Drifting snow sublimation: A high-resolution 3-D model with temperature and moisture feedbacks

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[1] The snow transport model of Alpine3D is augmented with a drifting snow sublimation routine. Contrary to other three-dimensional high-resolution snow transport models, Alpine3D now accounts for feedback mechanisms on the air temperature, humidity, and snow mass concentration in three dimensions. Results show that the negative feedbacks of sublimation on the snow mass concentration, temperature, and humidity are, in general, small but relevant on the slope scale. We analyzed the deposition on a leeward slope for simulations including sublimation and compared these to a reference simulation of the model without sublimation. Including sublimation, but neglecting sublimation feedbacks, leads to a reduction in deposition of approximately 12% on this slope. In a simulation including sublimation and its feedbacks, the reduction in snow deposition on the same slope was 10%. The feedbacks thus reduced the loss of snow due to sublimation by 2%. The sublimation process is therefore quite important for a leeward slope influenced by drifting snow. However, we also show that the spatial variability is large and that drifting snow sublimation will mainly affect small regions within a catchment. Averaged over our model domain (2.4 km^2) in the Swiss Alps, drifting snow sublimation causes a reduction in deposition of 2.3% during a 43 h test period, which is comparable to the sublimation loss from the snow cover during the same time.

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1. Introduction

[2] There are two ways in which a snow cover can lose mass due to sublimation: either directly, as in the case of surface sublimation, or indirectly, through the sublimation of snow transported by wind, i.e., drifting snow sublimation. Drifting snow sublimation has never been measured in an uncontrolled environment and model estimates vary greatly. The first controlled measurements of sublimation of snow grains in air were made by *Thorpe and Mason* [1966]. Single ice spheres, plates and dendrites were suspended on a fine fiber in a wind tunnel and the resulting mass loss was measured with a microbalance. This study was followed by similar measurement experiments, such as *Nelson* [1998], in which ice crystals were kept in a fixed position during sublimation. The only measurements of drifting snow sublimation where snow could freely move with the airflow are, as far as we know, those of *Wever et al.* [2009], in a closed system (wind tunnel). This study confirmed the significant effect of drifting snow sublimation.

[3] Sublimation is particularly relevant in the polar regions, where it can strongly influence the ice sheet and sea-ice mass balance [e.g., *Déry and Yau*, 2002]. The quantification of sublimation is also important in alpine terrain. For example, during warm and dry Föhn storms, suspended snow plumes that stem from crests and ridges are often observed dissipating downwind. The snow does not seem to reach the ground again. Possible explanations are that the snow plumes only seem to disappear because the snow becomes highly dispersed and the concentration of snow is no longer visible, or that the snow plumes dissipate because much of the drifting snow is sublimated. A combination of these two processes is most likely. In this model study, we investigate the magnitude of sublimation in such snow plumes, where in the plume most of the sublimation occurs and how sublimation affects the snow distribution in complex terrain.

[4] Drifting snow sublimation has been included in several physically based drifting snow models. Examples of one-dimensional models are PIEKTUK-T [*Déry et al.*, 1998], PIEKTUK-B [*Déry and Yau*, 1999], Windblast [*Mann*, 1998] and Snowstorm [*Bintanja*, 2000]. These models were compared in a thorough review [*Xiao et al.*, 2000], but have never been validated against measurements. In all of these models, the sublimation of drifting snow is accounted for based on the sublimation of a single ice sphere, as shown by *Thorpe and Mason* [1966]. However, as the models are only one-dimensional, advection effects cannot be included and they are not suitable for complex terrain. A distributed drifting snow model was developed by *Pomeroy et al.* [1997] based on the two-dimensional prairie blowing snow model [*Pomeroy et al.*, 1993]. This model is fetch dependent and

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describes the terrain as either sinks or sources of drifting snow. It can therefore estimate snow distribution. The sublimation routine, however, does not consider feedback on temperature or humidity. Examples of physically based, three-dimensional snow transport models are SnowTran-3D [*Liston and Sturm*, 1998], SYTRON3 [*Durand et al.*, 2005], and Alpine3D [*Lehning et al.*, 2008]. SYTRON3 uses a highly simplified parameterization for sublimation which only depends on a vertically averaged concentration of suspended snow and wind speed. SnowTran-3D has a more sophisticated parameterization of sublimation, depending on temperature-dependent humidity gradients between the particle and the atmosphere, conductive and advective energy- and moisture-transfer mechanisms, solar radiation intercepted by the particle and the particle size. The feedbacks on temperature, humidity and snow concentration, however, are not explicitly implemented. In Alpine3D, the sublimation of drifting snow has been neglected until now, but in this study we describe the implementation of a sublimation routine including the feedbacks.

[5] Thermodynamic feedbacks were studied by *Déry et al.* [1998] with a fetch-dependent drifting snow model. They found temperature decreases up to 0.5°C due to drifting snow sublimation and a significant increase in humidity near the surface. The sublimation process appeared to be self-limiting. Another study that specifically addressed drifting snow sublimation and the thermodynamic feedbacks is that of *Bintanja* [2001]. He used a one-dimensional model and showed that the thermodynamic feedbacks can have a strong influence. He found that, when removal mechanisms (such as advection) are weak, the air will quickly become saturated and the upward vertical turbulent flux of moisture might even be reduced in the presence of drifting snow. This latter result is unlikely and probably an artifact of not properly taking into account the vertical location of the local fluxes. Both studies showed the importance and complexity of thermodynamic feedbacks in drifting snow, but neither are suitable for complex terrain. Such feedbacks should be included in a three-dimensional model more suitable for complex terrain to investigate whether they are important in complex terrain as well.

[6] In alpine regions, few studies on drifting snow sublimation have been made, probably because drifting snow and the sublimation process in steep terrain are complex and the terrain is difficult to access. *Strasser et al.* [2008] simulated drifting snow sublimation with SnowTran-3D for the Berchtesgarden National Park and found that drifting snow sublimation has a large spatial variability. On crests and ridges, more than 1000 mm water equivalent (w.e.) would sublimate in one winter, which is roughly 70% of the local winter snowfall. Averaged over the complete domain, only 4.1% of snowfall was lost to sublimation from turbulent suspension. The study suggests a local significance of drifting snow sublimation in alpine regions. However, it reveals little about where in the catchment the deposition is mostly influenced by the sublimation of drifting snow and does not show how feedbacks can influence the sublimation, and thus, deposition. Furthermore, the simulations were based on wind fields with a horizontal resolution of 200 m, which has been shown to be inadequate for the type of terrain considered [*Mott and Lehning*, 2010].

[7] Most models neglect advection processes and temperature and humidity feedbacks on the sublimation of drifting snow. In this study, we focus on an iterative coupling of the three-dimensional transport of snow, humidity and temperature and quantify several feedbacks that are neglected in other models for complex terrain. We build on a sublimation model introduced by *Wever et al.* [2009], which was tested for ensembles of particles in a wind tunnel. The coupling of this model to a three-dimensional snowdrift model allows us to estimate the thermodynamic effects of sublimation. Furthermore, with a three-dimensional high-resolution model, we can address some of the local effects of drifting snow sublimation on the snow cover in complex alpine terrain. We predict a spatial lag between snow plumes and humidity plumes, and suggest that drifting snow sublimation is important for the snowpack evolution, particularly on the leeward sides of steep ridges.

[8] The focus of this study is on drifting snow sublimation. For simplicity and readability, we will often refer to this simply as sublimation. When we compare drifting snow sublimation with surface sublimation, we make it clear which of these we are referring to. We describe the drifting snow model and the numerical implementation in section 2 and our test site and a selection of observations during a case study in section 3. Section 4 focuses on our results and the influence of individual feedbacks, as well as the significance of drifting snow sublimation in an Alpine catchment. Conclusions are given in section 5.

2. Theory and Model

[9] Here we describe the drifting snow sublimation process, and then, rather generally, the snowdrift module of Alpine3D. Afterward, we explain how we coupled the sublimation routine to Alpine3D to find a steady state sublimation.

2.1. Drifting Snow Sublimation

[10] *Thorpe and Mason* [1966] estimated the mass loss of a single ice sphere due to sublimation with the following equation (omitting the influence of solar radiation):

$$\frac{dm}{dt} = \frac{2\pi r_0 \sigma}{\frac{L_s}{KT_a Nu}\left(\frac{L_s M}{RT_a} - 1\right) + \frac{1}{D\rho_s(T_a)Sh}}, \tag{1a}$$

where $\sigma = (\rho/\rho_s(T_a)) - 1$ is the supersaturation of water vapor with respect to ice, M (kg mol^{-1}) is the molecular weight of water, R (J mol^{-1} K^{-1}) is the universal gas constant, r_0 (m) is the particle radius, T_a (K) is the air temperature, D (m^2 s^{-1}) is the molecular diffusivity of water vapor in the atmosphere, K is the molecular thermal conductivity of the atmosphere (0.024 J m^{-1} s^{-1} K^{-1}), L_s (J kg^{-1}) is the latent heat of sublimation, ρ_s (kg m^{-3}) is the saturation density of water vapor, *Nu* is the Nusselt number, and *Sh* is the Sherwood number. The Nusselt and Sherwood number account for the ventilation of the particles and depend on the wind velocity and particle size.

[11] The particle size distribution is frequently given as a function of height, e.g., as by *Liston and Sturm* [1998]. We, however, assume that the radius (r_0 = 62.5 μm) is constant and that the particles in suspension are spherical to simplify

the calculations. This size was chosen based on model calculations of the sublimation for each particle of an ensemble of particles based on a gamma size distribution with a mean radius of 35 μm and $\alpha = 2$. We then searched for a single particle size that would give the same total sublimation for an equal concentration as the ensemble did (similar to *Wever et al.* [2009]), resulting in a particle size of 62.5 μm.

[12] The mean particle radius chosen was rather small compared to, for example, that in the work of *Xiao et al.* [2000], where a mean radius of 75 μm was assumed. The size is comparable to the mean radius at a height of about 3–4 m in other drifting snow models, such as SnowTran-3D and PIEKTUK-T. Although most particles are found close to the surface, we chose this rather small value because (1) we need to describe suspended particles at a height ranging from a few centimeters to a few meters above the ground, (2) the value should be representative for particles both close to and far away from their source, and (3) we assume that all particles are spherical. These assumptions lead us to conclude the particle must be rather small.

[13] Other findings support this conclusion. *Budd et al.* [1966] and *Schmidt* [1982] showed that the size of particles in suspension rapidly decreases with height, which means a somewhat smaller particle will be representative for a larger height range. Furthermore, there is a feedback between sublimation and particle size, probably causing particle size to decrease with the distance from the source. We do not take this directly into account in this study, but this effect can probably be canceled out by choosing a small particle size in a steady state situation.

[14] Our third assumption is supported by *Schmidt* [1982], who stated that ice crystals in wind-blown snow can be considered spherical as they rapidly lose their original precipitation crystal characteristics. This implies that, even though particles of many different shapes and sizes are present at the onset of drifting snow and close to the source of snow plumes, the spherical particle is more representative for the hourly steady state we use. Furthermore, *Wever et al.* [2009] show that their sublimation model tends to underestimate the sublimation, especially for experiments with freshly made dendrites. They believe that the underestimation compared to wind tunnel measurements may in part arise from not taking into account the irregularity of the snow crystals. However, they studied particles that were transported over only 13 m in less than 1 s. The particles therefore did not have much chance to become rounded, unlike in our hourly simulations. Nonetheless, we could reduce the general underestimation of sublimation arising from the assumption of spherical particles. By assuming the particle size is small in our case study, we can increase the number of particles, and thus the surface of all particles relative to the snow concentration and consequently sublimation. Therefore, even though the modeled sublimation is based on a spherical particle of fixed size, it is still representative of the sublimation of particles ranging broadly in size and shape.

[15] Equation (1a) solves the energy and mass transfer between a particle and the surrounding air based on the assumption of a thermal equilibrium. In this system, all energy flowing toward the particle is used for sublimation. In our case, we do not look at single particles but at snow plumes. As the plumes are highly turbulent and there are many snow particles within the plumes, there is an exchange between particles, particle boundary layers and surrounding air. Thus, the sublimation not only affects the particle boundary layer, but also the air temperature and the humidity of the plume. This effect can be accounted for by extending the sublimation of a single particle to an ensemble of particles and calculating the feedbacks on the plume. For a constant radius (r_0), the sublimation rate (S) (kg m^{-3} s^{-1}) is related to the snow concentration (c) via

$$S(\boldsymbol{x}, t) = \frac{dm}{dt}(T_a, \sigma, \mathbf{u}) \frac{c(\boldsymbol{x}, t)}{\frac{4}{3}\pi r_0^3}. \tag{1b}$$

Equation (1b) allows us to relate sublimation to the meteorological fields: suspended snow concentration, temperature and specific humidity, as shown in section 2.3. In Alpine3D, we use steady state fields of the mean wind. Therefore, we calculate snow transport in steady state during steps of 1 h. To be consistent with this approach, we use steady state sublimation as well. The procedure for retrieving steady state sublimation is explained in section 2.3.

2.2. Alpine3D

[16] Alpine3D is a model for alpine surface processes described by *Lehning et al.* [2006]. A description of the drifting snow module in Alpine3D is available from *Lehning et al.* [2008].

[17] The snow transport module combines high-resolution steady state wind fields from the mesoscale atmospheric model Advanced Regional Prediction System (ARPS) [*Xue et al.*, 1995; *Xue et al.*, 2001] with a saltation [*Clifton and Lehning*, 2008] and suspension model. The setup to retrieve mean flow fields (attained on a time scale of 10 s) is the same as that described by *Mott et al.* [2010]. The vertical resolution of the lowest level of ARPS varies between 0.7 m on ridges and 1.1 m on flatter terrain due to grid stretching. In this study we used a horizontal resolution of 10 m. Other important characteristics of the ARPS model are described by *Raderschall et al.* [2008]. *Lehning et al.* [2008] showed that mean flow fields can be used for snowdrift modeling in complex terrain.

[18] The mean flow fields are not calculated separately for each simulation time step. What we do is make a library of wind fields which are representative for several time steps within a simulation period. We then use a classification scheme to select which wind field we use at which particular time step, based on the wind data from one or more automatic weather stations. Except for the wind, Alpine3D is driven by meteorological data from weather stations.

[19] Snow cover is modeled by SNOWPACK [*Lehning and Fierz*, 2008], which is part of Alpine3D. The coupling between SNOWPACK and the drift module allows a realistic simulation of drifting snow, as the properties and the availability of snow at the surface are known.

2.3. Drifting Snow Sublimation in Alpine3D

[20] Drifting snow sublimation is implemented in the suspension module of Alpine3D. To introduce sublimation (S) within this model framework, the same advection-diffusion type of equation is assumed for specific humidity (q) and potential temperature (θ) as that already used for suspended

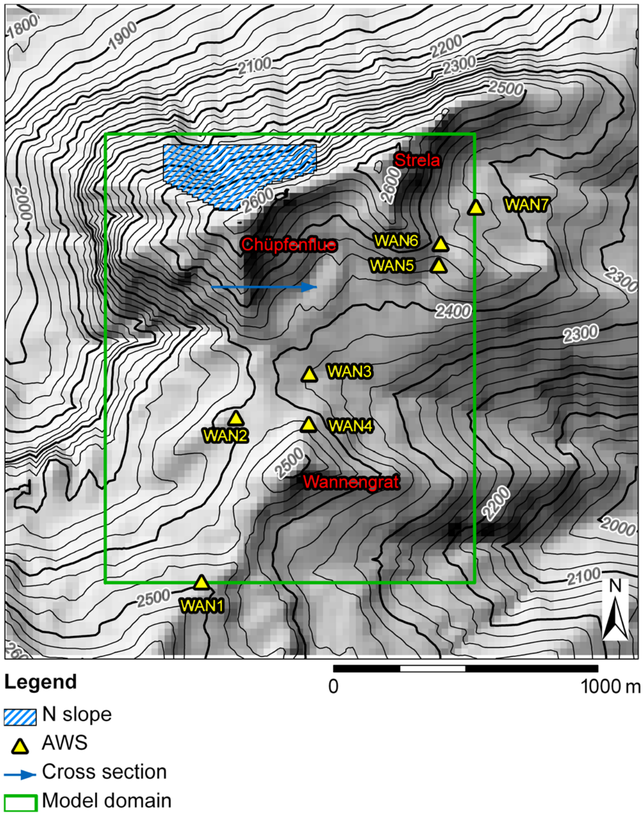

Figure 1. Map of the Wannengrat test site and the surrounding area, indicating the location of the automatic weather stations, the cross section used in section 4.2, and the north slope used in section 4.3. Three mountains are labeled in red. The interval between the thin lines represents 25 m, and the interval between the thick lines represents 100 m. The background shows the digital elevation model with standard hill shading.

snow concentration (c). This gives the following set of conservation equations:

$$\frac{\partial}{\partial \mathbf{x}}\left(K_c \frac{\partial c}{\partial \mathbf{x}}\right) + \mathbf{u}_p \cdot \frac{\partial c}{\partial \mathbf{x}} = -S, \quad (2)$$

$$\frac{\partial}{\partial \mathbf{x}}\left(K_q \frac{\partial q}{\partial \mathbf{x}}\right) + \mathbf{u} \cdot \frac{\partial q}{\partial \mathbf{x}} = \frac{S}{\rho_{\mathrm{air}}}, \quad (3)$$

$$\frac{\partial}{\partial \mathbf{x}}\left(K_\theta \frac{\partial \theta}{\partial \mathbf{x}}\right) + \mathbf{u} \cdot \frac{\partial \theta}{\partial \mathbf{x}} = -\frac{1}{\rho_{\mathrm{air}} \cdot C_P}(L_s S). \quad (4)$$

[21] In all equations, the first term represents turbulent diffusion, the second the advection and the third the sublimation as a source or sink. Here ρ_{air} is the density of the air, C_p is the specific heat capacity, and L_s is the latent heat of sublimation; $\mathbf{u}$ and $\mathbf{u}_p$ denote the wind speed (m s^{-1}) and particle velocity (m s^{-1}), respectively, and K (m^2 s^{-1}) is the diffusivity coefficient, which is attained from ARPS and is based on a 1.5 order subgrid-scale TKE-based turbulence closure scheme. As diffusion is closely related to turbulence, we do not use the diffusivity of mean flow fields, but rather that of flow fields that include turbulence. These turbulent flow fields are attained on a time scale of 70 s. Note that this set of equations does not take into account the effect of the changed temperature, humidity and snow concentration field on the flow dynamics as the flow field is calculated separately. However, we expect this to have only a minor effect.

[22] The sublimation is present as a source/sink term in all equations. It depends on all of the three state variables c, q, and θ via equation (1a), and thus couples the three conservation equations (2), (3), and (4). An iterative method was implemented in which this set of equations is solved several times until the steady state sublimation, consistent with temperature, suspended snow concentration and humidity, is found. The feedbacks between sublimation and snow concentration, humidity and temperature are negative as sublimation cools and saturates the air and leads to loss of snow mass. The initial sublimation is consequently reduced. The reduction depends on the initial conditions, as these determine the strength of the feedbacks. For instance, for a grid point in the model domain in a simulation where the suspended snow concentration does not limit sublimation and with an air temperature of 270 K and 95% relative humidity, the steady state sublimation was about 44% of the initial sublimation. The same conditions, but with a relative humidity of 70%, would in contrast, allow a steady state sublimation of only 8% compared to the much larger initial sublimation.

[23] The set of equations (2), (3), and (4) is solved with a biconjugate gradient solver as described by *Lehning et al.* [2008]. The boundary conditions are of Robin type for concentration. For humidity and temperature, we use Dirichlet boundary conditions at the lateral sides and top of the domain. For the bottom boundary we assume that there is no exchange of water vapor and energy as the sublimation of drifting snow will instantaneously saturate the saltation layer. This assumption might not hold when snow plumes overlie dry and warm air and latent heat exchange from the surface is still possible. We expect, however, that the temperature and humidity within the snow plumes at higher altitude will not be affected by this.

3. Experimental Setup and Data

3.1. Case Study

[24] We chose a typical spring storm where we expected sublimation to occur for our case study. We made several simulations for this storm which include either all feedbacks, no feedbacks or only feedbacks involving suspended snow concentration, humidity or temperature alone. Simulations were performed for Wannengrat (see Figure 1), a small catchment (2.4 km^2) near Davos in the SE of Switzerland. Meteorological observations from seven automatic weather stations (AWS) WAN1-WAN7, are available for the research area, providing records of air temperature, relative humidity, wind speed and direction at 2–3 m (depending on snow height), incoming longwave radiation, incoming and outgoing shortwave radiation, snow height and snow surface temperature. An overview of the test site is given in Figure 1.

[25] All meteorological data apart from wind direction and precipitation are from WAN3. We chose this AWS as it is positioned in the center of our research site on a relatively

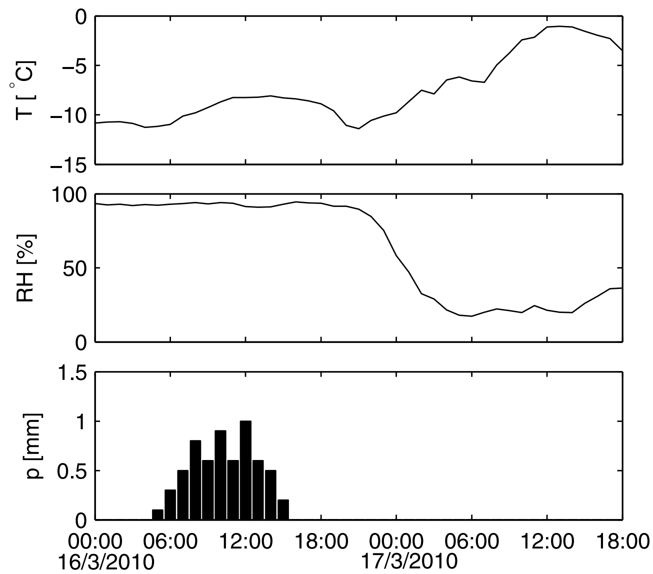

Figure 2. Meteorological observations (hourly averages) during the case study. (top) Air temperature (°C) and (middle) relative humidity (%), both about 2 m above ground, were measured at WAN3, and (bottom) precipitation (mm) was measured at Weissfluhjoch Versuchsfeld.

flat field. Furthermore, the humidity and temperature measurements at WAN3 are influenced by drifting snow sublimation only during very strong winds, according to a comparison between relative humidity and wind speed. During our case study, the wind was too weak to moisten the air near the station through drifting snow sublimation. Measurements of air temperature, relative humidity, longwave radiation, shortwave radiation and snow surface temperature were used as input for our simulations. We assume that air temperature follows a dry adiabatic lapse rate for an initial temperature profile. For the initial humidity field, we assume that the specific humidity is constant across our model domain.

[26] No precipitation measurements are available within the catchment. We measure snow height, which is affected by drifting snow. Precipitation amounts can only be derived from snow height if the settlement of the snowpack is accounted for [*Lehning et al.*, 1999]. Therefore, we used hourly precipitation observations from a nearby flat-field station, Weissfluhjoch Versuchsfeld, located at a similar altitude and 4 km northeast of Wannengrat.

[27] As a case study, we chose a period of 43 h, 16–17 March 2010. This was a spring situation with favorable conditions for drifting snow sublimation. At the beginning of this period, there was snowfall accompanied by a strong NW wind. This was followed by several hours with a warm, dry south wind, with the NW wind eventually returning at the end of the period. The observed temperature, humidity and precipitation are shown in Figure 2. More information on the observed and simulated wind is given in the next section.

3.2. Wind Fields

[28] The driving force behind drifting snow is wind. We therefore took care to produce realistic wind fields for the studied terrain. Below we briefly describe them and discuss their quality compared to the observations.

[29] In section 2.2 we mentioned that the wind fields are chosen according to wind observations. This requires representative stations for the main wind speed and synoptic wind direction in the area. At Wannengrat, the classification scheme is based on the frequency distribution of wind velocity observed at WAN3 (see Figure 1) and the wind direction observed at WAN2 (see Figure 1). In our case study, a set of five wind fields was used to drive the Alpine3D simulations.

[30] Using a library of wind fields means that wind speeds vary less than actually observed wind speeds as several values of wind speed (in a range of approximately 1–2 ms^{-1}) at one station are represented by the same wind field. Moreover, the decoupling between the wind field calculation and drifting snow simulation does not allow any feedback from the snow cover on the wind field. We do not account for the surface evolution with increasing snow height and the smoothing of the terrain by the snow cover [e.g., *Schirmer et al.*, 2011].

[31] A full validation of the wind fields is beyond the scope of this study. Other studies, such as *Mott et al.* [2010], have already shown that quite a good estimate of the snow distribution in this catchment (and therefore of drifting snow) could be obtained with similar wind fields, which are qualitatively validated by *Mott et al.* [2010]. Here we will thus restrict ourselves to describe the quality of the wind fields based on measured and simulated wind speeds at three stations (Figure 3). We chose these specific points because they represent a rather sheltered flat field area (WAN3) as well as more exposed sites (WAN2 and WAN6). The simulated wind speed shown in Figure 3 was actually obtained from the 3 m wind speed. As we use a grid with a 10 m resolution, the modeled local wind speed was estimated by an interpolation of the wind speed at the grid points closest to the station. From the interpolated 3 m wind speed, we retrieved the wind speed at the height of the measurements using a logarithmic profile based on a roughness length of 1 cm, which is typical for such terrain [*Doorschot et al.*, 2004]. The height of the wind measurement was about 2.3 m (depending on snow cover).

[32] The most striking feature in Figure 3 is probably the peak wind speed, which is underestimated. Wind field modeling confronts most difficulties during periods with either very weak winds, when ARPS tends to overestimate the wind speed but the measurements are less reliable, or with strong winds, when simulations tend to become numerically unstable. During the period with southerly winds at WAN2 (16 March, 2000 LT to 17 March, 0600 LT), the wind direction at other stations seems to have been poorly represented. At WAN6, we have observed two dominant flow regimes during the south wind since the installation of the weather station. The wind either comes from south to SW or from west to NW. With ARPS, we only capture the W-NW stream during south wind. For WAN3, we observed north wind in this period with south wind at WAN2. This is a very rare case as during all other periods with south wind we have studied, the wind direction at WAN3 was similar to that at WAN2. We must conclude that there were either problems with the observations, such as icing or the accuracy was decreased due to the low wind speed. Or, alternatively, it

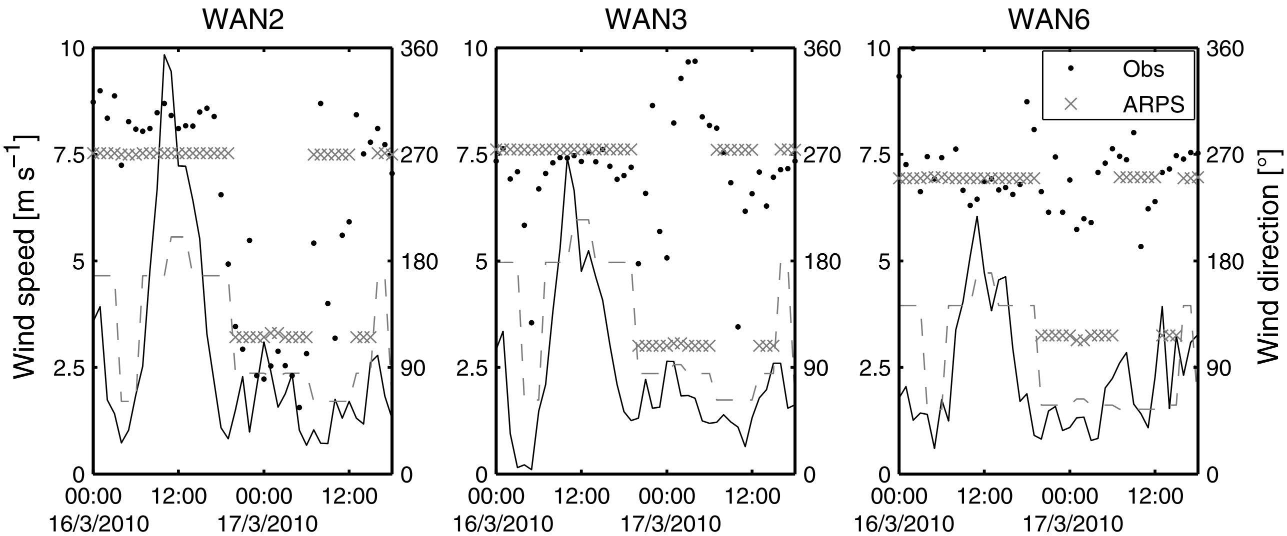

Figure 3. The observed (black) and simulated (Advanced Regional Prediction System) (gray) wind at WAN2, WAN3, and WAN6 during the case study. The wind speed is represented with lines (solid line, observation; dashed line, simulation), and wind direction is represented with points (circle, observation; cross, simulation).

may not be possible to capture such a rare case with ARPS as it might be, for instance, a thermally driven flow.

[33] We maintain that we have been able to represent the wind reasonably well, especially for the grid resolution used, but over the full period we tend to overestimate the wind speed. This can lead to inaccuracies in the estimates of deposition and sublimation quantities. We probably somewhat overestimate erosion and therefore the snow concentrations in snow plumes, and consequently, the sublimation for the period. Nonetheless, we are able to

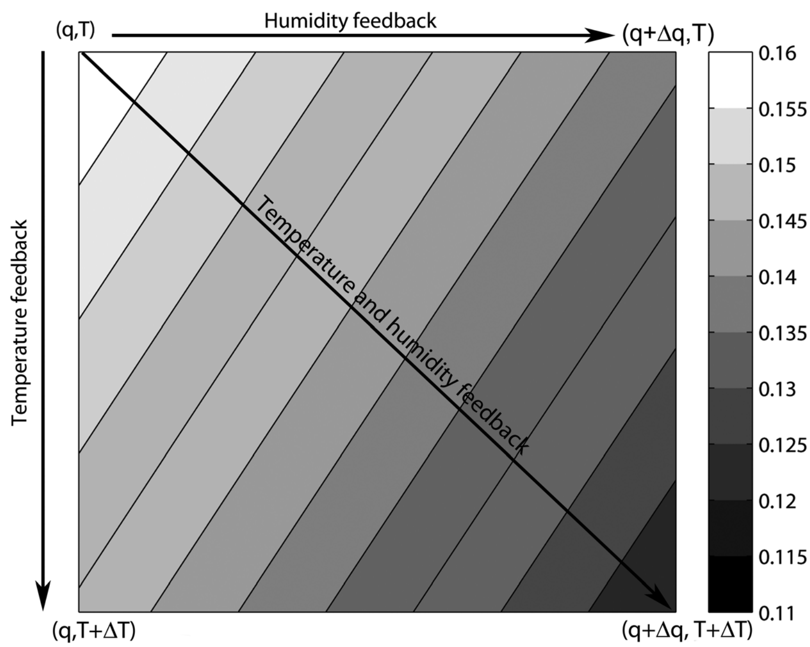

Figure 4. Sublimation rate (g m^{-3} s^{-1}) as a function of temperature and humidity feedbacks. The *x* axis represents the magnitude of the humidity feedback Δq, and the *y* axis represents the magnitude of the temperature feedback ΔT with respect to the initial sublimation after an interval of 1 s. The image can be interpreted by following the arrows from the upper left corner to another corner. The shaded contour at the first corner represents the initial sublimation (no feedbacks), and the gray scale at the second corner is the possible sublimation when the temperature, humidity, or both have been changed through sublimation feedbacks applied over 1 s. This image demonstrates that the humidity feedback is stronger than the temperature feedback.

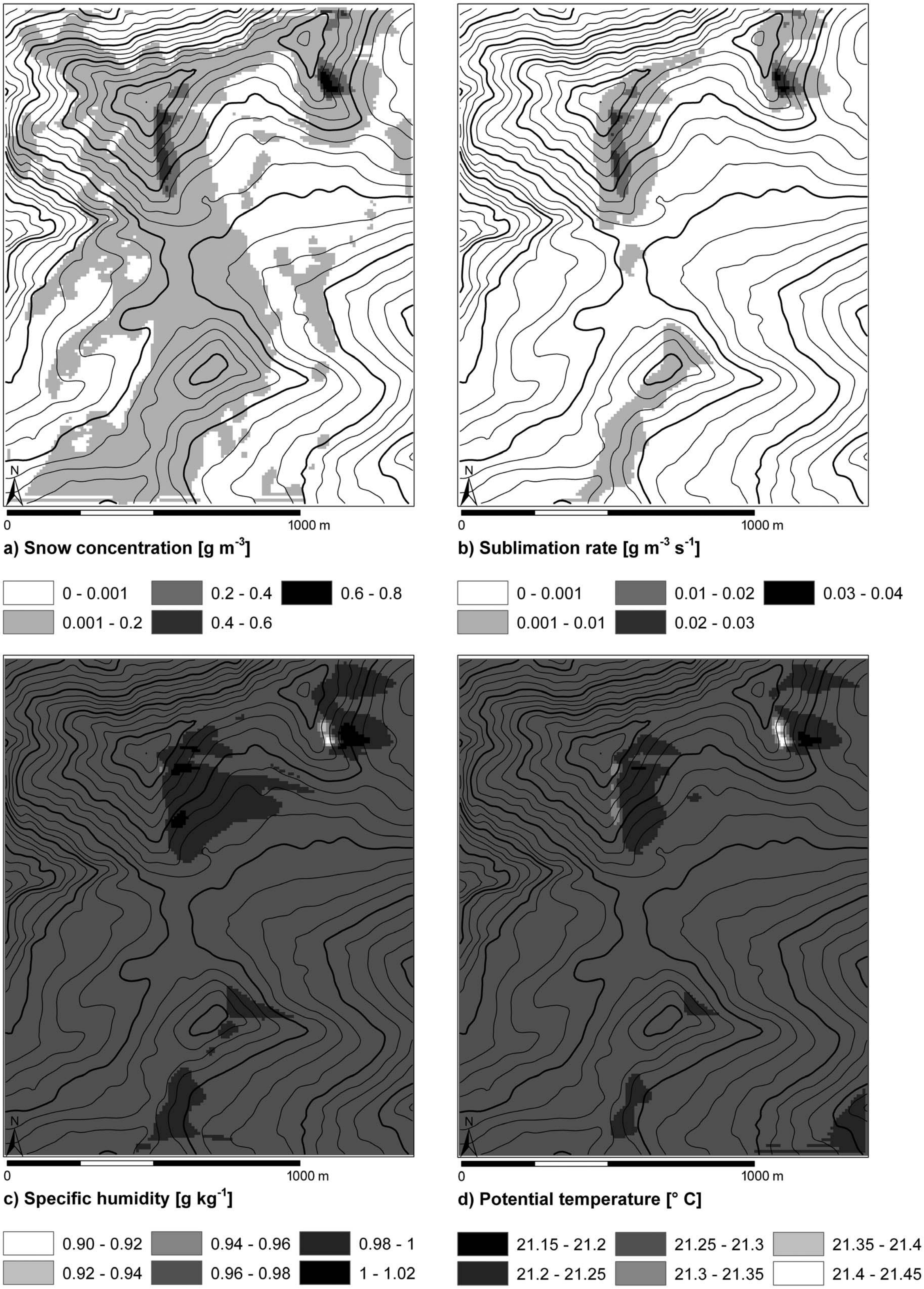

Figure 5. Vertical mean of (a) suspended snow concentration, (b) sublimation rate, (c) specific humidity, and (d) potential temperature within our testing domain for moderate NW wind on 17 March 2010 at 1600 LT. The lines show the topography: the intervals between the thin and thick lines represent 25 m and 100 m, respectively. The image shows that sublimation occurs mainly close to the drifting snow sources (the ridges) and that there is a spatial lag between humidity and snow plumes.

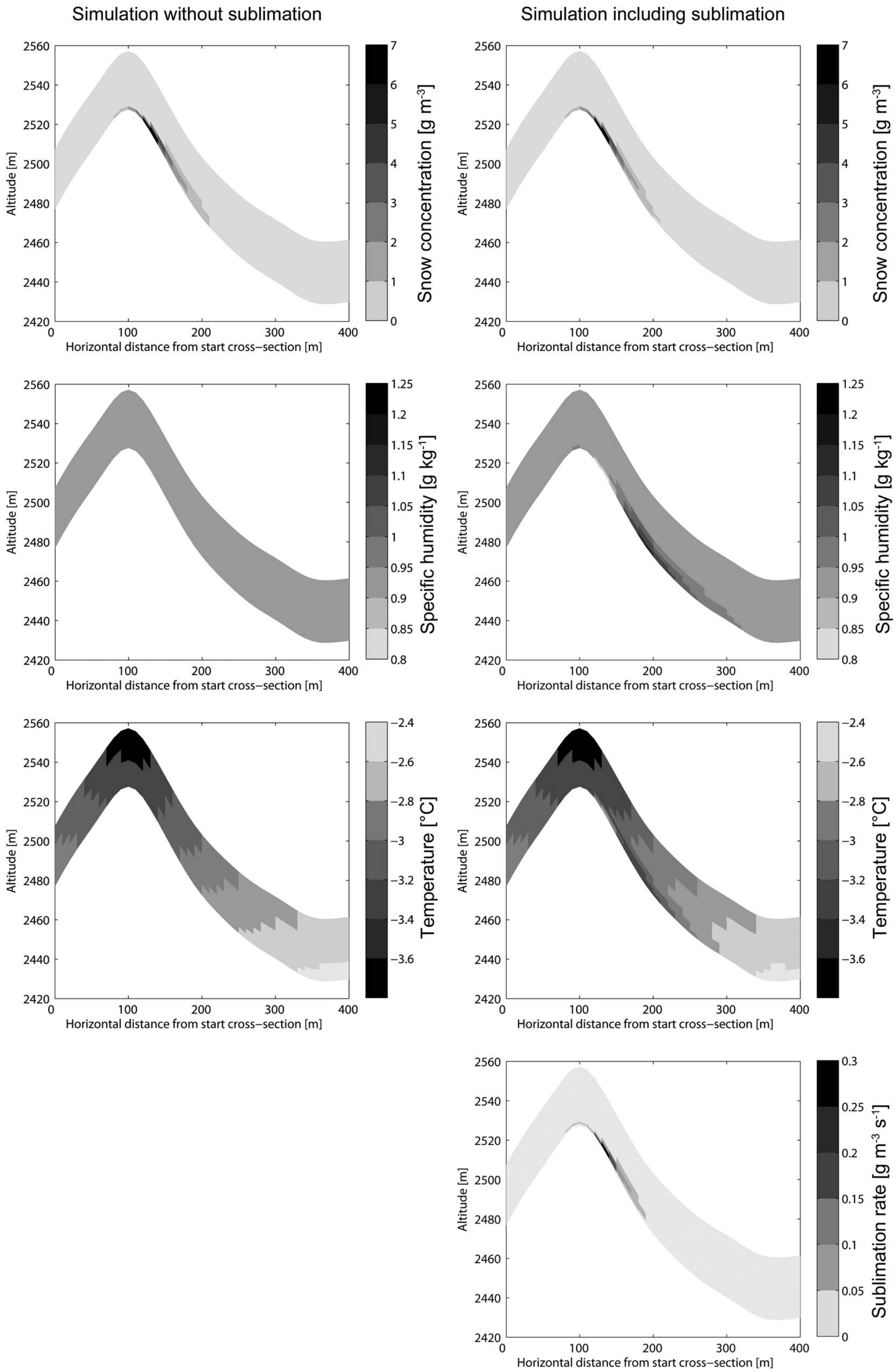

Figure 6. Suspended snow concentration (g m^{-3}), specific humidity (g kg^{-1}), temperature (°C), and sublimation rate (g m^{-3} s^{-1}) for moderate NW wind on 17 March 2010 at 1600 LT: (left) simulation without sublimation; (right) simulation including sublimation and feedbacks. W-E vertical cross section of Chüpfenflue. The location is indicated in Figure 1. The x axis shows the horizontal distance (m) from the start of the cross section, and the y axis shows the altitude (m).

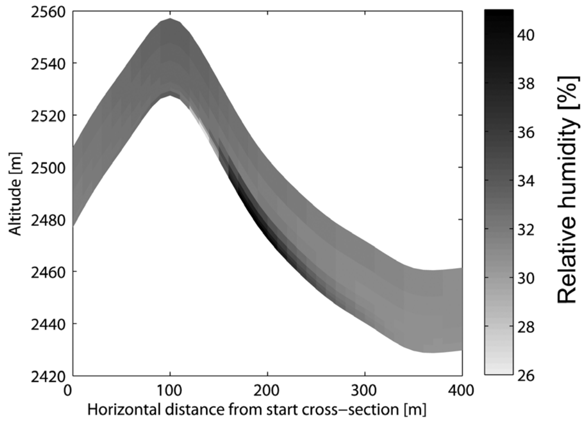

Figure 7. Relative humidity (%) in contours for a simulation with sublimation and moderate NW wind on 17 March 2010 at 1600 LT. W-E cross section of Chüpfenflue. The location is indicated in Figure 1. The *x* axis shows the horizontal distance (m) from the start of the cross section, and the *y* axis shows the altitude (m).

capture the main features of wind speed and direction with realistic values in this very steep terrain, which allows us to estimate sublimation in this catchment.

4. Results and Discussion

[34] Here we first focus on thermodynamic feedbacks, showing why they need to be accounted for and how they affect sublimation, the suspended snow concentration, or temperature and humidity fields. Then we will use our case study to show how important the sublimation is for the snow distribution in complex terrain.

4.1. Thermodynamic Feedbacks: Idealized Case

[35] As expressed by equation (1a), sublimation of an ice sphere is dependent on both the temperature and the humidity of the air. Since sublimation is a source of water vapor and requires energy, the humidity and temperature of the air will change as a function of sublimation, and have a major impact on estimates of total sublimation. One theoretical case is illustrated in Figure 4, where we assume a suspended snow concentration of 0.01 kg m^{-3}, a temperature of 270 K and a relative humidity of 70% as initial values. We used equations (1a) and (1b) to calculate the sublimation. The temperature was chosen to be close to the freezing point as the temperature influence on the sublimation is greatest near the freezing point. Figure 4 shows how the sublimation decreases as a result of temperature and humidity feedbacks. We start with an initial sublimation of 0.159 g m^{-3} s^{-1} shown in the upper left corner. If this sublimation is applied for 1 s, the amount of water vapor that is added to the air due to the humidity feedback is 0.12 g kg^{-1}. The sublimation possible at the same temperature, but with a slightly increased humidity (solving equation (1a) for T and $q + \Delta q$), is 0.134 g m^{-3} s^{-1}, which is only 84% of the initial sublimation. The change in sublimation due to the humidity feedback is expressed in Figure 4 with the contours between the upper left corner and the upper right corner.

[36] The same procedure (solving equation (1a) for q and $T + \Delta T$) can be done for temperature. In this case, the sublimation decreases to 91% of the initial sublimation due to the temperature feedback (shown in Figure 4 as the color difference between the upper left and lower left corners). When both feedbacks are accounted for, the sublimation is reduced to 74% of the initial value (see Figure 4, from the upper left to lower right corner.) This reduction in sublimation indicates that the humidity feedback is stronger than the temperature feedback, and that both feedbacks are significant. Consequently, both feedbacks were implemented in Alpine3D.

4.2. Thermodynamic Feedbacks: Simulation for Real Topography

[37] With a three-dimensional model that includes the advection and all feedbacks, it is possible to show where the most sublimation in the air occurs as well as where these impacts most influence snow deposition. To give a qualitative overview of the spatial influence of the thermodynamic feedbacks, we show the vertically averaged sublimation rate, humidity, potential temperature and suspended snow concentration for a simulation that includes all feedbacks (Figure 5).

[38] We chose a case with wind coming from NW (without precipitation), as the snow distribution in our catchment is dominantly influenced by NW wind [*Schirmer et al.*, 2011] and the full extents of the snow plumes at Chüpfenflue (see Figure 1) are captured in the model domain.

[39] Typically, drifting snow plumes start on ridges, and they have the largest suspended snow concentration at their start. The same pattern can be observed for drifting snow sublimation, which is highly dependent on the suspended snow concentration. However, the effects on humidity are visible further downstream in the snow plume. There, the air had longer exposure to drifting snow sublimation and the saturation effects are strongest (see Figures 5 and 6). Note that a simulation without feedback of sublimation on humidity would show a constant field of specific humidity. Therefore all values larger than 0.98 (g kg^{-1}) in Figure 5c indicate a moistening effect due to sublimation. Why values smaller than 0.98 (g kg^{-1}) occur even though we added a source of moisture is explained in the Appendix. Potential temperature has a similar effect to humidity, but smaller. The extent of the region where the temperature is affected by sublimation is not much larger than the snow plume itself.

[40] A vertical cross section through a snow plume is given in Figure 6. For all variables, we show the results of a simulation with and without sublimation for the same simulations as in Figure 5. The location of the vertical cross section is indicated in Figure 1. The spatial lag between the snow plume and the temperature and humidity plume is visible in Figure 6. Just as for humidity, the center of the maximum suspended snow concentration (roughly at 120 m from the start of the cross section) is not identical to the region predominantly cooled by sublimation (roughly at 180 m from the start of the cross section). Moreover, we observe a slight decrease in the extent of the snow plume in the vertical cross sections of snow concentration due to sublimation.

[41] The model indicates that the temperature on the leeward slope beyond the snow plume increases markedly. This is due to the initial temperature, which was defined with an adiabatic lapse rate so that the temperature rises with decreasing altitude.

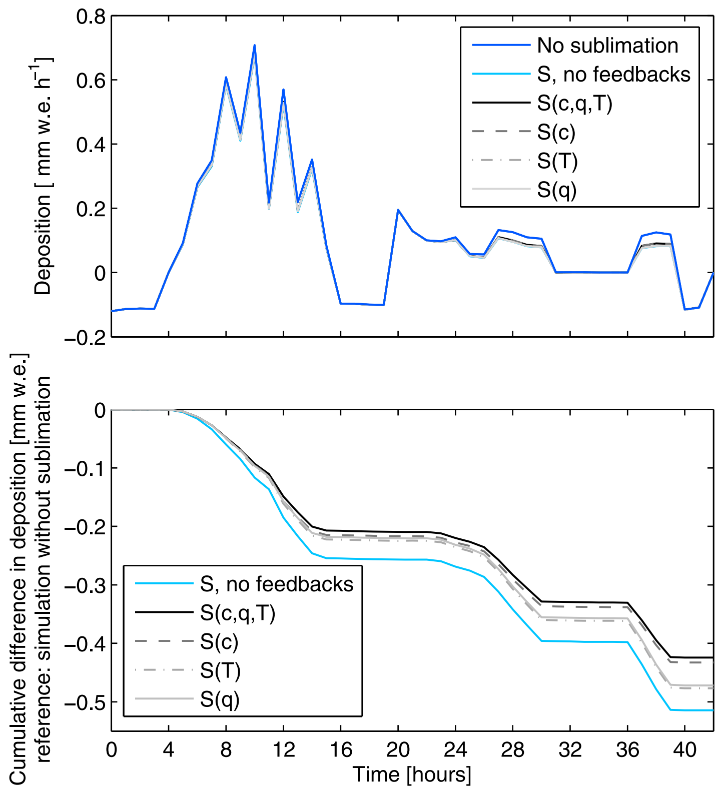

Figure 8. (top) Deposition (mm w.e. h^{-1}) and (bottom) the cumulative difference (mm w.e.) from that in a simulation without sublimation on the north slope (see Figure 1) for different simulations with sublimation with and without feedbacks. The simulation time step is 1 h.

[42] Sublimation in an uncontrolled environment cannot be measured directly with current measurement techniques, but indirect measurements to try to identify the effects of sublimation are possible. Our model suggests that sublimation causes a clear pattern in the humidity field and a small effect in the temperature field, and it would be interesting to see whether these effects can also be measured. The temperature changes predicted by the model are quite small, usually less than 1 K, and are thus within the error range of most standard measuring equipment. The relative humidity effect, however, should be possible to measure. In Figure 7, we show the modeled relative humidity for the same cross section as in Figure 6 in a simulation with sublimation. According to the simulation, the relative humidity can increase by up to approximately 15% due to drifting snow sublimation. On a leeward slope, it should be possible to detect this signal with a simple set of humidity measuring devices aligned with the snow plume. This might be a way to indirectly validate the model. No measurements could be carried out, however, on this site because the slope is steep and inaccessible as the avalanche risk is quite high.

4.3. Effects of Sublimation on the Snow Deposition on a Leeward Slope

[43] As explained in section 2, the feedbacks of both temperature and humidity are expected to have a significant impact on the sublimation amount. We have already shown how the spatial variability of the different fields is changed by sublimation and its feedbacks. We will now show how the single feedbacks influence the snow distribution. For this, we selected the slope where the sublimation had a relatively strong influence during our case study (see section 4.4). The location of the slope, which we will call the north slope, is indicated in Figure 1. For this region, we calculated the average deposition for each time step and the cumulative deposition for different simulations. The results are shown in Figure 8.

[44] Starting with deposition (Figure 8, top), only small changes due to sublimation ($S(c, q, T)$ versus no sublimation) are visible within the first 24 simulation steps. The reason is that during this period there is either snowfall with nearly saturated air or insufficient wind for drifting snow to occur. Once the southerly wind starts, bringing dry and warm air, we start to see some larger effects of sublimation on the deposition on this slope. The reduction of deposition due to the individual feedbacks varies between time steps. This is because the atmospheric conditions are crucial for the feedbacks. When it is warm and dry, initial sublimation is quite large and the feedbacks are stronger.

[45] In the plot of the cumulative difference in deposition (Figure 8, bottom) we can see the effects of the three separate feedbacks ($S(c)$, $S(q)$, and $S(T)$). The cumulative deposition for a simulation with sublimation including only the concentration feedback ($S(c)$) differs the most from the simulation without feedbacks (S, no feedbacks). We can therefore conclude that, in this situation, the suspended snow concentration feedback was most important. In this specific case, inclusion of all feedback mechanisms ($S(c,q,T)$) led to a result similar to a simulation with only the concentration feedback ($S(c)$).

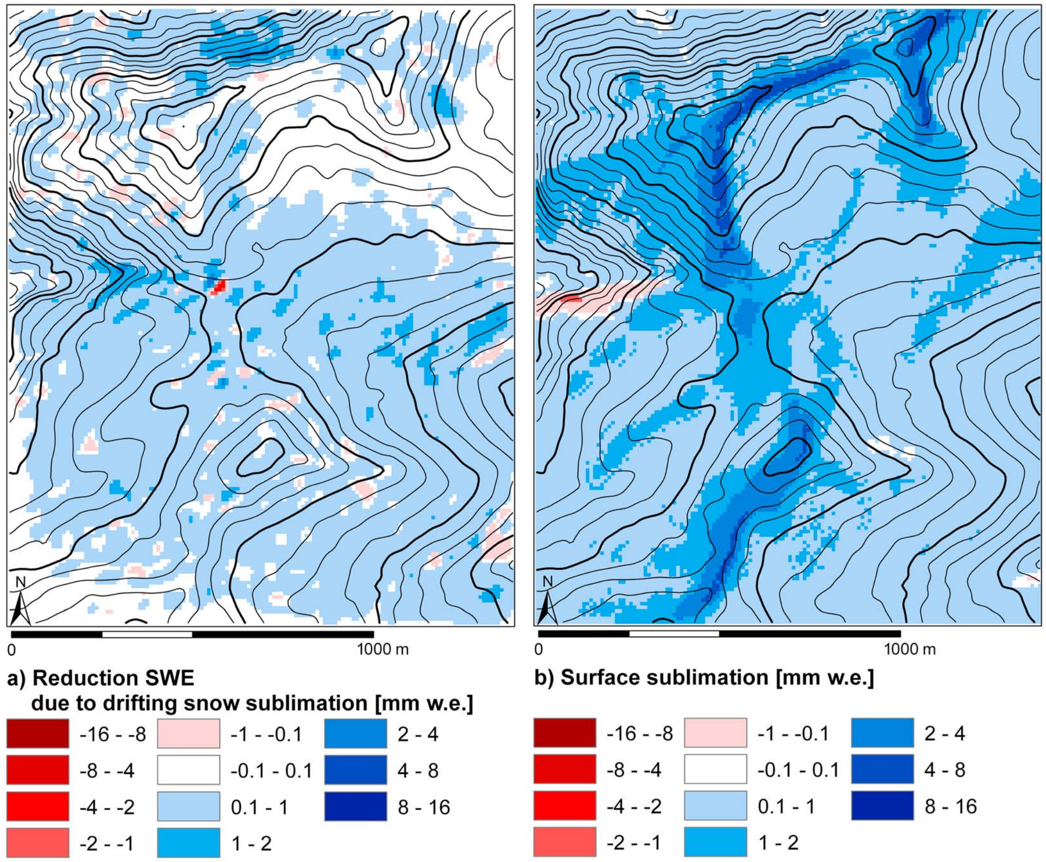

Figure 9. (a) Difference in snow water equivalent of the snow cover (SWE) (mm w.e.) at the end of the simulation between a simulation with and without sublimation (filtered with a low-pass filter). Positive values indicate that drifting snow sublimation has reduced the snow amount. (b) Sum of the latent heat flux from the surface over the complete test period (mm w.e.) (defined positive upward) for a simulation without drifting snow sublimation. The lines show the topography: the intervals between the thin and thick lines represent 25 m and 100 m, respectively.

[46] Absolute differences in deposition are quite small and its relevance can be best discussed by looking at the relative differences in deposition. A simulation without drifting snow sublimation gives a mean total deposition of 4.41 mm w.e. By aiming to improve the model with the inclusion of drifting snow sublimation, but without the feedbacks on temperature, humidity and snow concentration, the deposition is decreased by about 11.7%. However, when actually including drifting snow sublimation and all feedbacks, we estimate a deposition of 4.0 mm w.e., which means that sublimation reduces the deposition in this slope by approximately 9.6%. Looking solely at the drift period from hours 20 to 42, sublimation decreases deposition over this period by about 16%.

4.4. Effects of Sublimation on the Snow Distribution

[47] In the simulation, deposition was found to be reduced by sublimation on the north slope, but does sublimation affect the snow distribution in the whole domain? The differences in the snow water equivalent of the snow cover (SWE) at the end of the case study in a simulation with and without sublimation (including feedbacks) are shown in Figure 9a.

[48] At some points the snow mass increases when there is sublimation. This is a numerical effect, which is discussed in the Appendix. The sink of snow changes the steady state snow concentration field. In an advection dominated situation, this can numerically lead to an increase in the snow mass (and therefore in snow deposition), which is compensated for by a corresponding decrease elsewhere. As shown in the Appendix, the difference in the amount of snow at the end of the test period between two simulations is quite pixilated due to shifts in deposition (essentially caused by numerical issues). We therefore applied a low-pass filter twice on the field shown in the Appendix, resulting in Figure 9a. This gives a better view of the effect of drifting snow sublimation, which is easier to interpret.

[49] Sublimation had largest effects on the north slope when the southern wind brought warm dry air. In Figure 9, however, we see that most of the domain has been affected by drifting snow sublimation. The mean total reduction in drifting snow by sublimation within our model domain is approximately 2.3% of the mean deposition within this period. This value is slightly smaller than the result of *Strasser et al.* [2008] (4.1%), although a fair comparison is not possible as we only looked at a brief period and would

hesitate to speculate about the significance of drifting snow sublimation during a whole season. It is, however, realistic to assume that, for a full season, the average effect will be smaller since we picked a period with a strong sublimation event. This will be a subject for a future study.

[50] Nonetheless, we can, with the present model, compare the effects of drifting snow sublimation with those of other processes influencing the snow mass. As the conditions for drifting snow sublimation are quite suitable for surface sublimation as well, a comparison with the latent heat flux within our catchment could give us a general idea about the potential impact of drifting snow sublimation. We show that deposition is reduced by drifting snow sublimation (Figure 9a) and the total latent heat flux (surface sublimation) during the case study in Figure 9b. Although the spatial variability of the latent heat flux is quite different, the total amount of snow mass lost within our catchment is of the same order of magnitude. This means that, for the averaged snow mass balance in our domain, the surface and drifting snow sublimation have a similar influence. Other studies have compared drifting snow sublimation and surface sublimation, including *Strasser et al.* [2008] and *MacDonald et al.* [2010]. *MacDonald et al.* [2010] studied two seasons in the Canadian Rocky mountains and found a drifting snow sublimation of 86 mm w.e. during the 2007/2008 season along a transect and 69 mm w.e. during the following season (2008–2009). The surface sublimation was 2.8 mm w.e. in the first year and 62 mm w.e. the second, but they did not discuss the difference between the two seasons. In the second year they had a surface sublimation similar to drifting snow sublimation, which we found during our short study period as well, but the first year was completely different. *Strasser et al.* [2008] showed a seasonal drifting snow sublimation of 26.5 mm w.e. and a surface sublimation of 29.1 mm w.e., which is again a drifting snow sublimation amount in the same order of magnitude as the surface sublimation.

[51] Preliminary results of this study [*Groot Zwaaftink and Lehning*, 2010] showed a larger effect of sublimation. Here, a different set of wind field simulations was used, which later turned out to be inadequate. Some inconsistencies in the setup of the southern and northwestern wind fields were found and the wind velocity was overestimated. The wind fields used in the present study coincide better with AWS measurements in our domain, with a lower wind velocity, especially during southerly winds. This means there is less drifting snow after the snowfall period, and deposition in the case study is dominated by precipitation rather than drifting snow. Thus, the influence of drifting snow sublimation seems to be smaller than indicated in the preliminary results. A detailed analysis of sublimation sensitivity on the simulated wind fields may be given in a future publication.

5. Conclusions

[52] In this study, we implemented a routine for sublimation of drifting snow in Alpine3D. In contrast to other three-dimensional models, we included the feedback of sublimation on air temperature, suspended snow concentration and humidity. With our high-resolution model, we were able to show how sublimation affects the snow distribution very locally. Obviously, the spatial distribution of drifting snow sublimation is strongly related to the spatial distribution of drifting snow. However, we showed that the influence on snow concentration is slightly larger in the tails of the snow plumes, as sublimation shortens them. A similar effect was visible on the temperature and humidity: along the snow plume the cooling and moistening increases as the effects of sublimation build up, even though maximum sublimation values are found close to the source, where suspended snow concentrations are largest. These results suggest that not only the feedbacks, but also advection have considerable effect throughout the plume.

[53] On the leeward slope we selected, we saw that drifting snow sublimation can reduce the deposition by a substantial amount (9.6%). The spatially averaged mass loss due to sublimation was 2.3% of the deposition for the 43 h case we studied. This suggests that the sublimation of drifting snow is not only important for modeling the snow distribution, but could also be a significant mass sink over a larger area and time period. It should therefore be accounted for in mass balance calculations. A comparison with the latent heat flux from the surface showed that the mass loss due to drifting snow sublimation is of the same order of magnitude as the surface sublimation, but is concentrated mostly on the leeward slopes.

[54] A disadvantage of this model is that it does not take the feedback on the particle size distribution into account and it assumes that all particles are spherical. It can be expected that particle size will decrease with distance from the source. As it decreases, the surface area will become larger relative to the volume and sublimation is likely to increase. However, most sublimation occurs at the start of the snow plume where particle sizes are larger. The feedback on particle size would be centered downstream of the source (just as for humidity and temperature), where we already see a larger moistening effect from the sublimation. We believe that other feedbacks are likely to be limiting factors here, which would mean that the size effect is rather small. The assumption that the particles are spherical will become more realistic if drifting snow occurs longer and the distance from the source of the particles increases. As we estimate drifting snow with an hourly steady state assumption, spherical particles are most representative, but we reduced the size of the spheres to indirectly account for irregular shapes. Since sublimation is quite sensitive to the particle size and shape, further study on this topic should clarify whether our assumption is justified.

[55] When interpreting the results, it should be kept in mind that our case study is probably not representative for a full season at the catchment. We can show how feedbacks and advection act to limit drifting snow sublimation three-dimensionally and how sublimation affects the snow distribution on a short time scale, but it is difficult to predict their effects on the snow mass balance in the catchment for a full season. Further work is needed to quantify the effects of drifting snow sublimation on a seasonal alpine snow cover and larger areas. However, the results are transferable to any storms with similar temperature and moisture conditions, which may occur frequently in other midlatitude mountain ranges such as the Andes, Rocky Mountains or Himalayas.

Appendix: Numerical Effects of Steady State Sublimation

[56] To calculate the steady state fields we use a numerical solver. The results will be mathematically correct, but might not always be physical. Moreover, the input fields and

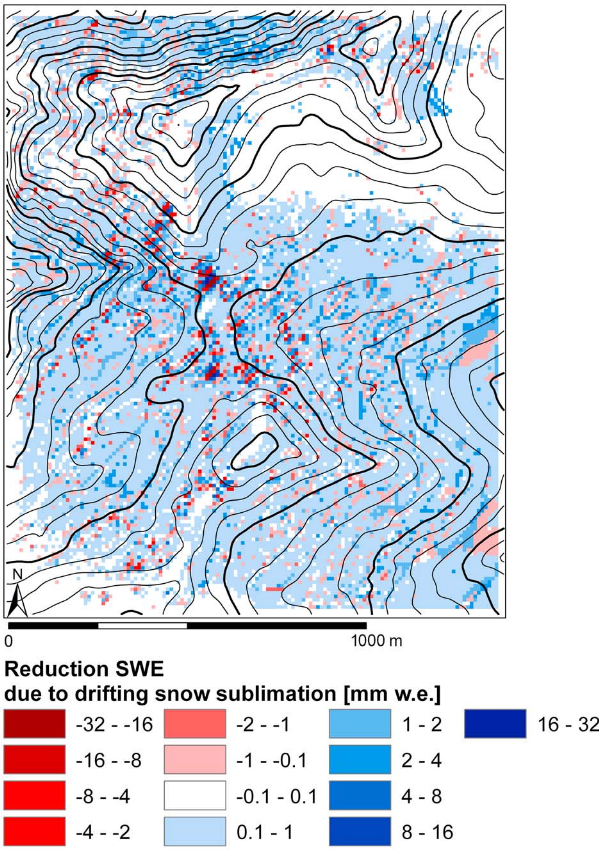

Figure A1. Difference in SWE (mm w.e.) at the end of the simulation between a simulation with and without sublimation. The lines show the topography: the intervals between the thin and thick lines represent 25 m and 100 m, respectively.

boundary conditions have to be chosen carefully. On ridges the air sometimes dries when we introduce sublimation into the simulation. This only occurs at the start of snow plumes where the gradient of the wind is strong. We briefly want to clarify this effect, even though its influence on the results is small. It helps to consider a highly simplified one-dimensional system without diffusion. The system consists of three points 10 m apart. We try to retrieve the humidity for a steady state at the middle point according to the following scheme:

$$q_1 = 1.2 \xrightarrow{u_a=10} \underset{S=0}{q_2 = 1.2} \xrightarrow{u_b=10} q_3 = 1.2.$$

If we prescribe values for the wind, q_3 and q_1, we can simply find q_2 as for a steady state we have $u_a\ ((q_2 - q_1)/\Delta x) = u_b\ ((q_3 - q_2)/\Delta x)$. If we start with $q_1 = q_3 = 1.2$ g kg^{-1}, then q_2 will be 1.2 g kg^{-1} as well.

[57] When we want to account for the feedback of sublimation on humidity, we have to introduce a source (e.g., +0.15 g kg^{-1}) at the middle point. We keep $q_1 = q_3 = 1.2$ g kg^{-1} for simplicity. We then have to solve $u_a\ ((q_2 - q_1)/\Delta x) + S = u_b\ ((q_3 - q_2)/\Delta x)$ and find $q_2 = 1.13$ g kg^{-1}:

$$q_1 = 1.2 \xrightarrow{u_a=10} \underset{S=+0.15}{q_2 = 1.13} \xrightarrow{u_b=10} q_3 = 1.2.$$

At this point we have found a steady state, but even though we added a source of moisture, the humidity is smaller. This situation will, of course, not occur in reality as q_3 would change. Restricted by the numerical solver, however, we are forced to use Dirichlet boundary conditions for humidity, which means that in snow plumes close to the boundaries of our domain, the above effect may occur. The example shows how the boundary conditions can affect our results and how it is possible to find a steady state solution which is implausible, despite being mathematically correct.

[58] Furthermore, this type of numerical solver can only be used for a wind field which is strictly divergence free. When transposing wind fields from ARPS to the finite element snow transport model, some necessary interpolations can produce a wind field that does not meet this condition. This can cause discrepancies in the modeled temperature, snow concentration and humidity fields similar to those described in the previous example. If we consider a similar case, but this time with a wind field that is not strictly mass conserving, without a source we obtain $q_1 = q_2 = q_3$, similar to the first case. If the source is 0.15 g kg^{-1} and the fixed boundary values are $q_1 = q_3$, we find $q_2 = 1.1$ g kg^{-1}:

$$q_1 = 1.2 \xrightarrow{u_a=5} \underset{S=+0.15}{q_2 = 1.1} \xrightarrow{u_b=10} q_3 = 1.2.$$

This simple example shows how problems arise due to the limitations of the boundary conditions, to the assumption of steady state and to the wind field not being divergence free. In Alpine3D this is much more complicated since we have to consider three dimensions and diffusion as well, but these examples serve to illustrate some of the numerical problems that may occasionally occur when using the solver of Alpine3D.

[59] An example where this effect occurs in our terrain can be seen in the vertical mean of specific humidity in Figure 5, where in the NE there are some plumes with relatively dry air at the starting point. This effect is also found with suspended snow concentrations, where concentrations increase when sublimation and wind gradients are strong at the start of the plume and decrease elsewhere, but this happens very sporadically. As this effect will not occur when the feedbacks are neglected, we compared the steady sublimation to the sublimation of simulations without feedbacks. We found that the vertical mean of steady sublimation is smaller than the initial sublimation, which implies that sublimation is still reduced by the feedbacks and the effect of this unrealistic solution of the numerical solver on the snow distribution is probably very small.

[60] Apart from this effect, a slight change in the concentration field due to the extra iterations that are required because of the sublimation calculations can also cause a minimal increase in the suspended snow concentration. The numerical solver will start with a different initial field and can only solve the fields to a certain level of accuracy (0.1 mg for suspended snow concentration). Therefore, the final fields may be somewhat different from the reference calculation. This is again a local effect which merely causes a shift in deposition of approximately one grid cell and it does not affect the spatially averaged mass balance. The actual difference in the final SWE is shown in Figure A1, where shifts of deposition to neighboring grid cells are clearly

visible (e.g., on the saddle between Wannengrat and Chüpfenflue; Figure 1). As shown in section 4.4, we filtered this field with a low-pass filter to eliminate the changes in snow amount arising from shifts in deposition rather than sublimation.

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