sublimation is in principle very efficient, as the total exposed surface of the suspended particles is relatively large and as suspended particles are continuously ventilated. The sublimation of snowdrifting particles induces a source of moisture and a sink of heat, both causing increasing relative humidity values. Because particle sublimation is proportional to the undersaturation of the air, these thermodynamic feedbacks will in turn reduce snowdrift sublimation. Reductions in the particle radius and the eddy exchange coefficients (due to stable particle and thermal stratifications) (Bintanja, 2000b) further strengthen this negative feedback.

With no additional mechanisms included, the simulations show that the extra moisture source of snowdrift sublimation tends to saturate the entire surface layer, starting near the surface. In stronger winds with more suspended particles and high ventilation rates the saturation of the air occurs fastest, reducing the snowdrift sublimation rates to almost zero. Additionally, the modelled surface sublimation rate is reduced to zero as the vertical moisture gradient at the surface vanishes. Thus, surprisingly, our model simulations suggest that the occurrence of an additional moisture producing mechanism of snowdrift sublimation can reduce the total (surface plus snowdrift) sublimation rates to practically zero. Hence, the surface–atmosphere moisture flux may largely vanish during snowdrift events. Interestingly, the integrated (over time) snowdrift sublimation rate peaks at intermediate values of the wind forcing. This is because the increase of snowdrift sublimation with wind speed (due to a higher amount of suspended particles and stronger ventilation rates) is, at higher winds, compensated for by a rapid saturation of the boundary layer and a strong reduction snowdrift sublimation rates. This obviously disagrees with most studies that evaluated snowdrift sublimation made so far, so it is unlikely that snowdrift sublimation is governed by the thermodynamic effect alone. Hence, there are probably other mechanisms that affect the moisture budget and increase sublimation rates.

The process of snowdrift sublimation is very efficient in transforming solid ice into water vapour if the extra moisture would somehow be removed and the ambient air would remain undersaturated. This is investigated by including the moisture removing mechanisms of entrainment and horizontal advection in a simple manner to illustrate their potential effects. We have used simple formulations to investigate the effect of these processes, using such values that the impact on the moisture budget is significant. Obviously, both mechanisms remove moisture from the boundary layer, enabling higher snowdrift sublimation rates than in cases without these terms. The calculations show that in the cases where moisture removal is largest snowdrift sublimation rates are significant as they produce 50–70% of the moisture flux from the surface to the atmosphere (the remainder originates from the surface). Interestingly, in these cases the relative contribution of the surface sublimation is largest (compared to the cases with less efficient moisture removal), indicating that both surface and snowdrift sublimation rates "benefit" from an efficient moisture removal. In the strongest winds considered here, the total (surface plus snowdrift) integrated sublimation is lower than in a similar case without snowdrifting in spite of the additional moisture removal. This clearly demonstrates the effectiveness and potential importance of snowdrift sublimation in reducing the surface–atmosphere moisture flux, as predicted by the model.

It is clear that snowdrift sublimation is a mechanism that cannot be neglected in studies concerning the mass balance of snow-covered regions and the associated boundary layer moisture budget. The model simulations show that the surface ablation and the surface–atmosphere moisture flux can either increase or decrease as a result of the occurrence of snowdrift sublimation, depending on the magnitude of the wind forcing and of the efficiency of removal of the "extra" moisture produced by snowdrift. They also show that this adaptation of the total upward moisture flux is strongest when the surface sublimation is normally found to be largest, namely at high winds with high snowdrift transport rates. It is evident that more snowdrift data supplemented by accurate wind, temperature and humidity data are needed to understand more fully the processes associated with snowdrift sublimation and to validate models about snowdrift sublimation. Many of the features regarding snowdrift sublimation and the moisture budget predicted by the model require observational evidence. Atmospheric models with a realistic treatment of the moisture and heat budgets of the atmospheric boundary layer should be used to realistically quantify the effects of snowdrift sublimation over snow covered regions.

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