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

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Abstract. Sublimation of blowing snow is an important parameter not only for the studying of polar ice sheets and glaciers, but also for maintaining the ecology of arid and semi-arid lands. However, sublimation of blowing snow near surface is often ignored in the most of previous studies. To study sublimation of blowing snow near surface, we established a sublimation of blowing snow model containing both vertical moisture diffusion equation and heat balance equation. The results showed that although sublimation of blowing snow near surface was strongly reduced by negative feedback effect, due to vertical moisture diffusion, the relative humidity near surface doesn't reach 100%. Therefore, the sublimation of blowing snow near surface will not stop. In addition, the sublimation rate near surface is 3-4 orders of magnitude higher than that at 10 m above the surface and the mass of snow sublimation near surface accounts for even more than half of the total snow sublimation when the friction wind velocity is less than about 0.55 m/s. Therefore, sublimation of blowing snow near surface should not be neglected.

1 Introduction

Blowing snow is the main source of polar ice sheets and mountain glaciers at snowy area with high latitude in the Northern Hemisphere (such as north of Canada, Greenland, etc), which have profound influence on the global hydrologic cycle, climate change and ecological system. Extensive studies have showed that sublimation of blowing snow is an important method to change the snow distribution, especially in the polar ice sheets, highland mountains and areas with high latitude in Northern Hemisphere. It has been shown the mass of sublimated blowing snow was equal to 18.3% of annual precipitation in coastal Antarctica (Pomeroy and Jone, 1995), 22% of winter precipitation in Arctic Alaska (Liston and Sturm, 2004), 17%-19% of annual precipitation in Rocky Mountains, Canada (MacDonald et al. 2010), and 24% of annual precipitation in western Chinese mountains (Zhou et al. 2014). In addition, the fluxes of sublimated blowing snow sublimation fluxes during blowing snow returned 10±50% of seasonal snowfall to the atmosphere in North American prairie and arctic environments (Pomeroy and Essery, 1999). These results indicate that sublimation of blowing snow is very important for studying of global and polar hydrological systems.

Some scientists (Pomeroy and Essery, 1999; Cullen et al., 2007; Marks et al., 2008; Reba et al., 2012) used eddy covariance to directly measure sublimation of blowing snow. However, since this method can only obtain information from a few points, it is difficult to be used 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, studying the sublimation of snow using numerical model is highly demanded.

The sublimation of blowing snow particles is normally accompanied with heat absorption and water vapor production, which will lead to decreased ambient air temperature and increased in humidity. The latter will in turn inhibit snow sublimation, and the former will decrease the saturated vapor pressure in the air, and subsequently inhibit the snow sublimation. Many researchers (Déry et al., 1998; Bintanja, 2001a; Mann et al., 2000) believed that the sublimation of snow particles near surface would be significant at the early stage of drifting snow process. However, the high concentration of snow particles near surface would result in a rapid air temperature decrease and humidity increase. Therefore, the humidity near surface would quickly reach saturation, leading to sublimation ceasing in the layer with saturated humidity. Therefore, the sublimation of snow particles near surface was negligible in the fully developed drifting snow (Déry et al., 1998; Bintanja, 2001a; Mann et al., 2000). However, some researchers (Schmidt, 1982; Groot Zwaadtink et al., 2011) found that humidity near surface didn't reach saturation in the drifting snow in the field or wind tunnel experiments and believed that caused by water transport (convection and diffusion). Déry and Yau (1999) fix the relative humidity at 95% instead of 100% at the surface when simulating the blowing snow sublimation and found that the time-integrated values of sublimation increased by 14% than at 95% relative humidity compared with that at 100% relative humidity. So they believed that humidity near surface is very important for the simulations of blowing snow sublimation. Huang et al. (2016) calculated the snow sublimation in the saltation layer by taking into consideration of the effect of horizontal moisture convection on the non-homogeneous snow cover. Their results showed that sublimation of blowing snow 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, snow-covered grassland, 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.

However, in the previous blowing snow sublimation model, the diffusion equation was often used to describe the movement of snow particles. Although the equation is good on describing the movement of small particles well, but it 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 saltating particles in near surface region with non-uniform snow cover. But this model did not take into consideration of turbulent suspension of snow particles. Furthermore, all the above exiting models did not take into consideration of the effects of vertical moisture diffusion on the sublimation.

In this study, a drifting snow model was first established to describe the movement of snow particles of both saltating snow particles near surface and suspended snow particles in the higher region. Then, a sublimation model of blowing snow was built in combination of the drifting snow model, a vertical moisture diffusion equation and a heat balance equation. Next, sublimation of blowing snow at three different wind speeds was calculated and the temporal evolution and vertical profiles of temperature, relative humidity, mass concentration of snow particles and snow sublimation rate were analyzed in details. At last, the proportions of the sublimation mass of snow particles near surface to the total sublimation mass were also given.

2 Methods

2.1 Basic flow equations

The horizontal wind field satisfies the Navier–Stokes equation at the atmospheric boundary layer (Nemoto and Nishimura, 2004).

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

where κ is the von Karman constant, ρ_a is air density, u is the horizontal wind speed and F is the reaction force of the snow particles on the flow field.

2.2 Snow particle motion equation

The snow particles jumping from the bed are divided into saltating 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 saltating particles are calculated by Lagrange particle tracing method, and the suspended particles are calculated by diffusion equation.

2.2.1 Judging criteria of saltating and suspended particles

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

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

where u_* is the friction velocity and w_s is the final sedimentation velocity of the particles which can be calculated by the following equations (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, υ_a is air viscosity coefficient, ρ_p is the density of snow particles, g is the acceleration of gravity.

2.2.2 Basic equations of saltating particles

The motion equation of the saltating particles 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, respectively, U_p and V_p are the horizontal and vertical velocities of snow particle, respectively, $V_r = \sqrt{(U_p - U_a)^2 + (V_p - V_a)^2}$ is the movement relative velocity of the snow particles in the flow field, F_B and F_D are the buoyancy and traction forces of snow particles, respectively, 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,

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

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

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

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, respectively.

2.2.3 Basic equations of suspended particles

The movement of suspended particles is described by the following vertical diffusion equation according to horizontal uniformity condition (Déry and Yau, 1999),

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

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

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

where β is the proportionality constant, w' is the vertical turbulent fluid velocity, and we set $\beta = 1$, and $\overline{w'^2} = u_*^2$.

2.2.4 Aerodynamic entrainment

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

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

where N_a is the number of snow particles taking off due to aerodynamic entrainment, ς is a non-dimensional coefficient, approximately equal to 1×10^{-3}, u_* is the friction velocity, and u_{*t} is the threshold friction velocity.

2.3 Sublimation formula

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

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

where RH is the relative air humidity, T_a is air temperature, L_s is the latent heat of sublimation (equal to 2.84×10^6 J kg^{-1}), K_a is the air thermal conductivity, 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, respectively (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 equations

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

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

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

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

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

where K_T and K_V are the molecular diffusion coefficients of heat and water vapor, respectively, and 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 and its initial value is

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

where $p_0 = 1000 hpa$, $R_d = 287 JKg^{-1}K^{-1}$ 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 $3 \times 10^{-5} m$ at snow bed (Nemoto and Nishimura, 2001), and $R_S = 1.9974 \times 10^{-2}$.

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 saltating particles, 10^{-2} s for suspended particles, and 10^{-3} s for wind, and the calculation time is 1500 s. The motion of saltating particles is only calculated for 10 s in consideration of the practical simplicity, since saltating particles will stabilize within a few seconds. The data of saltating 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 size distribution of snow particles used in this paper fits the results of Schmidt's (1982) field observations (Fig. 1).

2.6 Calculation process

The calculation process of our model is as follow,

(1) We set a logarithmic wind field as the initial wind field, and give the first take-off particle with random particle size and vertical velocity $\sqrt{2GD}$.

(2) All the snow particles in the air are divided into saltating particles and suspended particles by Eq. 2-3. The movement of saltating particles is calculated by Eq. 4-7 and the movement of suspended particles is calculated by Eq. 11-12.

(3) If the snow particles fall on the bed, they will rebound and eject other particles which are on the bed. This process will be calculated by Eq. 8-9.

(4) If the bed shear stress is greater than the threshold value, particles are entrained from their random positions on the snow surface at vertical speed $\sqrt{2GD}$ and the number of aerodynamically entrained snow particles can be calculated by Eq. 13.

(5) The reaction force of the snow particles on the flow field is calculated by Eq.4-5 due to Newton's third law, and then the new flow filed is calculated by Eq.1.

(6) The air temperature and humidity are calculated by Eq. 16-19.

(7) The sublimation of snow particles is calculated by Eq. 14-15.

(8) The step (2)-(7) will be recycled until the end of the simulation.

3 Results and Discussion

In order to verify the judging criteria in Eq.2, we divided the particles into sets varied by 10 μm

(1-600 μm), 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, recorded the particle diameter $D_{99\%}$ and compared it with the threshold particle diameter D_{th} calculated by Eq.2. The results are shown in Table1.

As shown in Table 1, particles with diameter larger than the threshold diameter 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 those of the field observations (Fig.2). The red dots in Fig. 2 are 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 in the above field observation results. It is clear from Fig.2 that our simulation result is basically consistent with those observed in the field, demonstrating the reliability of our simulations. It can be seen from Fig. 2 that there are some discontinuities in our results, and the discontinuity is at a height of about 0.1m, which is approximately equal to the maximum height of the saltating particles (Fig. 10a) for snow particles near the height of 0.1m is rare. Therefore the randomness of snow particles' number and their sizes at 0.1m is relatively large, which leads to the discontinuity of snow mass concentration. This problem is more serious in case the wind speed is smaller, for the smaller the wind speed is, the fewer number of snow particles in the air (See Fig.2a). It's much improved when the wind speed is higher (see Fig.2c).

We also verify the reliability of our simulation by comparing our sublimation results with that of the field observations (Fig.3). The red lines in Fig. 3 are the observation results of Schmidt (1982) in Wyoming, U.S.A, in 1982. The black line represents the simulated results obtained at the same environmental conditions as those of Schmidt's. It can be seen that the total sublimation rates calculated using our model (black line) are approximately the same as Schmidt's results, and the sublimation rate at 0.01 m is two orders of magnitude larger than that at 0.1 m. These results demonstrate that our 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 represents the result of the sublimation rate of suspended particles 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 saltating particles near surface was neglected. It is shown from Fig. 4 that all the sublimation rates of suspended particle increase with height first, and then start to decrease, reaching peak at about 0.1 m. Our results are higher than those of Xiao et al. (2001). The sublimation rate of the four models is zero below at height 0.05 m, which is different with the result of our model and Schmidt (1982) in Fig. 3. This is because the relative humidity below height of 0.05 m is set to 100% in the above-mentioned four models, but not in our model.

Fig. 5 is the temporal evolution of the mass of saltating particles and suspended particles for various friction velocities. It is shown that the masses of saltating and suspended particles increase with time, and eventually reach steady. The mass of saltating particles is much higher than that of suspended particles at the steady state. The time for saltating particles to reach steady state is about 2 s, while that is about 300 s for suspended particles.

Fig. 6 shows the changes of temperature and humidity with height at initial state and at 1500 s. It is shown that air temperature and relative humidity are changed by sublimation of blowing snow particles, and the amplitude of these changes increase with the friction velocity. The greater wind velocity will lead to more snow particles into the air and undergoing sublimation and subsequently more dramatic changes in air temperature and relative humidity.

Fig. 7 and Fig. 8 show the temporal evolution of temperature and relative humidity at various heights. It is clear from in Fig. 7 and 8 that the amplitude changes of temperature and relative humidity decrease with height increasing and sublimation becomes weaker with height increasing while the relative humidity becomes constant of about 2 s at 0.01 m and about 300 s at 10 m, consistent with the corresponding values for suspended snow particles. This is because the main part of snow particles near surface is saltating particles, while that in upper air is mainly suspended particles (Fig. 10).

Fig. 8 also shows that the relative humidity near surface with three friction velocities does not reach saturation when the blowing snow particles saturate, indicating that the snow sublimation does not stop. Moreover, the vertical diffusion of water vapor can effectively reduce the negative feedback effect.

It can be seen from Fig. 9a that the sublimation rate of saltating particles shows a trend of first increasing then decreasing with time. Its peaks at 2s and gradually decreases and reaches a steady state at about 300 s. The negative feedback effect on saltating particles is very obvious and the time to reach a steady state is about 300 s. Because the mass of saltating particles increases with time during the first 2 s, with a greater amplitude than that of relative humidity, and the saltation sublimation rate increases with time. However, the mass of saltating particles basically stays 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, resulting in the stability of sublimation rate. The saltating particles distribute mainly near surface, where the amplitude change of relative humidity is strong, resulting in a strong negative feedback effect on saltating 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 suspended particles increases with time during the first 300 s with an amplitude larger than that of the relative humidity. So the suspended sublimation rate increases with time. Then the mass of suspended particles and relative humidity both reach their steady states, leading to the sublimation rate of suspended particles becomes constant. Since the suspended particles mainly distribute in upper air where the amplitude change of relative humidity is weak, therefore, the negative feedback effect on suspended particles is also weak.

Although the effect of negative feedback on saltating particles is stronger than that on suspended particles, the sublimation rate of saltating particles is still greater than that of suspended particles, indicating that the sublimation of saltating 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 saltating particles is much higher than that of suspended particles. It can be seen from Fig. 10a that saltating particles mainly distribute at height below 0.1 m, which is consistent with the previous experimental results (Takeuchi, 1980).

Fig. 11 shows that sublimation rates increases with friction velocity. The sublimation rates of saltating and suspended particles show a trend of decrease after increasing, reaching peak at about 0.01 m for saltating particles, and about 0.1 m for suspended particles. This is because the mass concentration and relative humidity of snow particles decrease with height, while temperature increases. However, mass concentration of saltating particles changes more strongly than that of suspended particles with height. Therefore, sublimation rate of saltating 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, consistent with 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.

The snow sublimation near surface was ignored in most previous studies (Déry et al., 1998; Xiao et al. 2000; Vionnet et al. 2014). That is, to define a wind velocity related height, below which saltating particles move, saltating particles are moved due to wind velocity below certain height. Assuming that moisture below the height is saturated, therefore the snow sublimation would not be counted in the region (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. (2014) was the same as that of Pomeroy and Male (1992)). Fig. 12 shows the actual ratio of our simulated sublimation mass below the three heights to the total. It is clear that all the sublimation masses below the three heights account for more than half of the total sublimation mass. This is because the main part of snow particles is saltating particles (Mellor, 1965), which mainly distribute in near surface region. Although sublimation near surface leads to significant changes in 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, resulting in continuous snow sublimation. Therefore, the main part of the sublimation mass is sublimation of saltating particles. Thus, it is not appropriate to neglect blowing snow sublimation near surface in previous reports methods (Déry et al., 1998; Xiao et al. 2000; Vionnet et al. 2014). 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 decrease in proportion of snow particles near surface, the proportion of the sublimation mass near surface will decrease as well.

Fig.13 shows the vertical profiles of vapor flux. It is clear that vapor flux increases rapidly in near surface region, where most of saltating particles move, and slows down greatly after reaching a certain height. Because 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. It also can be seen from Fig. 13 that vapor flux increases with friction velocity, similar to that for humidity (Fig.5) and moisture diffusion coefficient (Eq.17).

4 Conclusions

We have established a blowing snow sublimation model with consideration of 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 decreases air temperature while increases air humidity. Meanwhile, the snow sublimation is reduced by the negative feedback effect of temperature and humidity, especially at near surface region, in agreement with previous researches. However, moisture near surface is not saturated due to the vertical moisture diffusion, so snow sublimation near surface is a continuous process. The sublimation rate near surface is even larger than that in the upper air, because mass concentration of snow particles near surface is 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, at low wind speed, the mass of sublimation near surface accounts for more than half of the total sublimation mass, and could not be neglected. Most of the air vapor in blowing snow is from 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.

We will continue to develop our model in the future. Two possible improvements in the future are that: (1) extend the model to three dimensions and take into consideration of the effects of turbulence on the sublimation of both saltating and suspended particles in the atmospheric turbulent boundary layer, which will lead to a more accurate and realistic model; (2) propose a parametric model of the blowing snow sublimation, which will provide parameterized values for the mesoscale climate model of the polar ice sheet, the alpine glacier, snowy area with the high latitude and so on.

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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 snow particles at 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 of most of saltating particles distributed below at 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

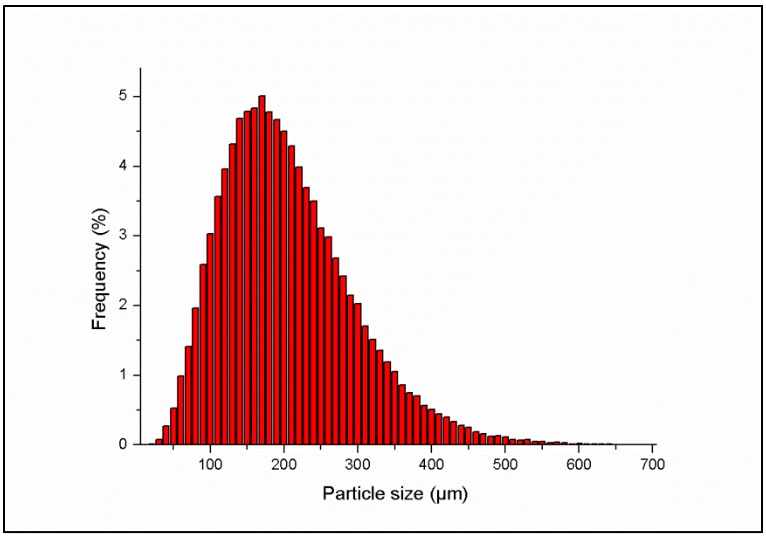

Figure 1: Particle size distribution used in this paper, which fits the results of Schmidt's (1982) field observations.

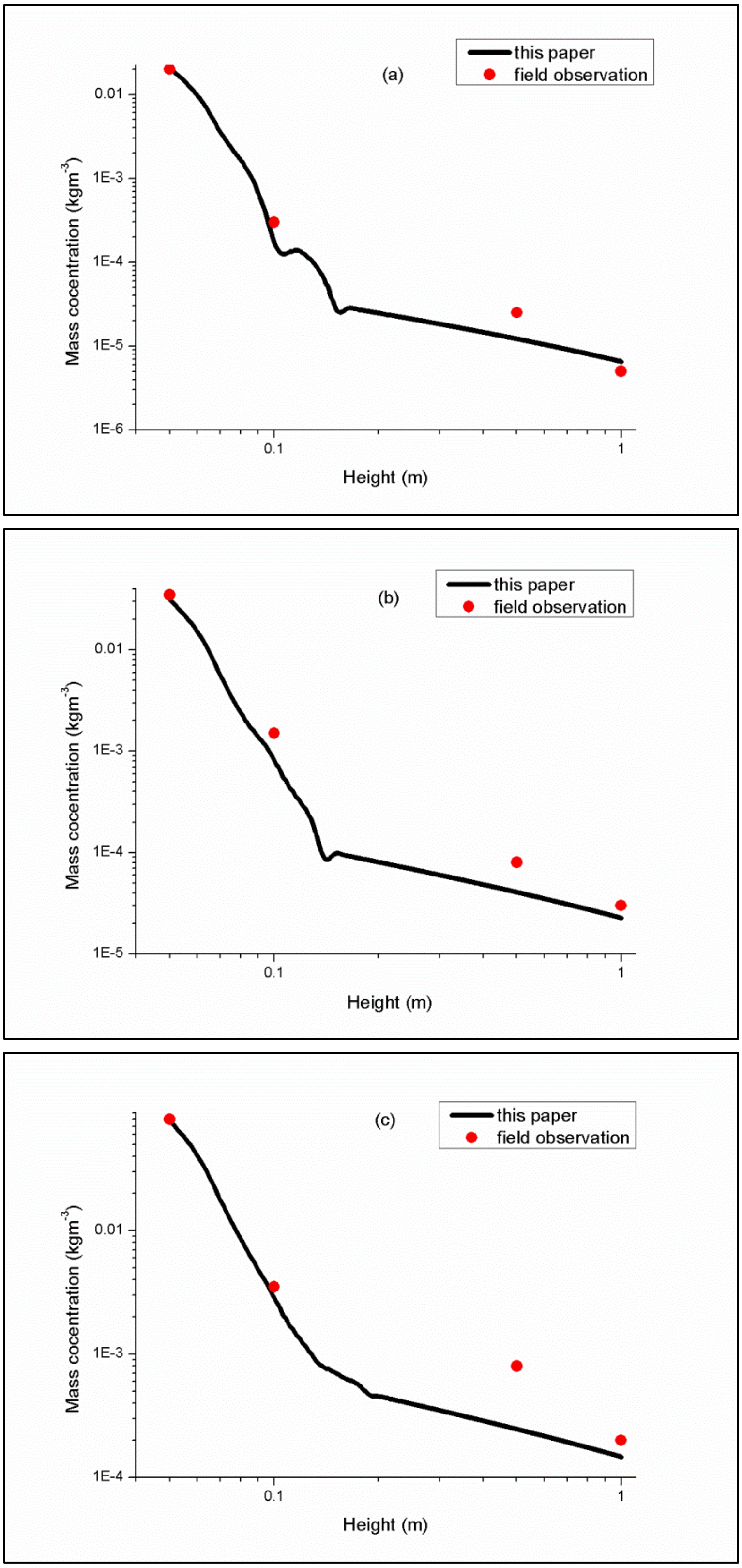

Figure 2: Comparison of mass concentration for this paper and field observation (a: $u_* = 0.35ms^{-1}; T = 268.65K$; b: $u_* = 0.41ms^{-1}; T = 268.65K$; c: $u_* = 0.54ms^{-1}; T = 268.65K$). The results of red dot are from near Saskatoon, Canada in 26 January 1987.

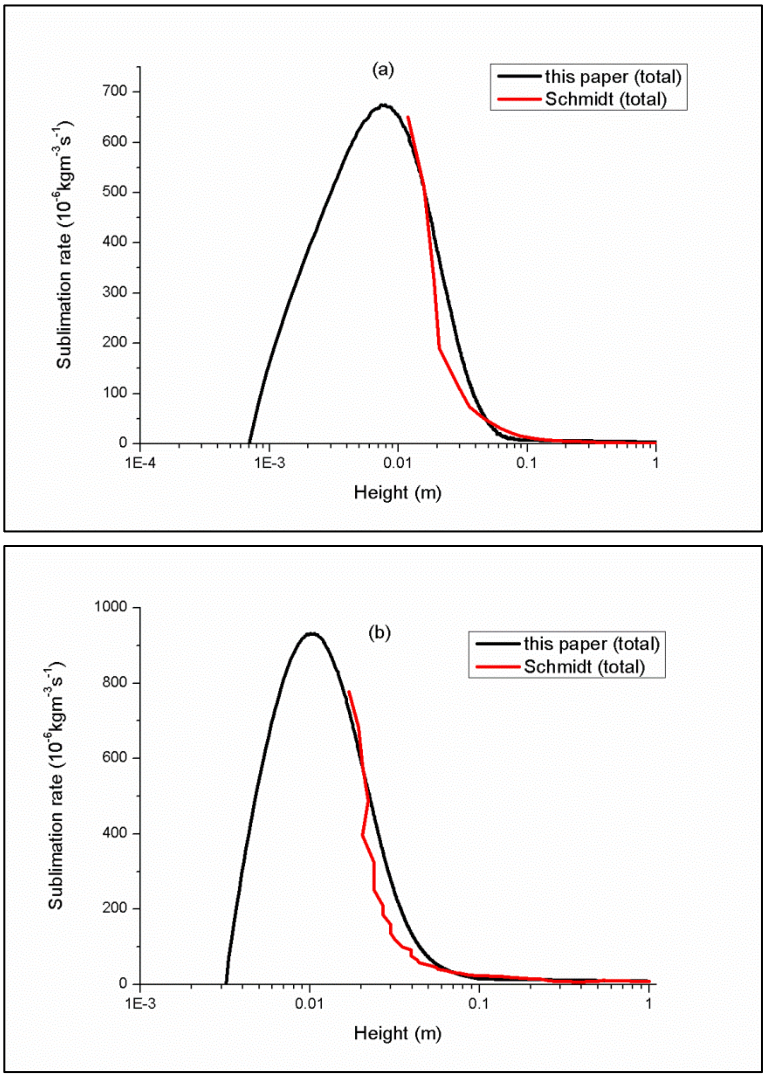

Figure 3: Comparison of sublimation rate obtained this paper and by Schmidt (1982) (a: $u_* = 0.632ms^{-1}, T = 267.45k$; b: $u_* = 1.072ms^{-1}, T = 265.65K$). The results of red line are from the data observed by Schmidt (1982) in Wyoming, U.S.A, in 1982.

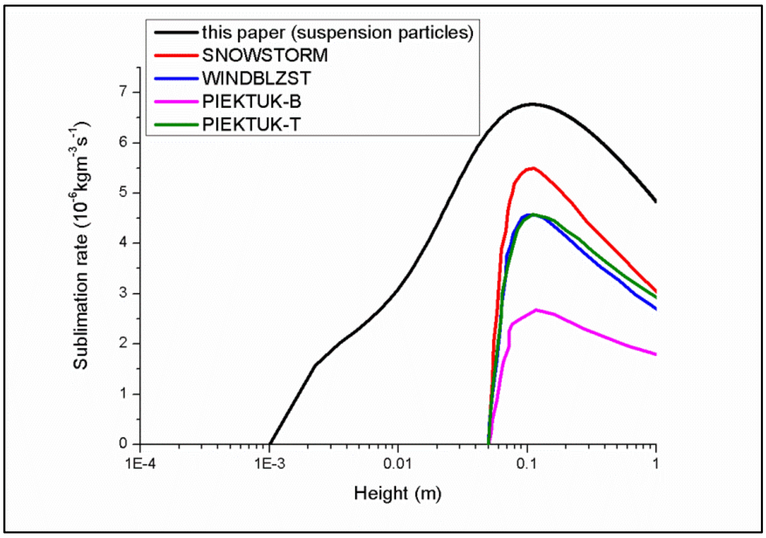

Figure 4: Comparison of sublimation rate for this paper and four blowing snow's models (Xiao et al., 2000). The friction velocity is set to 0.89m/s, and the temperature is set to 253.15K.

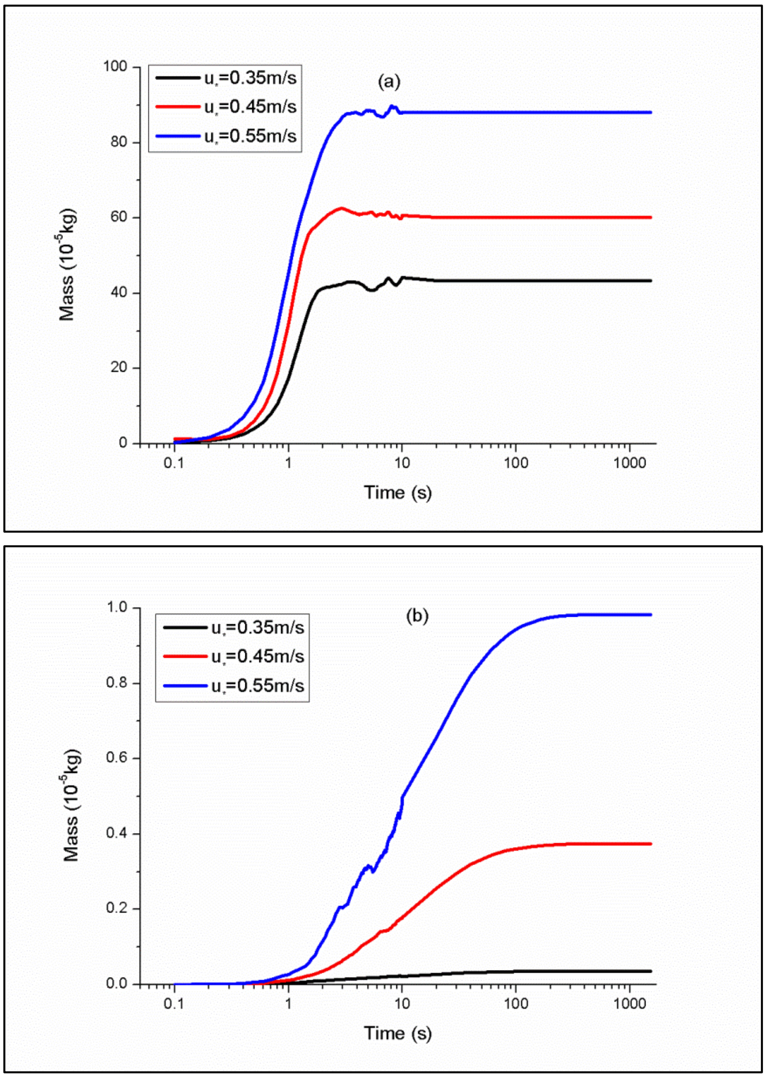

Figure 5 : Temporal evolution of mass of saltating particles and suspended particles (a: saltating particles；b: suspended particles)

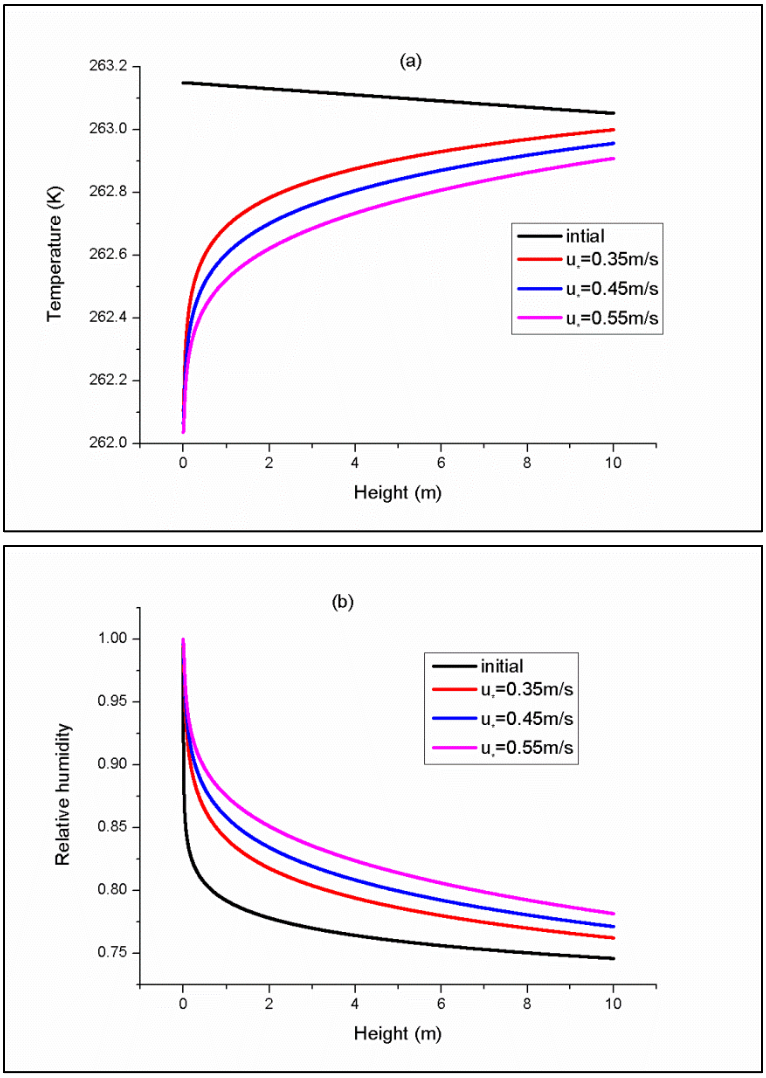

Figure 6: Vertical profiles of temperature and relative humidity

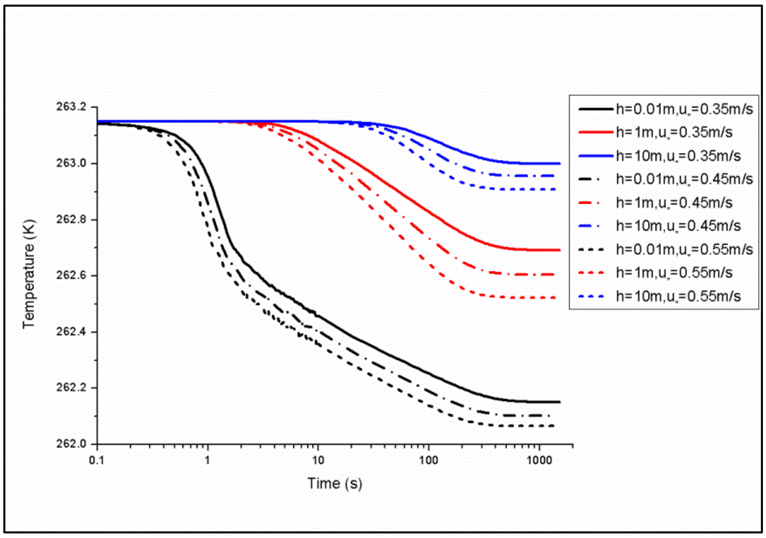

Figure 7: Temporal evolution of temperature for various heights

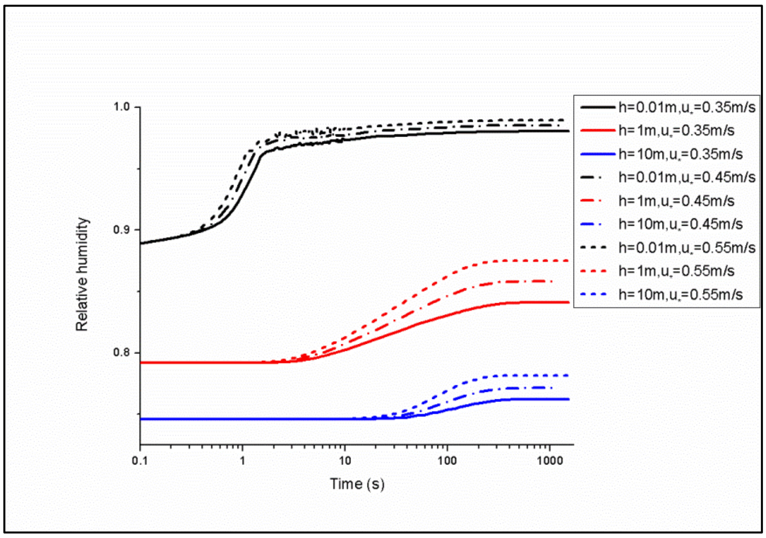

Figure 8: Temporal evolution of relative humidity for various heights

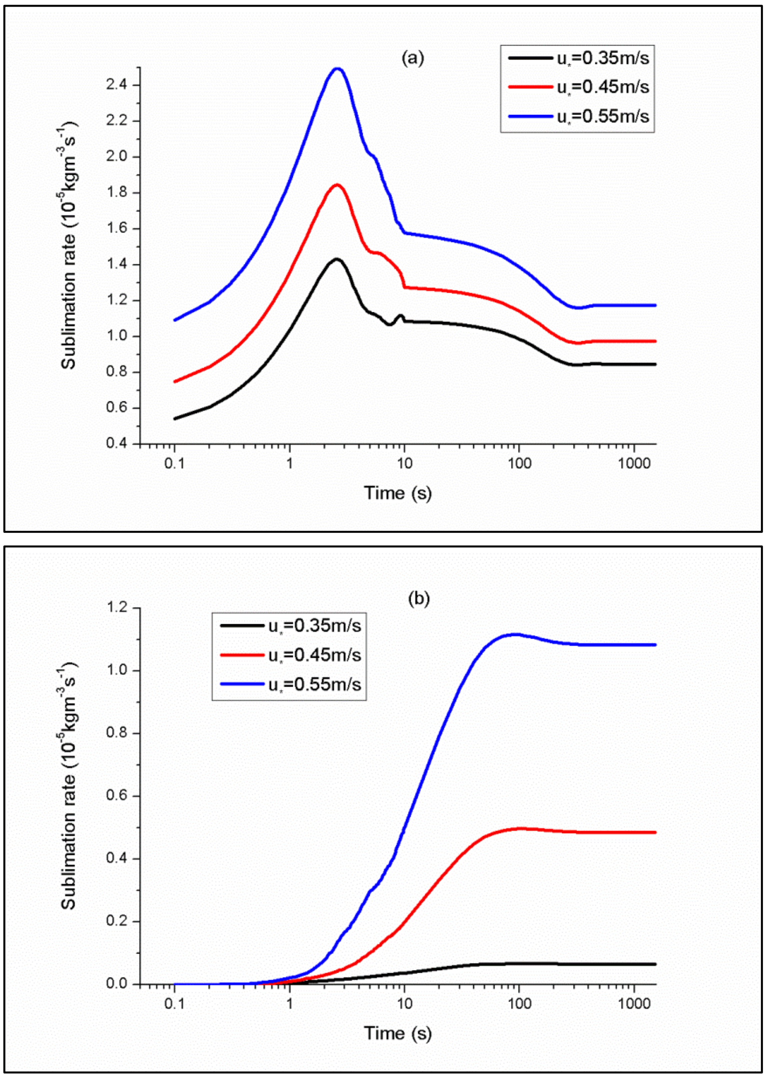

Figure 9: Temporal evolution of saltation sublimation rate and suspension sublimation rate (a: saltating particles; b: suspended particles)

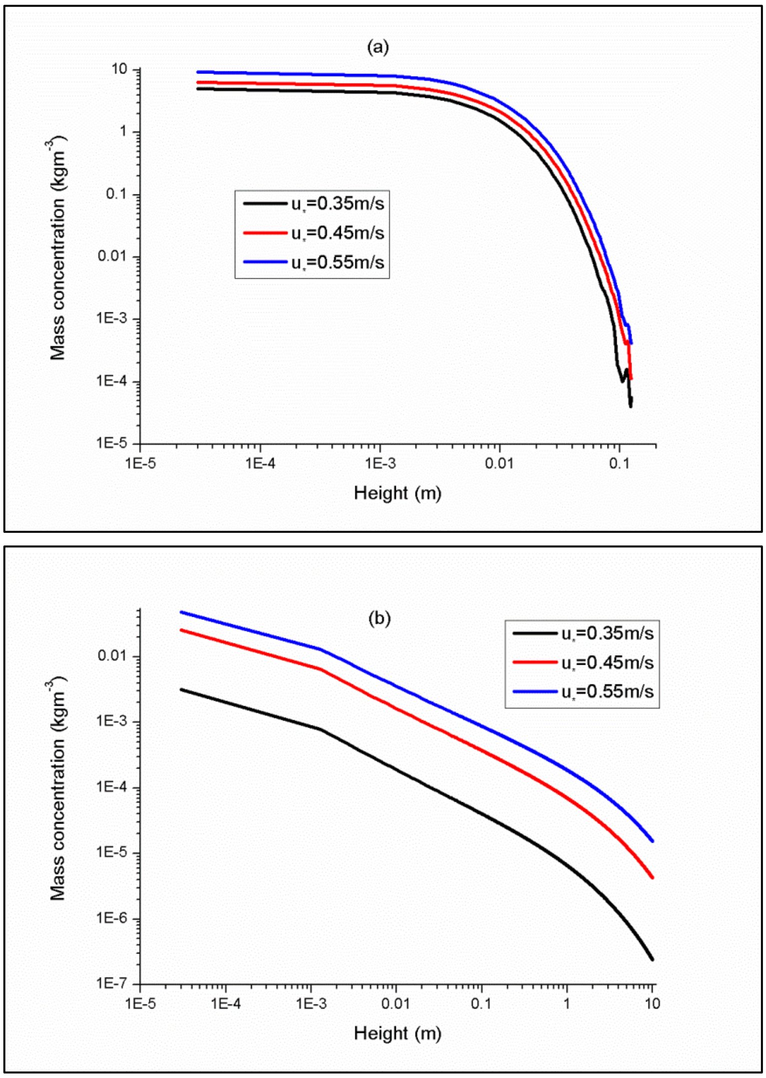

Figure 10: Vertical profiles of mass concentration for saltation and suspension (a: saltating particles, b: suspended particles)

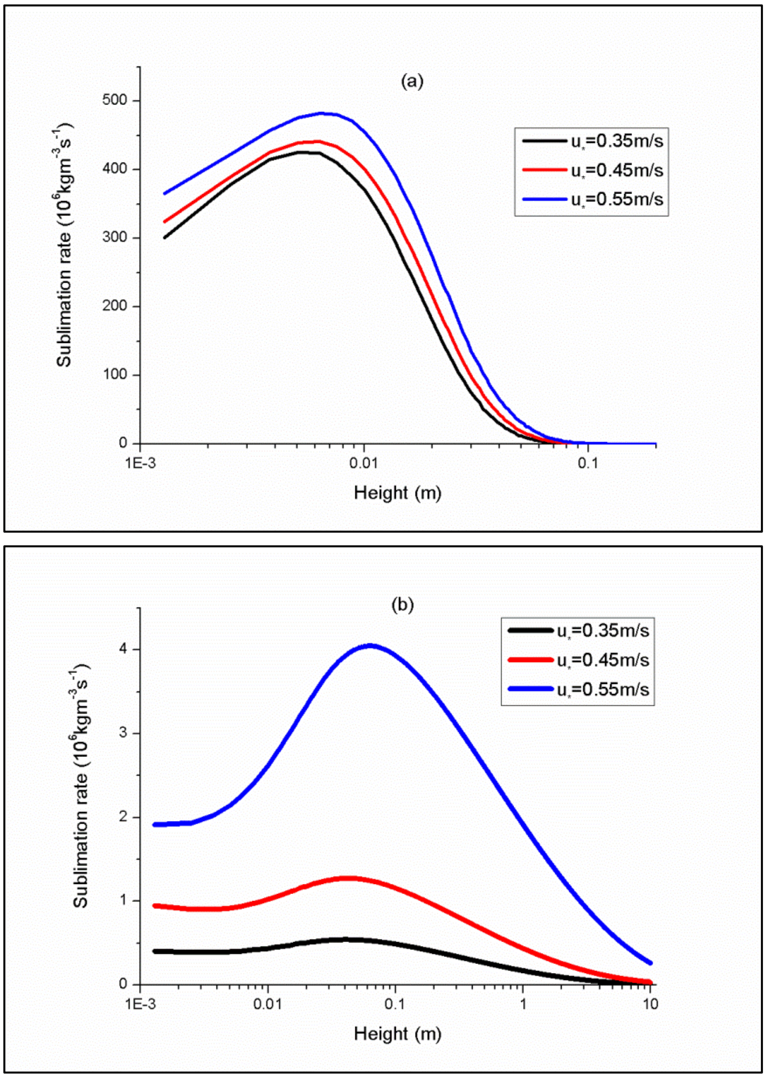

Figure 11: Vertical profiles of sublimation rate for saltation and suspension (a: saltating particles; b: suspended particles)

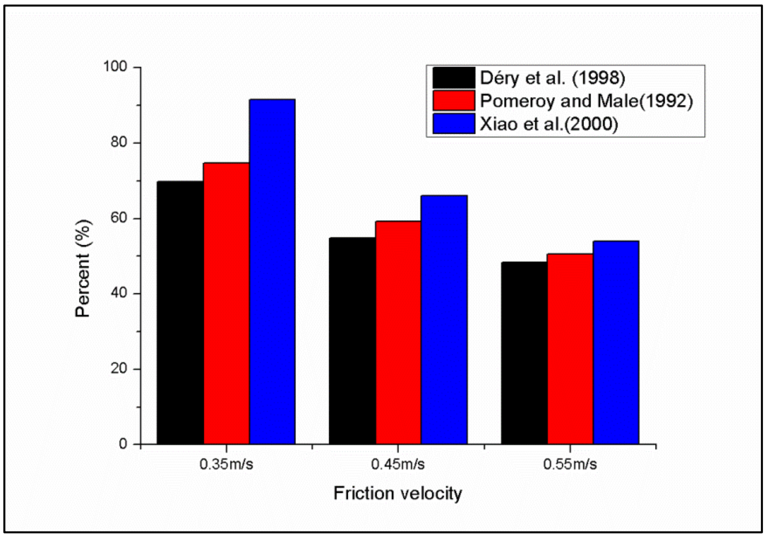

Figure 12: The ratio of sublimation mass below three heights to the total. Sublimation mass below a certain height is the sublimation mass that was ignored by other's models (Déry et al. 1998; Pomeroy and Male, 1992, and Xiao et al., 2000).

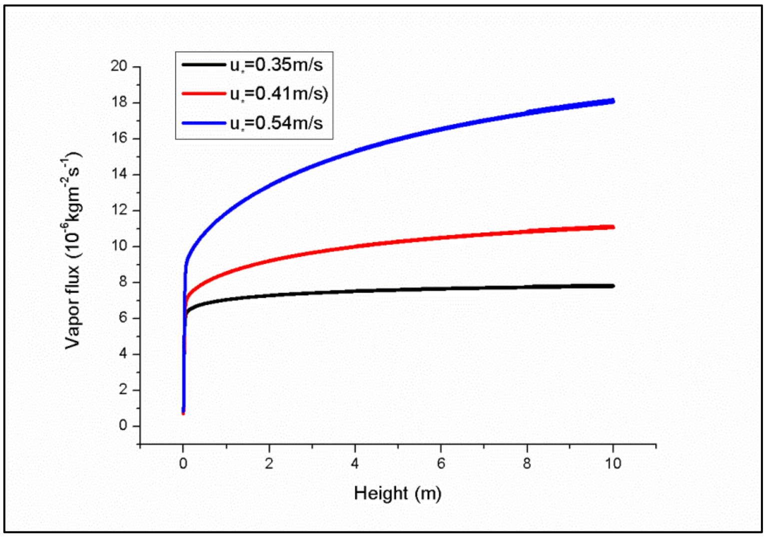

Figure 13: Vertical profiles of vapor flux