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A simulation of the large-scale drifting snow storm in a turbulent boundary layer

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Abstract. Drifting snow storm is an important aeolian process that reshapes alpine glaciers and polar ice shelves, and it may also affect the climate system and hydrological cycle since flying snow particles exchange considerable mass and energy with air flow. Prior studies have rarely considered the full-scale drifting snow storm in the turbulent boundary layer, thus, the transportation feature of snow flow higher in the air and its contribution are largely unknown. In this study, a large eddy simulation is combined with a subgrid scale velocity model to simulate the atmospheric turbulent boundary layer, and a Lagrangian particle tracking method is adopted to track the trajectories of snow particles. A drifting snow storm that is hundreds of meters in depth and exhibits obvious spatial structures is produced. The snow transport flux profile at high altitude, previously not observed, is quite different from that near the surface, thus, the extrapolated transport flux profile may largely underestimate the total transport flux. At the same time, the development of a drifting snow storm involves three typical stages, the rapid growth, the gentle growth and the equilibrium stages, in which the large-scale updrafts and subgrid scale fluctuating velocities basically dominate the first and second stage, respectively. This research provides an effective way to get an insight into natural drifting snow storms.

1 Introduction

Snow, one type of solid precipitation, is an important sources of material to mountain glaciers and polar ice sheets, which are widespread throughout high and cold regions (Chang et al., 2016; Gordon and Taylor, 2009; Lehning et al., 2008). A common natural phenomenon over snow cover is the drifting snow storm, which occurs when the wind speed exceeds a critical value (Doorschot et al., 2004; Li and Pomeroy, 1997; Sturm and Stuefer, 2013). Drifting snow can entrain loose snow particles on the bed into the air, which may be further transported to high altitude by turbulent eddies (King, 1990; Mann et al., 2000; Nemoto and Nishimura, 2004). Drifting snow clouds typically can range in thickness from tens to thousands of meters (Mahesh et al., 2003; Palm et al., 2011), which may not only affect people's daily life by reducing the visibility and producing local accumulation (Gordon and Taylor, 2009; Mohamed et al., 1998) , but also can influence the global climate system evolution by changing the mass and energy balance of ice shelves (Cess and Yagai, 1991; Hanesiak and Wang, 2005; Hinzman et al., 2005; Lenaerts and Broeke, 2012).

Several field experiments on drifting snow storm have been performed (Bintanja, 2001; Budd, 1966; Dingle and Radok, 1961; Doorschot et al., 2004; Gallée et al., 2013; Gordon and Taylor, 2009; Guyomarch et al., 2014; Kobayashi, 1978; Mann et al., 2000; Nishimura and Nemoto, 2005; Nishimura et al., 2015; Pomeroy and Gray, 1990; Sbuhei, 1985; Schmidt, 1982; Sturm and Stuefer, 2013) since the middle of the last century. However, the measurements are commonly conducted near the surface, thus, the drifting snow features at high altitude are unknown, and the impacts of these

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features are difficult to assess. A thorough investigation documenting the evolution process and structure of a full-scale drifting snow storm is essential to understand this natural phenomenon and assess its impacts.

Drifting snow models, on the other hand, offer a panoramic view of the evolution process of drifting snow and thus have become one of the most useful research approaches. Many continuum medium models of drifting snow (Bintanja, 2000; Déry and Yau, 1999; Schneiderbauer and Prokop, 2011; Uematsu et al., 1991; Vionnet et al., 2013) have advanced the knowledge of natural drifting snow to a great extent. However, a particle-tracking drifting snow model is still needed since the particle characteristics and its motion require further investigation. Although a series of particle tracking models (Huang et al., 2016; Huang and Shi, 2017; Huang and Wang, 2015; 2016; Nemoto and Nishimura, 2004; Zhang and Huang, 2008; Zwaaftink et al., 2014) have been established, these models have generally focused on the grain-bed interactions and particle motions near the surface. Thus, a drifting snow model aimed at producing a large-scale drifting snow storm in a turbulent boundary layer deserves further exploration.

In this study, a drifting snow model in the atmospheric boundary layer that focuses on the full-scale drifting snow storm is established. The wind field is solved using a large eddy simulation for the purpose of generating a turbulent atmospheric boundary layer. A subgrid scale (SGS) velocity is also considered to include the diffusive effect of small scale turbulence. Finally, particle motion is calculated using a Lagrangian particle tracking method. The large-scale drifting snow storm is produced and its

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spatial structures and transport features are analyzed.

2 Model and methods

2.1 Simulation of a turbulent atmospheric boundary layer

The mesoscale atmosphere prediction pattern ARPS (Advanced Regional Prediction System, version 5.3.3) is adopted to simulate the turbulent atmospheric boundary layer, in which the filtered three-dimensional compressible non-hydrostatic Naiver-Stokes equation is solved (Xue et al., 2001):

$$\frac{\partial \rho}{\partial t}+\frac{\partial}{\partial x_i}(\rho \tilde{u}_i)=0 \tag{1}$$

$$\frac{\partial \rho \tilde{u}_i}{\partial t}+\frac{\partial \rho u_i \tilde{u}_j}{\partial x_j}=-\frac{\partial \tilde{p}^*}{\partial x_i}+B\delta_{i3}-\frac{\partial \tau_{ij}}{\partial x_j} \tag{2}$$

where '~' represents variables that are filtered and the filtering scale is $\tilde{\Delta}=\left(\Delta x_1 \Delta x_2 \Delta x_3\right)^{1/3}$, in which Δx_i is the grid spacing along streamwise ($i=1$), spanwise ($i=2$) and vertical direction ($i=3$), respectively. u_i is the instantaneous wind speed component, and x_i is the position coordinate. t is time, δ_{ij} is the Kronecker delta, $B=-g\,\rho'/\rho$ is the buoyancy caused by the air density perturbation ρ', and g is the acceleration due to gravity. $p^*=p'-\alpha\nabla(\rho\mathbf{u})$ contains the pressure perturbation term and damping term, where α is the damping coefficient and ∇ is the divergence. The subgrid stress τ_{ij} can be expressed as (Smagorinsky, 1963):

$$\tau_{ij}=-2\nu_t \tilde{S}_{ij}=-2\left(C_s \tilde{\Delta}\right)^2\left|\tilde{S}\right|\tilde{S}_{ij} \tag{3}$$

where $\tilde{S}_{ij}=0.5\left(\partial \tilde{u}_i/\partial x_j+\partial \tilde{u}_j/\partial x_i\right)$ is the strain rate tensor and $\left|\tilde{S}\right|=\sqrt{2\tilde{S}_{ij}\tilde{S}_{ij}}$, C_s is Smagorinsky coefficient that is determined locally by the dynamic Lagrangian model (Meneveau et al., 1996).

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Considering the large grid spacing in simulating an atmospheric boundary layer (where the information about turbulent vortices smaller than the grid size is missing), the SGS velocity is also included. Namely, the local wind velocity $\tilde{u}_i\left(\vec{x}(t)\right)$ is composed of a resolved Eulerian large-scale part $\tilde{u}_i\left(\vec{x}(t)\right)$ (obtained from the linear weighting of surrounding grid points) and a fluctuating SGS contribution $u_i'(t)$. The SGS velocity can be calculated from the SGS stochastic model of Vinkovic et al. (2006):

$$du_i' = \left(-\frac{1}{T_L}+\frac{1}{2\tilde{k}}\frac{d\tilde{k}}{dt}\right)u_i'dt + \sqrt{\frac{4\tilde{k}}{3T_L}}\,d\eta_i\left(t\right) \tag{4}$$

where $T_L = 4\tilde{k}/\left(3C_0\tilde{\varepsilon}\right)$ is the Lagrangian correlation time scale. Here, C_0 is the Lagrangian constant, $\tilde{\varepsilon} = C_\varepsilon \tilde{k}^{3/2}/\tilde{\Delta}$ is the subgrid turbulence dissipation rate, C_ε is a constant, and $d\eta_i$ is the increment of a vector-valued Wiener process with zero mean and variance dt. $\tilde{k}$ is the subgrid turbulent kinetic energy and can be obtained from the transport equation (Deardorff, 1980):

$$\frac{\partial \tilde{k}}{\partial t} + \tilde{u}_j\frac{\partial \tilde{k}}{\partial x_j} = \frac{\nu_t}{3}\frac{g}{\theta_0}\frac{\partial \tilde{\theta}}{\partial x_3} + 2\nu_t\tilde{S}_{ij}^2 + 2\frac{\partial}{\partial x_j}\left(\nu_t\frac{\partial \tilde{k}}{\partial x_j}\right) + \tilde{\varepsilon} \tag{5}$$

where θ is the potential temperature and θ_0 is the surface potential temperature.

2.2 Governing equation of particle motion

The trajectory of each snow particle is calculated using a Lagrangian particle tracking method. Since a snow particle has is almost 10^3 times more dense than air, airborne particles are assumed to process only gravity and fluid drag forces, and the governing equations of particle motion can be expressed as (Dupont et al., 2013; Huang and Wang, 2016; Vinkovic et al., 2006):

$$\frac{dx_{pi}}{dt} = u_{pi} \tag{6}$$

$$\frac{du_{pi}}{dt} = m_p \frac{V_{ri}}{T_p} f(Re_p) + \delta_{i3} g \tag{7}$$

where x_{pi} and u_{pi} are the position coordinate and velocity of the snow particle, respectively. m_p is the mass of the solid particle, V_r is the relative speed between the snow particle and air, and $T_p = \rho_p d_p^2 / 18\rho\nu$ is the particle relaxation time, where ρ_p is the particle density, d_p is the particle diameter and $\nu = 1.5e-5$ is the dynamic viscosity of air. $f\left(Re_p\right)$ can be expressed as (Clift et al., 1978):

$$f(Re_p) = \begin{cases} 1 & (Re_p < 1) \\ 1 + 0.15\,\mathrm{Re}_p^{0.687} & (Re_p \geq 1) \end{cases} \tag{8}$$

where $Re_p = V_r d / \nu$ is the particle Reynolds number.

2.3 Initial conditions of snow particles

To generate a large-scale drifting snow storm, a steady-state snow saltation condition is set as the bottom boundary condition for particles. During drifting snow events, the sum of residual fluid shear stress τ_f and particle-borne shear stress τ_p should be equal to the total fluid shear stress τ, thus, the particle-borne stress can be expressed as:

$$\tau_p = \tau - \tau_f \tag{9}$$

Here, the residual fluid shear stress τ_f is set to be the threshold shear stress τ_{tf} of drifting snow, which can be read as (Clifton et al., 2006):

$$\tau_{tf} = A^2 g d \left(\rho_p - \rho\right) \tag{10}$$

in which $A = 0.2$ is a constant, g is the gravity acceleration and d is the mean diameter of the snow particles.

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At the same time, the particle-borne shear stress at the surface can be calculated from the particle trajectories as (Nemoto and Nishimura, 2004):

$$\tau_p = \sum_{i=1}^{n_\downarrow} m_i u_{pi\downarrow} - \sum_{i=1}^{n_\uparrow} m_i u_{pi\uparrow} \tag{11}$$

where m_i is the mass of particle and $u_{pi\downarrow}$ and $u_{pi\uparrow}$ are the horizontal speeds of impact and lift-off particles, respectively. $n_\downarrow$ and $n_\uparrow$ are the particle number per unit area in unit time of impact and lift-off grains, respectively, which should be equivalent in steady-state saltation. Thus, the number of lift-off particles per unit area is:

$$n_\uparrow = n_\downarrow = \frac{\tau_p}{\langle m_i \rangle \left(1 - \langle e_h \rangle\right) \langle u_{pi\downarrow} \rangle} \tag{12}$$

in which $\langle\ \rangle$ indicates the overall average, and e_h is the horizontal restitution coefficient of snow particle. According to Sugiura and Maeno (2000), the mean horizontal restitution coefficient can be expressed as:

$$\langle e_h \rangle = \begin{cases} 0.48\theta_i^{0.01} & v_i \le 1.27ms^{-1} \\ 0.48\left(\dfrac{v_i}{1.27}\right)^{-\log\left(\frac{v_i}{1.27}\right)} \theta_i^{0.01} & v_i > 1.27ms^{-1} \end{cases} \tag{13}$$

where θ_i and v_i are the impact velocity and angle, respectively. Here, θ_i has a mean value of approximately 10° (Sugiura and Maeno, 2000), and $\langle v_i \rangle$ is set to be the threshold of impact velocity, which is determined by setting ejection number $n_e = 0.51 v_i^{0.6} \theta_i^{0.16}$ equal to 1. In this way, the mean horizontal velocity of impact particles can be obtained through $\langle u_{pi\downarrow} \rangle = \langle v_i \rangle \cos\langle \theta_i \rangle$.

Then, the velocities of lift-off particles can be obtained from the restitution coefficient of snow. The horizontal restitution coefficient obeys the normal distribution with a mean value given in Eq. 13, and a standard variance as follow (Sugiura and Maeno, 2000):

$$\sigma^2 = \begin{cases} 0.07\theta_i^{-0.06} & v_i \le 0.52\, ms^{-1} \\ 0.07(\frac{v_i}{0.52})^{-\log(\frac{v_i}{0.52})}\theta_i^{-0.06} & v_i > 0.52\, ms^{-1} \end{cases} \quad (14)$$

On the other hand, the vertical restitution coefficient can be described by a two parameter gamma function (see Eq. 17), in which the parameter α and β can be expressed as (Sugiura and Maeno, 2000):

$$\alpha = \begin{cases} 1.22\theta_i^{0.47} & v_i \ge 0.84\, ms^{-1} \\ 1.22(\frac{v_i}{0.84})^{\log(\frac{v_i}{0.84})}\theta_i^{0.47} & 0.84 < v_i \le 1.23\, ms^{-1} \\ 1.22(\frac{v_i}{0.84})^{\log(\frac{v_i}{0.84})}(\frac{v_i}{1.23})^{-2\log(\frac{v_i}{1.23})}\theta_i^{0.47} & v_i \ge 1.23\, ms^{-1} \end{cases} \quad (15)$$

$$\beta = \begin{cases} 12.85\theta_i^{-1.41} & v_i \ge 0.84\, ms^{-1} \\ 12.85(\frac{v_i}{0.84})^{-\log(\frac{v_i}{0.84})}\theta_i^{-1.41} & 0.84 < v_i \le 1.23\, ms^{-1} \\ 12.85(\frac{v_i}{0.84})^{-\log(\frac{v_i}{0.84})}(\frac{v_i}{1.23})^{\log(\frac{v_i}{1.23})}\theta_i^{-1.41} & v_i \ge 1.23\, ms^{-1} \end{cases} \quad (16)$$

2.4 Simulation details

The computational domain is 1000×500×1000 m, with a uniform horizontal grid size of 5 m adopted to solve finer vortex structure in the atmospheric boundary layer. The mean grid size in the vertical direction is 20 m, with a grid refinement algorithm adopted near the surface (the finest grid size is 1 m). Periodic boundaries are used along streamwise and spanwise dimensions, and the bottom is set as a grid wall. The top is set as an open radiation boundary with a Rayleigh damping layer that is 250 m in depth.

The atmosphere is neutral with an initial potential temperature of 300K, and an initial relative humidity of 90%. The initial wind profile is logarithmic with a surface roughness of 0.1m (Doorschot et al., 2004). Atmospheric turbulence is induced by random initial potential temperature perturbations at the first-level grid level with a maximum magnitude of 0.5K, and is sustained by a constant heat flux at the bottom. The constant heat flux is 50 Wm^{-2} according to the observation of Pomeroy and Essery (1999).

For particles, periodic boundary conditions are also used at lateral boundaries, and a rebound boundary condition without energy loss is adopted at the model top. The bottom boundary condition for particles is given in Sect. 2.3, and is updated every 0.5 s. Additionally, each particle represents one particle parcel for the purpose of reducing computational complexity.

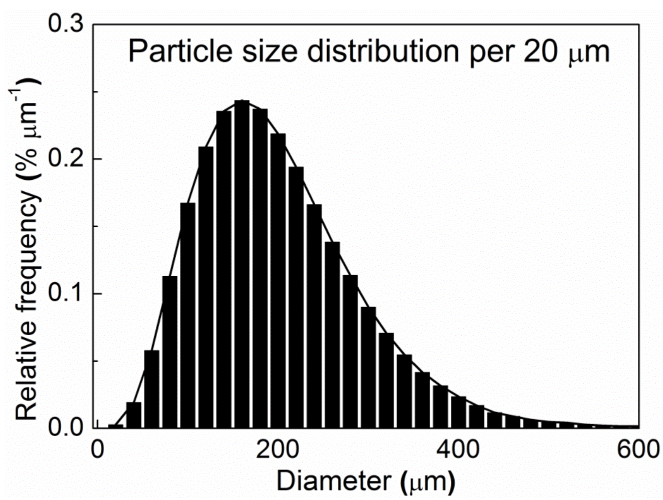

Figure 1. Size distribution of lift-off snow particles in this simulation.

The size distribution of lift-off particles in drifting snow can be well described by the two-parameter gamma function (Budd, 1966; Gordon and Taylor, 2009;

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Nishimura and Nemoto, 2005; Schmidt, 1982):

$$f\left(d\right)=\frac{d^{\alpha-1}}{\beta^{\alpha}\Gamma\left(\alpha\right)}\exp\left(-\frac{\beta}{d}\right) \tag{17}$$

where d is the particle diameter, and α and β are the shape and scale parameter of the distribution, respectively. In this simulation, the diameters of lift-off snow particles are given randomly from a gamma function with the parameters of $\alpha=4$ and $\beta=50$, as shown in Fig. 1, which is also consistent with observed particle size distributions (Nishimura and Nemoto, 2005; Schmidt, 1982).

3 Results and discussions

3.1 Model validation

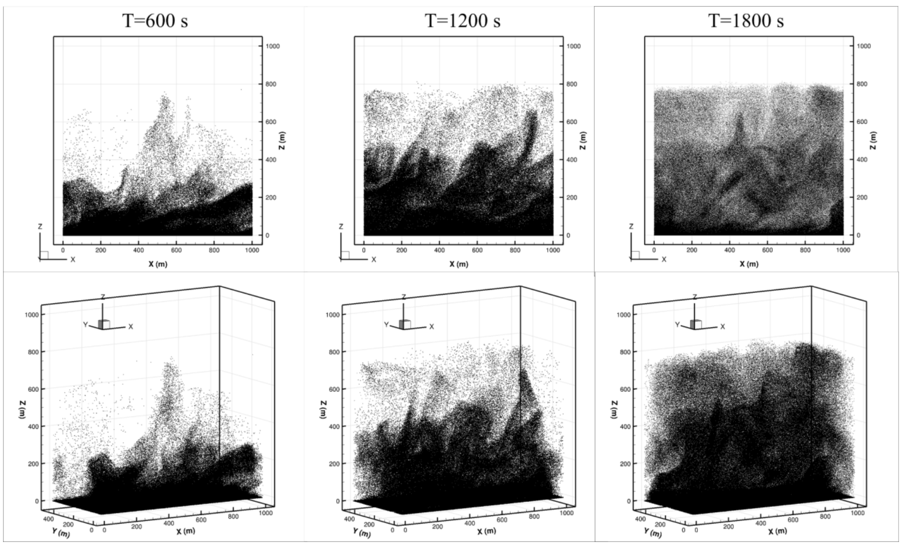

Figure 2. Drifting snow storm at different moments under the friction velocity of 0.29 ms^{-1}.

When drifting snow occurs in the atmospheric boundary layer, updrafts and turbulence fluctuations can send snow particles to high altitude, forming a fully developed drifting snow storm. Fig. 2 shows the drifting snow storm in the

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atmospheric boundary layer at different moments, in which the friction velocity is $u_* = 0.29\ ms^{-1}$ and dark spots represent snow particles. It can be seen that drifting snow storm experiences an evolution process from near the surface to high altitudes, which induces the fact that particle concentration decreases along increasing height. The high concentrations of drifting snow cloud are generally below 500 m, though snow particles may reach up to approximately 800 m under this condition. This is also consistent with observations (Mahesh et al., 2003; Palm et al., 2011).

Since a drifting snow storm exhibits a different structure from bottom to top, the evolution of particle number density profile in the drifting snow storm is shown in Fig. 3, which is also compared with measurements of Mann et al. (2000) . From this figure, the thickness of the drifting snow layer obviously increases with time, and almost approaches its steady state after 1200 s. At the same time, the particle number density basically decreases with height, which is consistent with the measurements of Mann et al. (2000) at various friction velocities. The predicted particle number density at the surface is much larger than at higher altitude and observations, mainly because the saltating particles are also included.

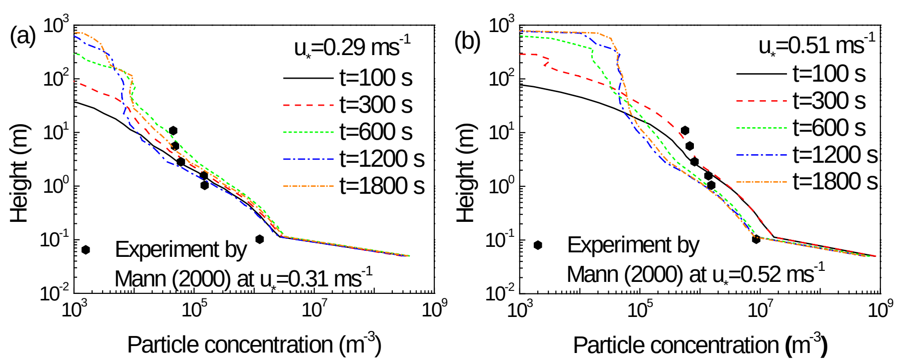

Figure 3. Evolution of particle number density under various friction velocities (a)

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0.29 ms^{-1} and (b) 0.51 ms^{-1}.

Generally, smaller particles are more likely to be transported higher in the air. Fig. 4 shows the variation of modeled average particle diameter versus height, which is also compared with various field measurements (Nishimura and Nemoto, 2005; Schmidt, 1982). Similar to the field observations, the average particle size basically decreases with height at lower altitude but is almost constant above 1 m. The average particle diameter is approximately 75 μm ranging from one meter to hundreds of meters in height, which is also consistent with the measurements of K Nishimura and Nemoto (2005).

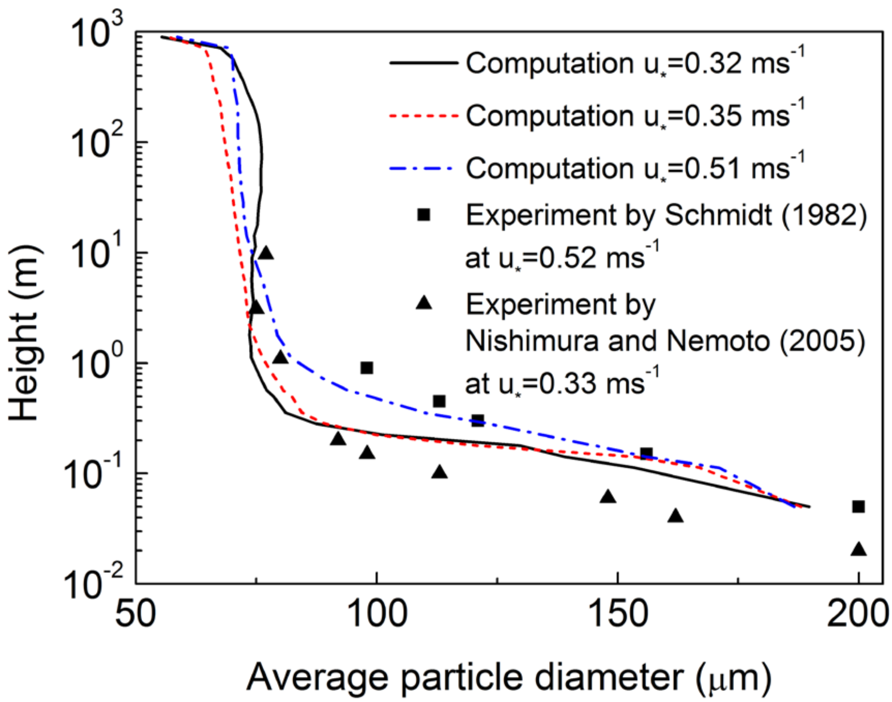

Figure 4. Variation of average particle diameter versus height.

Then, the particle size distributions at various heights are also compared with experiment results. As shown in Fig. 5, the heights are 0.05 m, 0.5 m and 1 m. The modeled particle size distributions at various heights are consistent with the measurements (Nishimura and Nemoto, 2005; Schmidt, 1982). Therefore, the established model is able to produce a large-scale drifting snow storm.

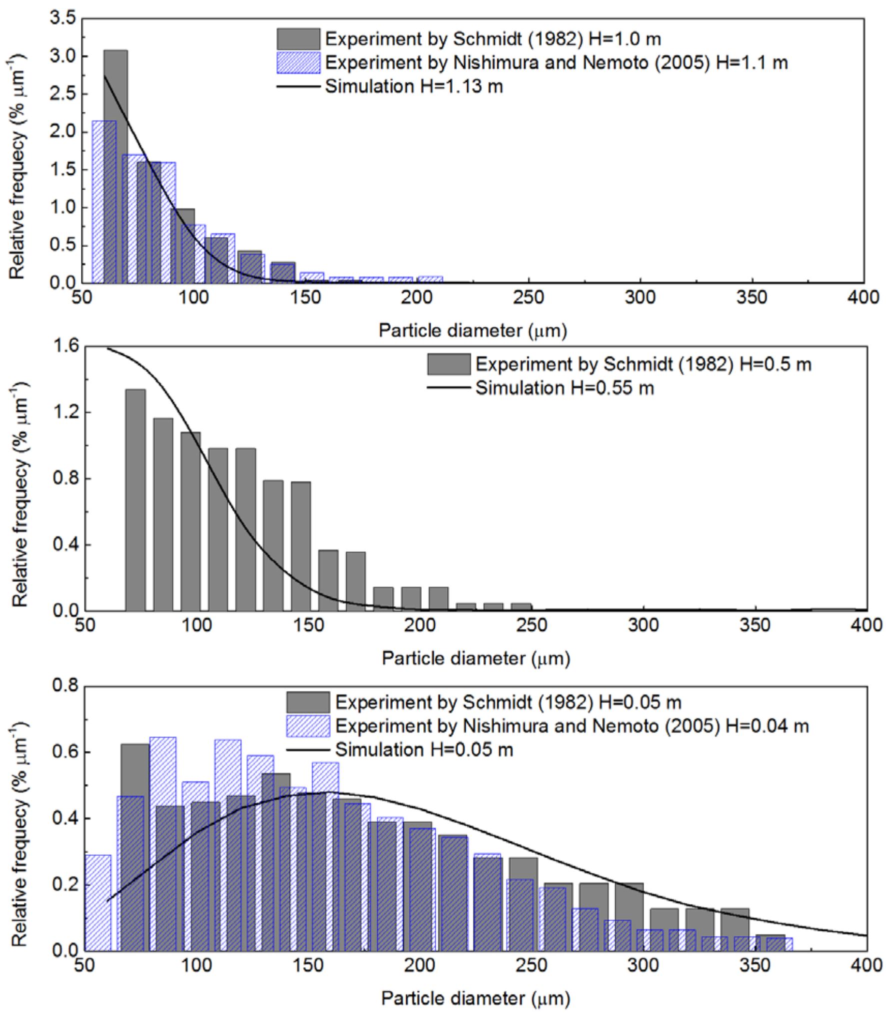

Figure 5. Particle size distribution at various heights.

3.2 Snow transport flux

The snow transport flux is of great importance to predict the mass and energy balances of ice sheets. The total transport flux can be obtained from vertical

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integration of the snow transport flux profile.

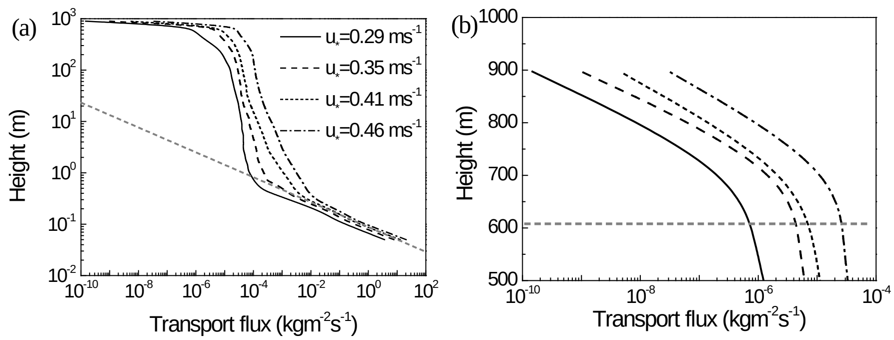

Figure 6. Variations of snow transport flux versus height.

The profiles of snow transport rate, per unit area, per unit time, under various friction velocities are shown in Fig. 6(a). It can be seen that the transport flux undergoes a sharp decrease with height at lower altitude (e.g., below 1.0 m), however, the transport flux tends to decrease rather gentle until almost the top of the drifting snow storm, as shown in Fig. 6(b), probably due to the large-scale turbulent motion and increasing wind speed with height. In other words, the suspension flux of drifting snow at higher altitudes, previously not observed, may be much larger than we previously thought.

In previous studies, the transport flux profile is commonly described using an exponential decay form based on the extrapolation from measurements near the surface (Mann et al., 2000; Nishimura and Nemoto, 2005; Schmidt, 1982; 1984; Tabler, 1990), which may result in a considerable underestimate of the total transport flux. The proportions of suspension flux above a given height h_c (referred as Q_c) to the total suspension flux Q_s are shown in Fig. 7, in which snow particles below 0.1 m are not calculated (Mann et al., 2000).

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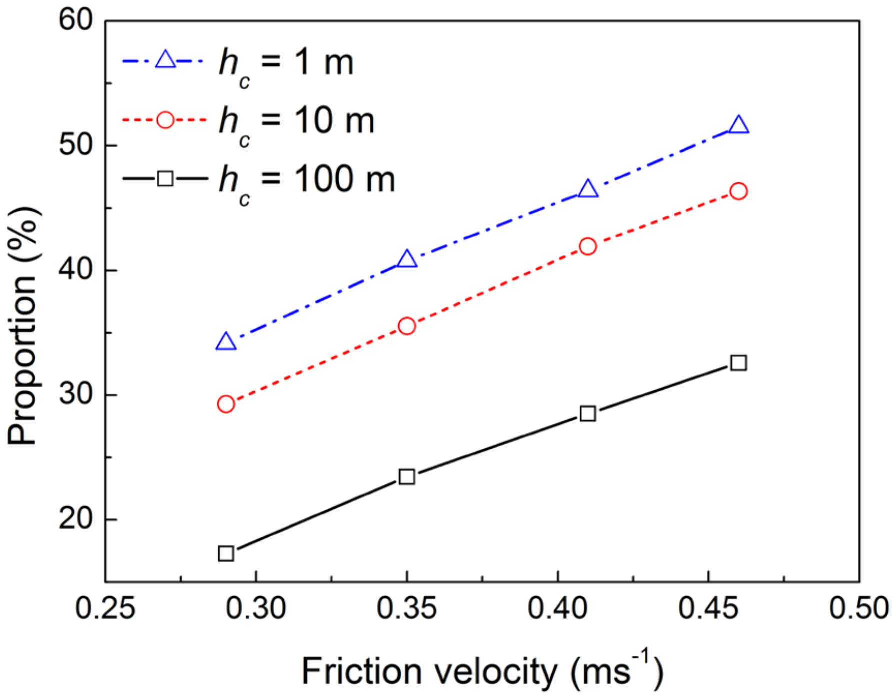

Figure 7. Proportion of suspension flux above h_c to the total suspension flux under various friction velocities.

From Fig. 7, the contribution of Q_c to the total suspension flux is non-negligible under various h_c, the proportion of Q_c when h_c =100 m to the total suspension flux has exceeded 30% when the friction velocity is 0.46 ms^{-1}. At the same time, the proportion of Q_c to the total suspension flux increases with friction velocity but decreases with increasing h_c.

In this way, not only the snow transport flux, but also the sublimation of suspended snow particles should be reevaluated because the sublimation rate of snow particles higher in the air may be much larger than near the surface due to the lower air humidity and greater wind speed at higher altitude (Mann et al., 2000; Nishimura and Nemoto, 2005; Schmidt, 1982; 1984; Tabler, 1990).

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3.3 Structures in a drifting snow storm

In a drifting snow storm, particles aggregate locally and produce special spatial structures (as shown in Fig. 2). These structures should be directly related to the turbulence structures present in the atmospheric boundary layer. Drifting snow storms without atmospheric turbulence are shown in Fig. 8. Compared with Fig. 2, drifting snow particles mainly travel at the near surface with a uniform spatial distribution when atmospheric turbulence is not included.

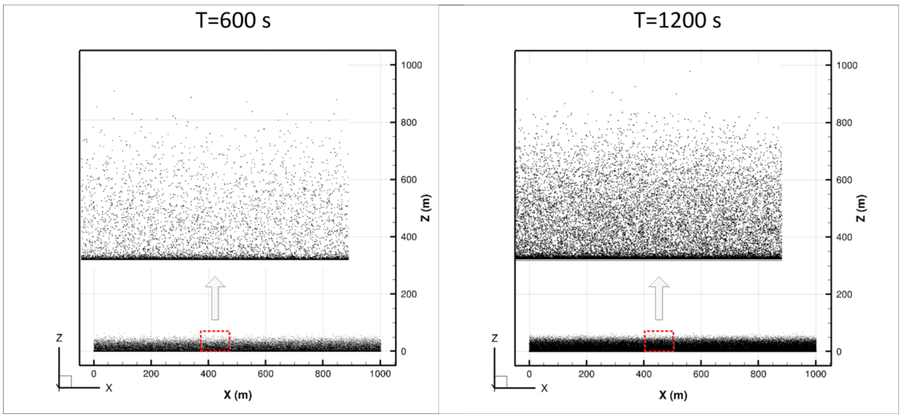

Figure 8. Drifting snow storm without atmospheric turbulence under friction velocity of 0.35 ms^{-1}.

It is known that snow particles will become suspended if the local vertical wind speed exceeds the terminal velocity of particle. In a turbulent atmospheric boundary layer, there exists a large amount of turbulent structures with different scales and shapes. The vertical wind speed component of large-scale turbulence (namely, updraft) plays an important role in carrying snow particles to high altitude, while small scale turbulence (e.g., the SGS fluctuating velocity) tends to spread particles from high concentration zones to low concentration zones. As shown in Fig. 9(a), at the initial

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period of a drifting snow storm, the structures in the drifting snow storm are consistent with large-scale updrafts, and snow particles are mainly located in the updraft. With the further development of the drifting snow storm, as shown in Fig. 9(b), more snow particles are scattered around the updraft bubbles although high concentration particle clouds are still in the wind bubbles. When drifting snow storm approaches its saturated state, snow particle clouds are almost connected together with numerous high concentration zones inside.

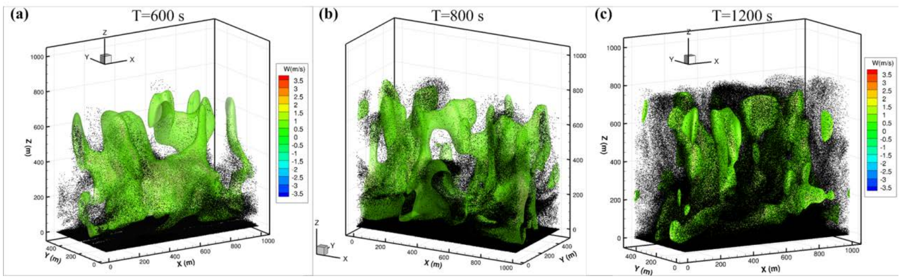

Figure 9. Evolution of drifting snow storm and vertical wind speed bubbles under friction velocity of 0.35 ms^{-1}, and wind bubbles are iso-surface of vertical wind speed with a value of 1.0 ms^{-1}.

The evolution of the depth of drifting snow storm can be divided into three typical stages. In sequence, these phases are the rapid growth phase, the gentle growth stage, and an equilibrium state, as shown in Fig. 10. Here, the depth of drifting snow storm refers to the average height of the topmost particle during this period (100 s). The rapid growth stage is mainly driven by large-scale turbulent motion, while the turbulent diffusion by the SGS fluctuating velocity is the main contributor to the gentle growth stage. The duration of second stage decreases with increasing friction velocity, which mainly comes from the stronger turbulent diffusion under larger

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friction velocities.

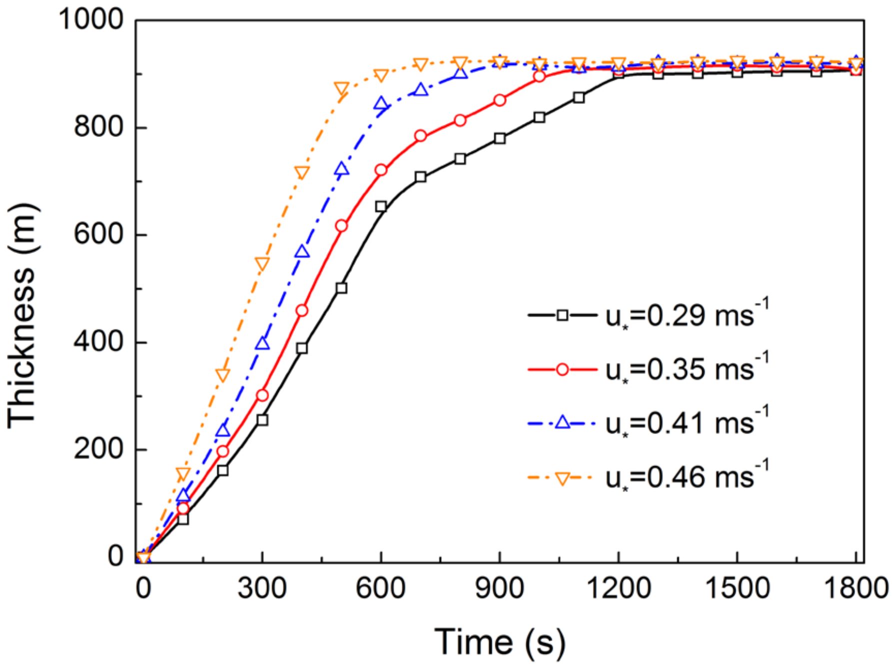

Figure 10. Time evolutions of the thickness of drifting snow storm under various friction velocities.

At the same time, the time required for the drifting snow storm to reach its maximum thickness decreases with friction velocity, ranging from about 1200 s to approximate 600 s when the friction velocity increases from 0.29 ms^{-1} to 0.46 ms^{-1}. The thickness of saturated drifting snow storms is almost constant with a value approximately 900 m under different friction velocities, probably because the boundary layer depth as well as the surface heat flux are unchanged. Thus, the final thickness of a drifting snow storm should be largely dependent on the maximum height of atmospheric turbulences.

4 Conclusion

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In this work, large-scale drifting snow storms are simulated in a large eddy simulation combined with a particle tracking model that includes subgrid scale velocity fluctuations. A typical drifting snow storm of several hundred meters in depth is generated, and the structure of the particle cloud with different concentrations is also produced. The transport flux profile has obviously different slopes near the surface compared to higher altitudes, that is, transport flux at near surface decreases with height sharply, but decreases more gentle at higher altitude. Previous studies may largely underestimate the total transport during drifting snow storms.

At the same time, the evolution of the thickness of drifting snow storm generally contains three stages. Drifting snow storm development generally begins with a rapid growth stage driven by the large scale atmospheric turbulent motions, followed by a gentle growth stage driven by the SGS fluctuating wind speed, before reaching an equilibrium stage when the drifting snow approaches a saturated state. The second stage becomes shorter with increasing friction velocity, mainly because stronger turbulence under higher friction velocity enhances the turbulent diffusion of particles.

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