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# The significance of vertical moisture diffusion on drifting Snow sublimation near snow surface

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**Abstract.** Drifting snow sublimation is a physical process containing phase change and heat change of the drifting snow, which is not only an important parameter for the studying of polar ice sheets and glaciers, but a significant one for the ecology of arid and semi-arid lands, where snow cover is the main fresh water resource. However, in the previous studies drifting snow sublimation near surface was ignored. Herein, we built a drifting snow sublimation model containing vertical moisture diffusion equation and heat balance equation, to study drifting snow sublimation near surface. The results showed that though drifting snow sublimation near surface was strongly reduced by negative feedback effect, relative humidity near surface didn't reach the saturation state caused by vertical moisture diffusion. Therefore, the sublimation near surface will not stop in drifting snow near surface. The sublimation rate near surface is 3-4 orders of magnitude higher than that at 10 m. And the mass of snow sublimation near surface accounts for even more than half of the total if the wind velocity is small. Therefore, drifting snow sublimation near surface can't be neglected.

## 1 Introduction

The polar ice sheets, mountain glaciers, snowy area in high latitude of Northern Hemisphere (such as North of Canada, Greenland, etc), whose main source is snow, have profound influence on the global hydrologic cycle, climate change and ecological system. Extensive researches showed that drifting snow sublimation was an important method to change the snow distribution, especially in the polar ice sheets, highland mountains and high latitude of Northern Hemisphere. For example, Pomeroy and Jone (1995) found that the mass of drifting snow sublimation was equal to 18.3% of annual precipitation in coastal Antarctica; while Liston and Sturm (2004) found that it was equal to 22% of winter precipitation in Arctic Alaska. Pomeroy and Essery (1999) found that blowing snow sublimation fluxes during blowing snow return 10±50% of seasonal snowfall to the atmosphere in North American prairie and arctic environments. MacDonald et al. (2010) found that the mass of drifting snow sublimation was equal to 17%-19% of annual precipitation in Rocky Mountains, Canada. Zhou et al. (2014) pointed out that the mass of drifting snow sublimation was equal to 24%

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of annual precipitation in western Chinese mountains. These results indicate that drifting snow sublimation is very important to the study of global and polar hydrological systems.

Some scientists directly measured drifting snow sublimation using eddy covariance, but this method can only obtain a few points of information, and it is difficult to predict the whole sublimation in snowy areas (Pomeroy and Essery, 1999; Cullen et al., 2007; Marks et al., 2008; Reba et al., 2012). Therefore, there is a high demand of studying the sublimation of snow using numerical model.

The sublimation of snow particles in the drifting snow is normally accompanied with heat absorption and water vapor production, which will cause a decrease in the ambient air temperature and an increase in humidity. The increased humidity will in turn inhibit the sublimation of snow particles; while the lower temperature will lead to a decrease in the air saturated vapor pressure, which will also inhibit the snow sublimation. Many researchers believed that the sublimation of snow particles near surface would occur violently at the early stage of drifting snow process, since the high concentration of snow particles near surface would result in a rapid air temperature decrease and humidity increase. Then the humidity would reach saturation quickly near surface, and the sublimation would stop at the saturated layer of humidity. Therefore, the snow sublimation near surface was negligible in the fully developed drifting snow (Déry et al., 1998; Bintanja, 2001a; Mann et al., 2000). However, some researchers found that humidity near surface didn't reach saturation in the drifting snow in the field or wind tunnel experiments, which they thought was caused by water transport (convection and diffusion) (Schmidt, 1982; Groot Zwaadtink et al., 2011). Déry and Yau (1999) fix the relative humidity at 95% instead of 100% at the surface when they simulated the blowing snow sublimation. They found that the time-integrated values of sublimation increased 14% than the results which fix the relative humidity at 100%, so humidity near surface is very important for the simulations of blowing sublimation. Huang et al. (2016) calculated the snow sublimation in the saltation layer, taking into consideration of the effect of horizontal moisture convection on the non-homogeneous snow cover. Their results showed that drifting snow sublimation in the saltation layer could not be neglected in the presence of horizontal moisture convection. But they did not discuss the sublimation near surface of areas such as polar ice sheets, grassland covered by snow, etc., where the snow cover was very large and the water convection was very weak. Therefore, studies on the snow-sublimation in these regions are of great significance for the understanding of global hydrological systems and ecosystems.

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However, in the previous blowing snow sublimation model, the diffusion equation was often used to describe the movement of snow particles, which can describe the movement of small particles well. But the diffusion equation is difficult to describe the movement of large snow particles which are mainly distributed in the near surface area (Déry et al., 1998; Xiao et al., 2000; Vionnet et al. 2014). Huang et al. (2016) used the Lagrangian particle tracing method to describe the movement of near-surface snow particles, and for the first time calculated the sublimation of saltation particles in near surface region on non-uniform snow cover. But this model can not describe snow particles suspending in upper air. Furthermore, all above exiting models did not take into consideration of the effects of vertical moisture diffusion on the sublimation.

Therefore, a drifting snow model has firstly been built to describe the movement of snow particles of both saltating near surface and suspending in the higher region. Then, a drifting snow sublimation model has been built the combination of the drifting snow model, a vertical moisture diffusion equation and a heat balance equation. Then drifting snow sublimation with three wind speeds was calculated. The temporal evolution and vertical profiles of temperature, relative humidity, mass concentration of snow particles, snow sublimation rate were analyzed in details. Meanwhile, the proportions of the sublimation mass of saltation snow grains and saltation layer to the total sublimation mass were also given.

# 2 Method

## 2.1 Basic Equations of the Flows

The horizontal wind field satisfies the Navier–Stokes equation at the atmospheric boundary layer. Considering a fully developed steady flow field on an infinite polar ice sheet where the changes of wind field in the lateral and flow direction are negligible, the fully developed horizontal direction flow field equation can be obtained according to the theory of mixing length by Prandtl.

$$\frac{\partial}{\partial z}(\rho_a \kappa^2 z^2 \left|\frac{du}{dz}\right| \frac{du}{dz}) + F = 0 \quad (1)$$

where $\kappa$ is the von Karman constant, $\rho_a$ is air density, u is the horizontal wind speed and F is the reaction force of the snow particle on the flow field.

### 2.2 Snow particle motion equation

The snow particles jumping from the bed are divided into saltation and suspended particles when calculating snow particle movement. These two types of particles are distinguished based on the particle size and flow field conditions. Then the saltation particles are calculated by Lagrange particle tracing method, and the suspension particles are calculated by diffusion equation.

#### 2.2.1 Judging criteria of saltation and suspended particles

The judging criterion of saltation and suspended particles is as follows (Scott, 1995):

$$\begin{cases} w_s/(ku_*)>1, & \text{saltation particle} \\ w_s/(ku_*)\le 1, & \text{suspension particle} \end{cases} \tag{2}$$

where $u_*$ is the friction velocity and $w_s$ is the final sedimentation velocity of the particles (Carrier, 1953):

$$\begin{aligned} & w_s = -\frac{A}{D} + \sqrt{\left(\frac{A}{D}\right)^2 + BD} \\ & A = 6.203\upsilon_a \\ & B = \frac{5.516\rho_p}{8\rho_a} g \end{aligned} \tag{3}$$

where D is diameter of snow particle, $\upsilon_a$ is air viscosity coefficient, $\rho_p$ is the densities of snow particle, $g$ is the acceleration of gravity.

#### 2.2.2 Basic equations of saltation particles

Saltation particle motion equation is as follows (Huang et al., 2011):

$$m\frac{dU_p}{dt} = F_D\left(\frac{U_a - U_p}{V_r}\right) \tag{4}$$

$$m\frac{dV_p}{dt} = -G + F_B + F_D\left(\frac{V_a - V_p}{V_r}\right) \tag{5}$$

$$\frac{dx_p}{dt} = U_p \tag{6}$$

$$\frac{dy_p}{dt} = V_p \tag{7}$$

where $m$ is the mass of snow particle, $G$ is the gravity of snow particle, $U_a$ and $V_a$ are the horizontal and vertical velocity of air, $U_p$ and $V_p$ are the horizontal and vertical velocities of snow particle, $V_r = \sqrt{(U_p - U_a)^2 + (V_p - V_a)^2}$ is the relative velocity of movement of the snow particles and the flow field, $F_B$ and $F_D$ are the buoyancy and traction forces of snow particles, $x_p$ and $y_p$ are the horizontal and vertical positions of snow particles.

The splash function fitted by Sugiura and Maeno (2000) according to the observations of the low temperature wind tunnel experiment was chosen,

$$S_v\left(e_v\right) = \frac{1}{b^a G\left(a\right)} e_v^{a-1} \exp\left(-\frac{e_v}{b}\right) \tag{8}$$

$$S_h\left(e_h\right) = \frac{1}{\sqrt{2\pi\sigma^2}} \exp\left[-\frac{\left(e_h - \mu\right)^2}{2\sigma^2}\right] \tag{9}$$

$$S_e\left(n_e\right) = {}_mC_{n_e} p^{n_e} \left(1 - p\right)^{m-n_e} \tag{10}$$

where $S_v\left(e_v\right)$, $S_h\left(e_h\right)$ and $S_e\left(n_e\right)$ are the probability distribution functions of the vertical restitution coefficient $e_v$, horizontal restitution coefficient $e_h$, and the number of grains ejected $n_e$.

### 2.2.3 Basic Equations of Suspended particles

The movement of suspension particles is described by the following vertical diffusion equation according to horizontal uniformity condition,

$$\frac{\partial q}{\partial t} = \frac{\partial}{\partial y}\left(K_s \frac{\partial q}{\partial y} + w_s q\right) + S \tag{11}$$

where q is the snow particle mass concentration, $K_s$ is the vertical diffusion coefficient, S is the volume sublimation rate of snow grain. $K_s = \delta \kappa u_* z$, $\delta$ is as follows (Csanady, 1963),

$$\delta = \frac{1}{\sqrt{1 + \frac{\beta^2 f^2}{w_a{}^2}}} \tag{12}$$

where $\beta$ is the proportionality constant, $w'$ is the turbulent fluid velocity in the vertical, and we set

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$\beta = 1$, $\overline{w'^2} = u_*^2$ ..

### 2.2.4 Aerodynamic Entrainment

The aerodynamic entrainment equation of Shao and Li (1999) is chosen,

$$N_a = V u_* \left( 1 - \frac{u_{*t}^2}{u_*^2} \right) D^{-3} \tag{13}$$

where $N_a$ is the number of snow particles taking off causing by aerodynamic entrainment, $\varsigma$ is a non-dimensional coefficient, approximately equal to $1\times10^{-3}$, $u_*$ is the friction velocity, $u_{*t}$ is the threshold friction velocity.

## 2.3 Sublimation formula

The sublimation formula is as follows (Thorpe and Mason, 1966),

$$\frac{dm}{dt} = \frac{\pi D(RH-1)}{\frac{L_s}{KNuT_a}\left(\frac{L_s}{R_v T_a}-1\right)+\frac{R_v T_a}{ShK_l e_s}} \tag{14}$$

where $RH$ is the relative humidity of air, $T_a$ is air temperature, $L_s$ is the latent heat of sublimation (equal to $2.84\times10^6$ J kg$^{-1}$), $K_a$ is the thermal conductivity of air, $R_v$ is the gas constant of water vapor (equal to 461.5 J kg$^{-1}$ K$^{-1}$), $K_l$ is the molecular diffusion of water vapor of atmosphere, $e_s$ is the saturated vapor pressure relative to the ice surface. $Nu$ and $Sh$ are the Nusselt and Sherwood numbers (Thorpe and Mason, 1966; Lee, 1975),

$$Nu = Sh = \begin{cases} 1.79 + 0.606\,\mathrm{Re}^{0.5} & 0.7 < \mathrm{Re} \le 10 \\ 1.88 + 0.580\,\mathrm{Re}^{0.5} & 10 < \mathrm{Re} < 200 \end{cases} \tag{15}$$

where $R_e = \frac{DV_r}{\upsilon_a}$ is Reynolds number.

## 2.4 Heat and humidity equation

The heat and humidity equations of air are as follows (Déry and Yau, 1999; Bintanja，2000),

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left( K_\theta \frac{\partial\theta}{\partial z} \right) - \frac{L_s S}{\rho_f C} \tag{16}$$

$$K_\theta = \kappa u_* z + K_T \tag{17}$$

$$\frac{\partial h_u}{\partial t} = \frac{\partial}{\partial z}\left(K_q \frac{\partial h_u}{\partial z}\right) + \frac{S}{\rho_f} \tag{18}$$

$$K_h = \kappa u_* z + K_V \tag{19}$$

where $K_T$ and $K_V$ are the molecular diffusion coefficients of heat and water vapor, C is the specific heat of air.

**2.5 Initial and boundary conditions**

The initial potential temperature $\theta_0 = 263.15K$, and the initial absolute temperature is

$$T_0 = \theta_0 \left(\frac{p}{p_0}\right)^{0.286} \tag{20}$$

Where p is atmospheric pressure, its initial value is

$$p = p_0 \exp\left(-\frac{yg}{R_d \theta_0}\right) \tag{21}$$

where $p_0 = 1000hpa$, $R_d = 287JKg^{-1}K^{-1}$ is the gas constant for dry air.

The initial relative humidity profile is

$$RH = 1 - R_s \ln(z / z_0) \tag{22}$$

where $z_0$ is the surface roughness, and its value is $3\times10^{-5} m$ at snow bed (Nemoto and Nishimura, 2001), and $R_S = 1.9974\times10^{-2}$.

The conversion relationship of relative humidity and specific humidity is

$$q = 0.622 \cdot \frac{e_s}{p - e_s} \cdot RH \tag{23}$$

where $e_s = 610.78\exp\left[21.87(T - 273.16)/(T - 7.66)\right]$.

The calculation area is set to 1 m in length, 10 m in height, and 0.01 m in width. The time step is $10^{-5}$ s for saltation particles, $10^{-2}$ s for suspended particles, $10^{-3}$ s for wind, and the calculation time is 1500 s. The motion of saltation particles is only calculated for 10 s in consideration of the practical simplicity, since saltation particles will stabilize within a few seconds. The data of saltation particles in the air and the jumping particles from bed are then replaced by the data averaged in 10 s. The threshold friction velocity is 0.21 m/s (Nemoto and Nishimura, 2001).

The snow particle size distribution fits the results of Schmidt (1982) field observations (Fig. 1).

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## 3 Results and Discussion

In order to verify the judging criteria in eq.2, we divided the particles into sets varied by 10 $\mu m$ (1-600 $\mu m$), and used eq.16 to simulate all the jumping particles. Then we accumulated the mass of snow particles in the air from small to large particles until the mass was equal to 99.9% of the total mass of snow particles in the air, and the particle diameter $D_{99\%}$ was recorded. $D_{99\%}$ and threshold particle diameter $D_{th}$ calculated by eq.2 were compared, and the results is shown in Table1.

As shown in Table 1, particles which are larger than the threshold particle do not enter into air according to the vertical diffusion, indicating that these particles can not be described by the diffusion equation. Thus, the judging criteria in eq.2 are reliable.

In order to verify the reliability of the blowing snow model in this paper, we compared our mass concentration results with that of the field observations (Fig.2). The red dot in Fig. 2 is the field observation results near Saskatoon, Canada in 26 January 1987 (Pomeroy and Male, 1992). And the black line in Fig.2 is our numerical simulation results using the same conditions with the above filed observation results. It is shown that our simulation results are basically consistent with those observed in the field, which demonstrates the reliability of our simulations.

We also compared our sublimation results with that of the field observations to verify their reliability (Fig.3). The red lines in Fig. 3 are the results gotten from the observed data by Schmidt (1982) in Wyoming, U.S.A, in 1982. The black line was the simulated results using the same environmental conditions as those of Schmidt's. It can be seen that the total sublimation rates calculated by the model of this paper (black line) are approximately the same as Schmidt's results, and the sublimation rate at 0.01 m was two orders of magnitude larger than that at 0.1 m. These results demonstrate that our snow sublimation results are reliable too.

We further compared our results with corresponding results of other models under the same conditions. The black line in Fig. 4 is the result of the suspension particles sublimation rate calculated by our model ( $u_* = 0.89, T = 253.15K$ ). The other four lines are the results calculated by Xiao et al. (2001) using four existing blowing snow sublimation models, in which the sublimation of saltation particles near surface was neglected. It is shown from Fig. 4 that all the rates of suspension particle increase with height first, and then start to decrease, and the peak is at about 0.1 m. The results of this paper are higher than that of Xiao et al. (2001). The peaks of total sublimation rate using our model and Schmidt (1982) are all at a height about 0.01 m, which is lower than that of the four blowing snow models in Fig. 4. But the values of peak in this paper and Schmidt (1982) are two orders of magnitude larger than that of the four blowing snow models. This is because the sublimation of saltation particles is neglected in the four models, which is the main movement of snow particles near surface.

Fig. 5 is the temporal evolution of the mass of saltation particles and suspended particles versus various friction velocities. It is shown that the mass of saltation and suspended particles increase with time, and finally reach steady. The mass of saltation particles is much larger than that of suspension particles in the steady state. The time for saltation particles to reach steady state is about 2 s, and about 300 s for suspended particles.

Fig. 6 shows the curves of temperature and humidity with height in the near-surface region of saltation particles and they are compared with their initial conditions. It is shown that drifting snow sublimation changes air temperature and relative humidity, and the change amplitude increases with the friction velocity. This is because the larger the friction velocity is, the more snow particles in the air are, and the more sublimation will occur, which makes a greater impact on temperature and humidity.

We compared the temperature and humidity with height. It is shown in Fig. 7 and 8 that the change amplitude of temperature and relative humidity increases while the height decreases. Combined with the results from Fig. 10, the mass concentration of snow particles increases while height decreases, which can make a stronger sublimation.

It is shown in Fig. 8 that the time for humidity to reach steady is about 2 s at 0.01 m, which is consistent with the stability time of saltation snow particles; and at 10 m is about 300 s, which is consistent with the stability time of suspension snow particles. This is because the main part of snow particles near surface is saltation particles, opposite to that in upper air which is mainly suspension particles (Fig. 10).

Fig. 8 shows that the relative humidity near surface with three kinds of friction velocities does not reach saturation when the blowing snow reaches steady, which indicates that the snow sublimation does not stop. It also shows that the vertical diffusion of water vapor can reduce the negative feedback effect effectively.

It can be seen from Fig. 9a that the sublimation rate of saltation particles increases with time first, then starts to decrease, in which the peak is at about 2 s and finally reaches stability at about 300 s. The negative feedback effect on saltation particles is very obvious and the time to reach a steady state is about 300 s. Because the mass of saltation particles increases with time during the first 2 s, and the increasing amplitude of which is larger than that of relative humidity, and the saltation sublimation rate increases with time. However, the mass of saltation particles basically stay unchanged after 2 s, while the relative humidity near surface gradually increases. Therefore, the sublimation rate decreases with time. The relative humidity near surface also reaches steady after 300 s, which results in the stability of sublimation rate. The saltation particles distribute mainly near surface, where the change amplitude of relative humidity is strong which results in a strong negative feedback effect on saltation particles.

It is shown in Fig. 9b that sublimation rate of suspended particles increases with time and finally reaches steady at about 300 s. The negative feedback effect on suspended particles is not obvious. The mass of suspension particles increases with time during the first 300 s, which the increase amplitude of is larger than that of relative humidity, so the suspension sublimation rate increases with time. Then the mass of suspended particles and relative humidity both reach stable, which leads to the sublimation rate of suspended particles reaching stable. Since the suspended particles mainly distribute in upper air where the change amplitude of relative humidity is weak, the negative feedback effect on suspended particles is not strong.

Although the effect of negative feedback on saltation particles is stronger than suspended particles, the sublimation rate of saltation particles is still greater than that of suspended particles, indicating that the sublimation of saltation particles is very strong even under the effect of negative feedback.

Fig. 10 shows that the mass concentration of snow particles increases with friction velocity and decreases with height, and the mass concentration of saltation particles is much higher than that of suspended particles. It can be seen from Fig. 10a that saltation particles mainly distribute below 0.1 m, which is consistent with the previous experimental results (Takeuchi, 1980).

Fig. 11 shows that sublimation rates increases with friction velocity. Sublimation rates of saltation and suspended particles increase with height first, then start to decrease. The peak of saltation particles is at about 0.01 m, and that of suspended particles is at about 0.1 m. This is because the mass concentration and relative humidity of snow decrease with height, while temperature increases. However, mass concentration of saltation particles changes more strongly than that of suspension particles with height. Therefore, sublimation rate of saltation particles reaches peak at lower height.

Table 2 shows that the sublimation rate at 0.01 m is two orders of magnitude faster than that at 0.1 m, which is same as the experimental results in Fig. 3, and it's 3-4 times faster than that at 10 m, although the negative feedback effect near surface is stronger than other regions. Because the mass concentration of snow particles near surface is much higher than that in other regions (Fig. 8), and water vapor near surface is not saturated, the sublimation rate near surface is much faster than that in other regions.

In the previous studies the snow sublimation near surface was ignored. That is, to define a wind velocity related height, below which saltation particles move. Then assumed that moisture in the region was saturated and therefore the snow sublimation would not be counted (Déry et al., 1998; Xiao et al. 2000; Vionnet et al. 2014). Three heights at several wind velocities proposed by Déry et al. (1998), Pomeroy and Male (1992), and Xiao et al. (2000) were respectively given in Table 3 (The height by Vionnet et al. was the same as that of Pomeroy and Male). Fig. 12 shows the actual ratio of our simulated sublimation mass below the three heights to the total. It is shown that all the sublimation masses below three heights account for more than half of the total sublimation mass. This is because the main part of snow particles is saltation particles (Mellor, 1965), which mainly distribute in near surface region. And although sublimation near surface leads to significant changes of temperature and humidity, which have a strong inhibition effect on sublimation, moisture near surface does not reach saturation due to the vertical diffusion of water vapor, which results in continuous snow sublimation. Therefore, the main part of the mass of sublimation is sublimation of saltation particles, and the previous methods neglecting blowing snow sublimation near surface is not appropriate. Fig. 12 also shows that the proportion of the sublimation mass near surface decreases with friction velocity. Because more snow particles can enter into upper air with increased wind velocity, which will lead to decreasing proportion of snow particles near surface, the proportion of the mass of sublimation near surface will decrease as well.

Fig.13 shows the vertical profiles of vapor flux. It is shown that vapor flux increases rapidly in near surface region, where most of saltation particles move, then slows down greatly after reaching a certain height. For there is no horizontal flux of water vapor， the water vapor flux at any height must be equal to the total amount of water vapor generated per second below the height. So most of the water vapor is coming from near surface regions. From Fig. 13 it can also be seen that vapor flux increases with friction velocity, for humidity (Fig.5) and moisture diffusion coefficient (eq.17) increase with friction velocity.

## 4 Conclusions

We have established a blowing snow sublimation model, which includes vertical moisture diffusion and heat balance, to study the snow sublimation near surface in large snow cover area in this paper. The simulation results showed that the blowing snow sublimation decreased air temperature and increased humidity of air. Meanwhile, the snow sublimation was reduced by the negative feedback effect of temperature and humidity, especially for near surface, which is in agreement of previous researches. However, moisture near surface was not saturated due to the vertical moisture diffusion, so snow sublimation near surface continued. The sublimation rate near surface was even larger than that in the upper air, because mass concentration of snow particles near surface was much higher than that in other regions. The sublimation rate at 0.01 m is two orders of magnitude greater than that at 0.1 m, and is 3-4 orders of magnitude greater than that at 10 m. Furthermore, when the wind speed was low, the mass of sublimation near surface accounted for more than half of total mass of sublimation, and could not be neglected. Most of the air vapor in bellowing snow is form near surface region. Therefore, blowing snow sublimation near surface should be taken seriously in the study of snow sublimation and water vapor transport in the future.

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**Table 1: Comparison of $D_{th}$ and $D_{99\%}$**

| | $u_* = 0.35ms^{-1}$ | $u_* = 0.41ms^{-1}$ | $u_* = 0.54ms^{-1}$ |
|---|---|---|---|
| $D_{th}$ | 80.55μm | 87.84μm | 102.61μm |
| $D_{99\%}$ | ≤80μm | ≤90μm | ≤110μm |

**Table 2: Sublimation rate at 1500s for various heights (*: friction velocity (m/s); **: height (m); ***: sublimation rate (kgm$^{-3}$s$^{-1}$))**

| | $u_* = 0.35ms^{-1}$ | $u_* = 0.45ms^{-1}$ | $u_* = 0.55ms^{-1}$ |
|---|---|---|---|
| **h=0.01**** | 3.71E-04*** | 4.05E-04 | 4.21E-04 |
| **h=0.05** | 1.22E-05 | 2.31E-05 | 3.18E-05 |
| **h=0.1** | 6.11E-07 | 3.08E-06 | 5.37E-06 |
| **h=1** | 1.68E-07 | 1.12E-06 | 2.29E-06 |
| **h=5** | 2.93E-08 | 2.88E-07 | 7.52E-07 |
| **h=10** | 8.44E-09 | 1.09E-07 | 3.31E-07 |

**Table 3: Height which most of saltation particles distributed below for various friction velocities**

| | $u_* = 0.35ms^{-1}$ | $u_* = 0.45ms^{-1}$ | $u_* = 0.55ms^{-1}$ |
|---|---|---|---|
| **Déry et al. (1998)** | 0.0196m | 0.0253m | 0.0316m |
| **Pomeroy and Male(1992)** | 0.0222m | 0.0306m | 0.0395m |
| **Xiao et al.(2000)** | 0.05m | 0.05m | 0.05m |

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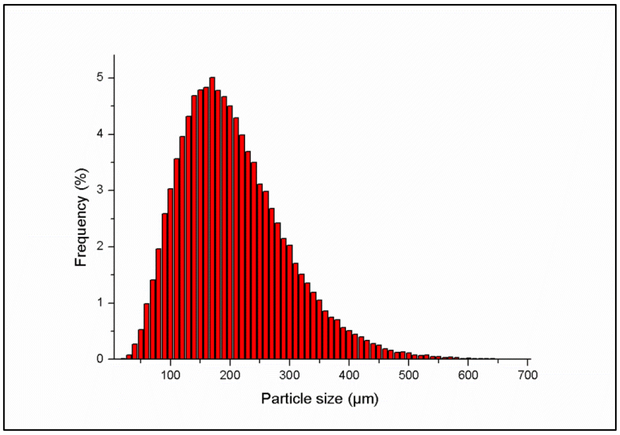


**Figure 1: Particle size distribution**

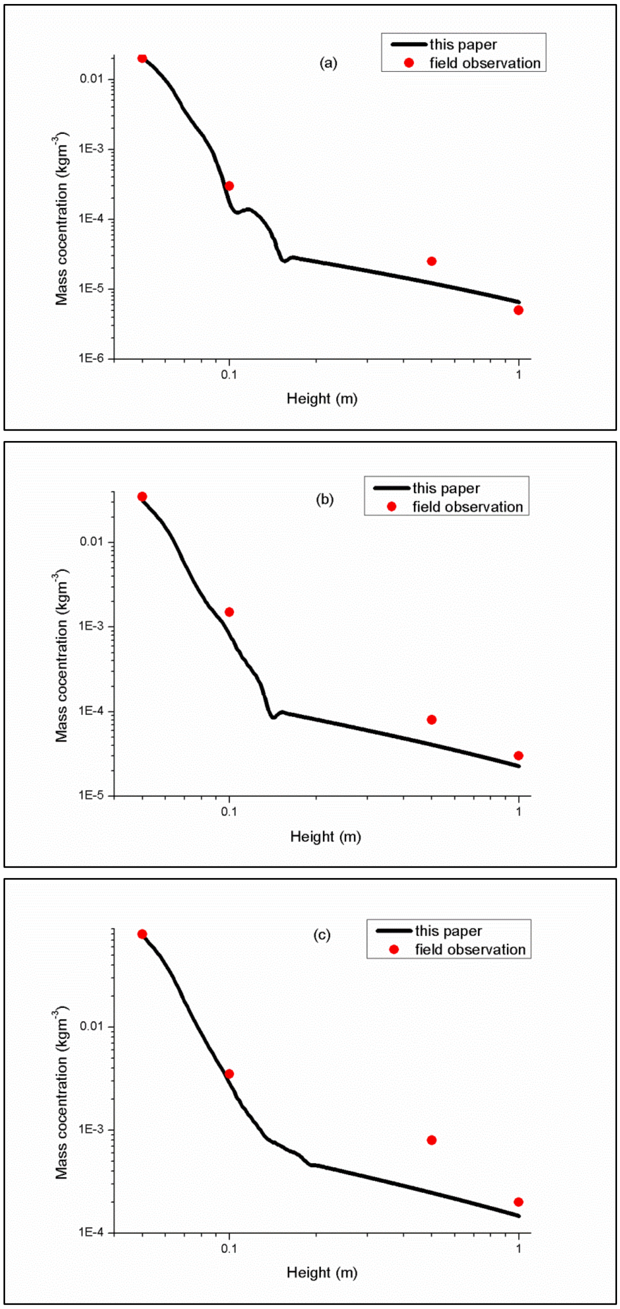


**Figure 2: Comparison of mass concentration for this paper and field observation (a: $u_* = 0.35ms^{-1}$ ; b: $u_* = 0.41ms^{-1}$ ; c: $u_* = 0.54ms^{-1}$ )**

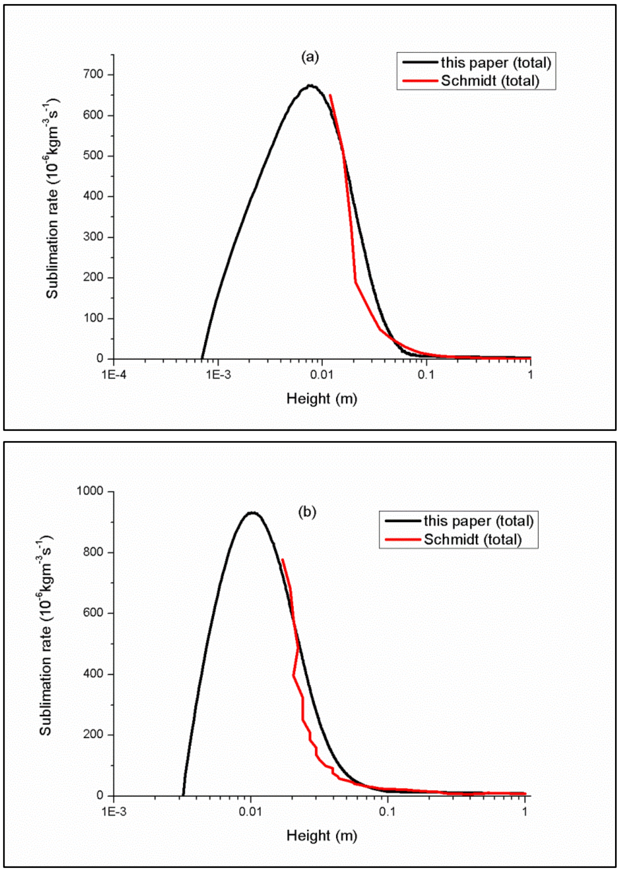


**Figure 3: Comparison of sublimation rate for this paper and Schmidt (1982) (a:** $u_* = 0.632ms^{-1}, T = 267.45k$ **;**

**b:** $u_* = 1.072ms^{-1}, T = 265.65K$**)**

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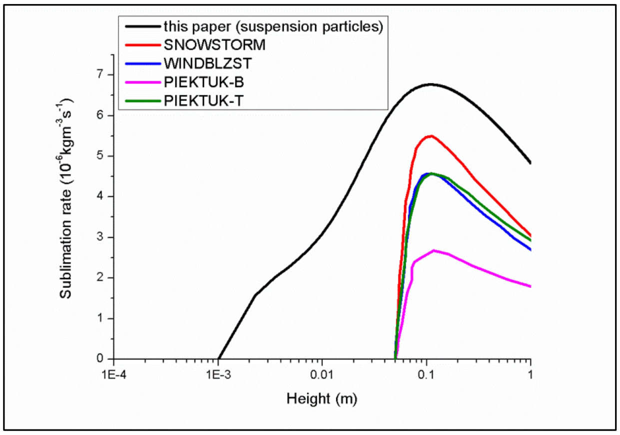


**Figure 4: Comparison of sublimation rate for this paper and four blowing snow's models (Xiao et al., 2000)**

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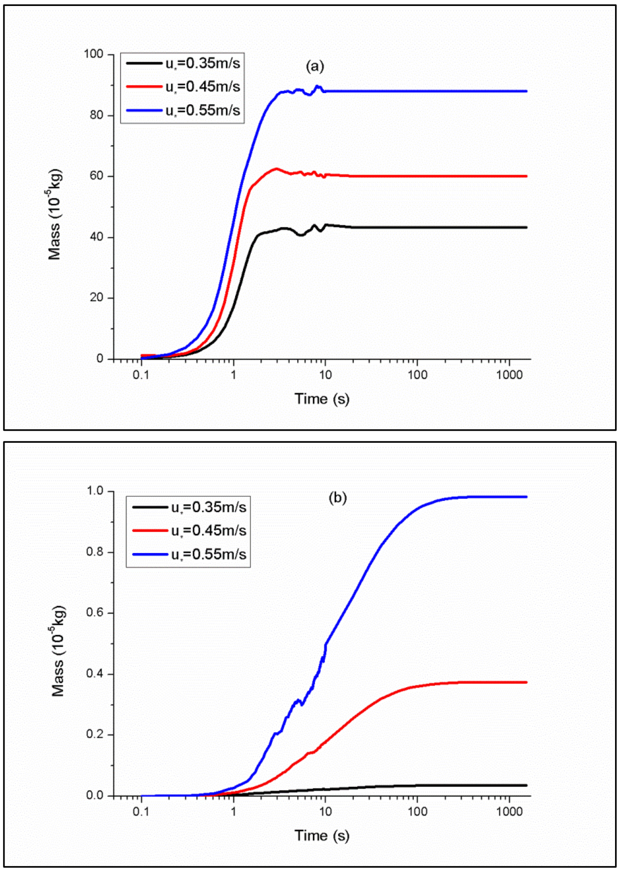


**Figure 5 : Temporal evolution of mass of saltation particles and suspension particles (a: saltation particles; b: suspended particles)**

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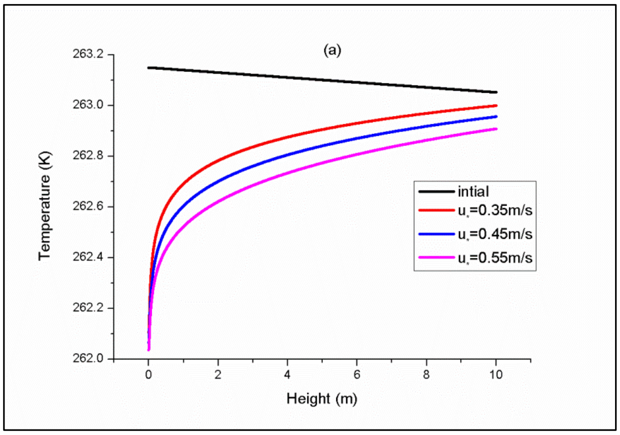


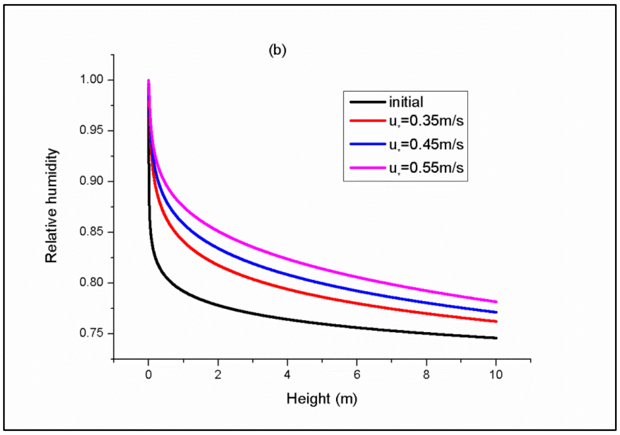


**Figure 6: Vertical profiles of temperature and relative humidity**

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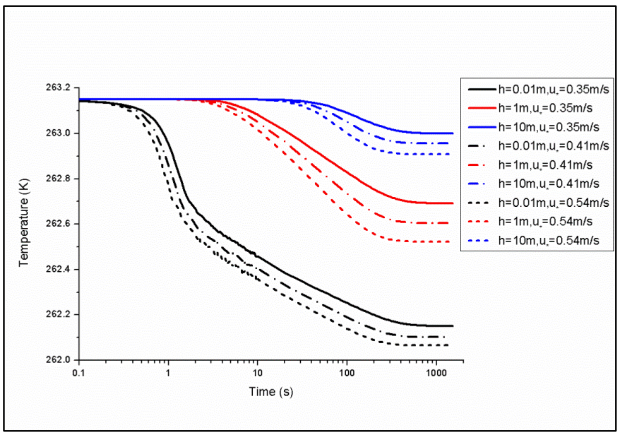


**Figure 7: Temporal evolution of temperature for various heights**

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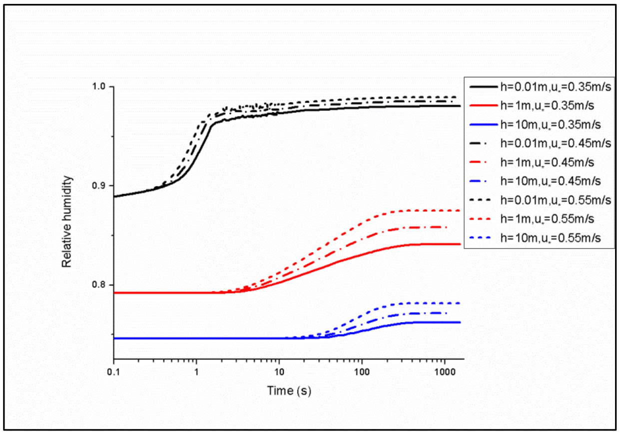


**Figure 8: Temporal evolution of relative humidity for various heights**

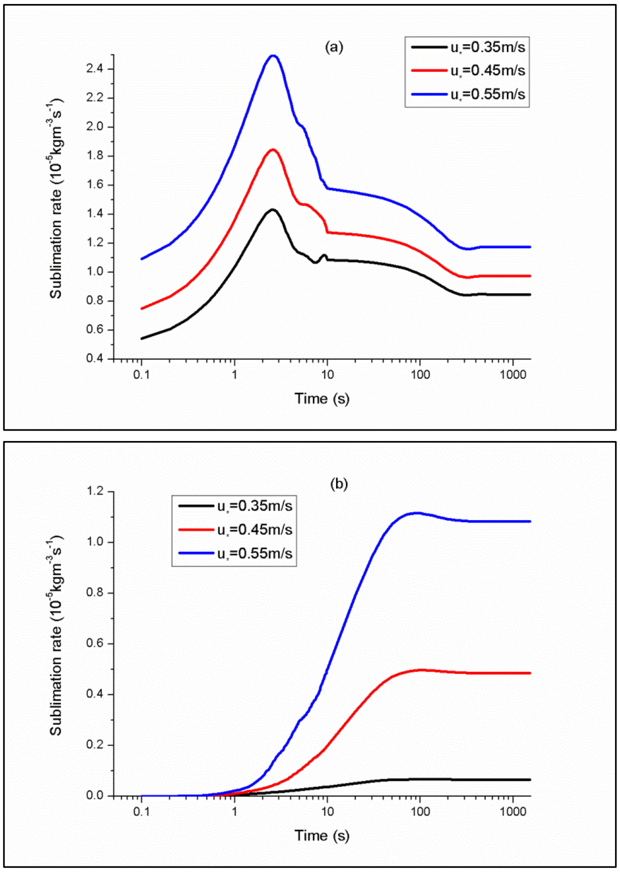


**Figure 9: Temporal evolution of saltation sublimation rate and suspension sublimation rate(a: saltation particles; b: suspended particles)**

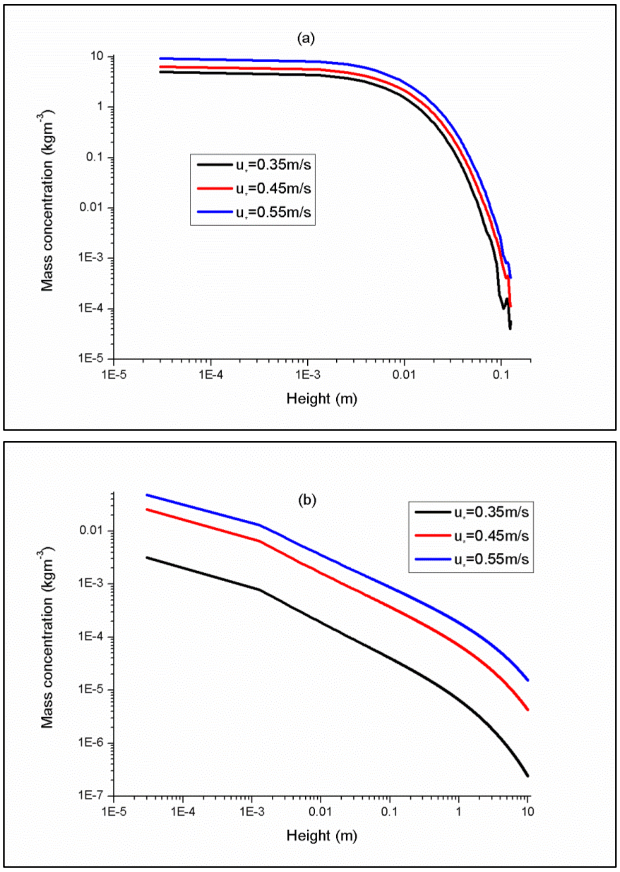


**Figure 10: Vertical profiles of mass concentration for saltation and suspension (a: saltation particles, b: suspended particles)**

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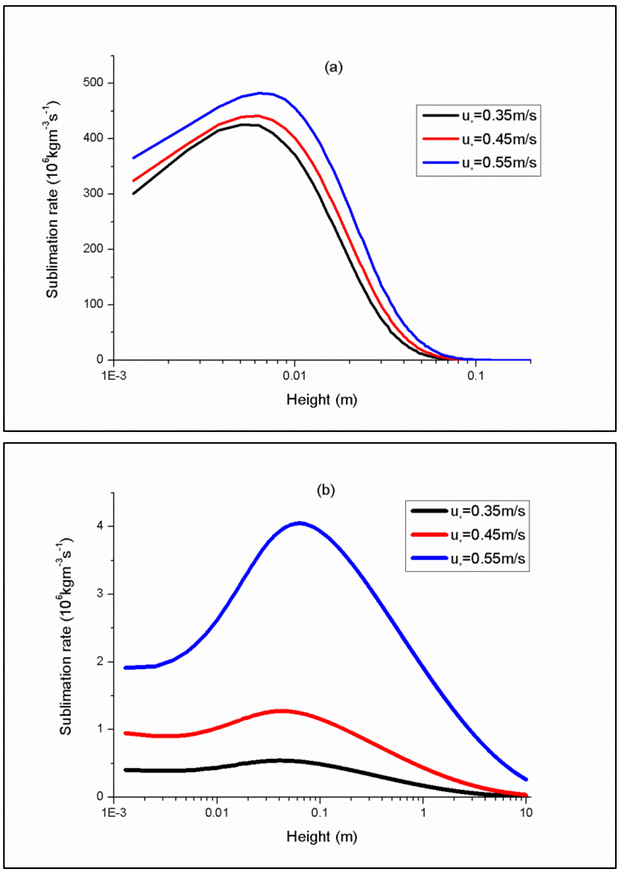


**Figure 11: Vertical profiles of sublimation rate for saltation and suspension (a: saltation particles; b: suspended particles)**

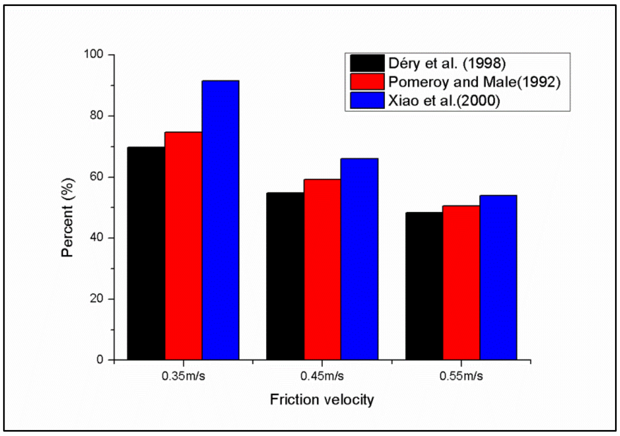


**Figure 12: The ratio of sublimation mass below three heights to the total (the sublimation mass below a height is the sublimation mass that was ignored by other's model , such as Déry et al. (1998), Pomeroy and Male (1992), and Xiao et al. (2000).)**

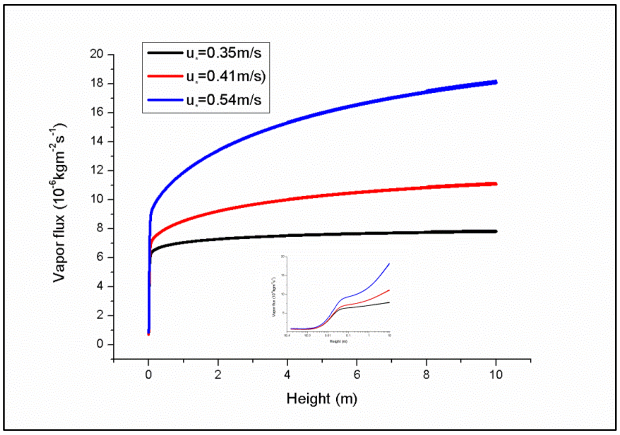


**Figure 13: Vertical profiles of vapor flux**