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Seasonal simulation of drifting snow sublimation in Alpine terrain

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[1] We estimate seasonal drifting snow sublimation at a study site in the Swiss Alps with the numerical model Alpine3D using external wind fields from the Advanced Regional Prediction System on a high-resolution grid (10 m). Novel in the field of snow transport modeling, the transport module of Alpine3D accounts for the feedbacks among drifting snow sublimation, snow concentration, temperature, and humidity of the air in three dimensions. Due to these feedbacks, drifting snow sublimation is a self-limiting process. Model results show that the domain averaged drifting snow sublimation over a season is small (about 0.1% of precipitation) but spatially highly variable. Simulation results show strongest seasonal reduction of snow amount by 1.8% due to drifting snow sublimation in a leeward slope during SE wind. This can be explained by the generally warmer and dryer conditions during events with SE wind. In the Wannengrat study area, which covers typical alpine terrain, drifting snow sublimation is thus only significant locally or on short time scales. Note that we only consider drifting and blowing snow in the absence of concurrent precipitation. Furthermore, our results show that drifting snow sublimation is much smaller than surface sublimation in this area.

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1. Introduction

[2] Drifting snow causes erosion and deposition of large snow amounts within short time periods and changes the properties of the snow. During transport, the small snow particles with their relatively large surface are exposed to a turbulent environment where much snow may be lost to sublimation. Drifting snow thus has a major influence on the alpine snow cover, particularly on the avalanche danger [e.g., *Schweizer et al*., 2003, *Schirmer et al*., 2009] and the amount and timing of snow melt [e.g., *Luce et al*., 1998].

[3] Previous model studies estimate snow sublimation amounts during drifting snow for different environments. For mountainous terrain the number of studies is limited, however, due to the complexity of the process and the limited accessibility of the terrain. With use of the Prairie Blowing Snow Model (PBSM) [*Pomeroy et al*., 1993], *Pomeroy and Gray* [1995] estimated blowing snow sublimation to remove 15%–41% of annual snowfall at four stations in the Canadian Prairies. *Pomeroy et al*. [1997] used a simplified version of the PBSM to model the snow cover of a 68 km^2 catchment in the low arctic of northwest Canada. According to this study 19.5% of annual snowfall sublimated from blowing snow. In their model, drifting snow sublimation becomes more important with the size of a fetch, meaning that sublimation is stronger over level plains. *MacDonald et al*. [2010] estimated the drifting snow sublimation for a transect in the Canadian Rocky Mountains with the PBSM. In their simulations 17%–19% of seasonal snowfall is lost to drifting snow sublimation.

[4] Based on simulations with SnowTran-3D [*Liston and Sturm*, 1998; *Liston et al*., 2007] at the Berchtesgarden National Park, *Strasser et al*. [2008] showed that drifting snow sublimation has a large spatial variability in complex terrain. They state that especially at crests much of the winter snowfall sublimates. Averaged over the complete domain about 4% of snowfall is lost to drifting snow sublimation. In a following study *Bernhardt et al*. [2012] included gravitational snow transport in a similar model setup. This reduced the amount of snow at ridges and lowered the estimate of drifting snow sublimation to a seasonal average over the total area of 1.6% of annual snowfall.

[5] Thus, there is a wide spread in the estimates of drifting snow sublimation within this limited number of publications. A key cause of this spread is the varying climate in the studied regions but also differences in snow transport models. A limitation of these studies is that the used models simplify the sublimation process. In a previous study we have presented a routine for drifting snow sublimation in Alpine3D [*Lehning et al*., 2006, 2008]. In contrast to other 3-D snow transport models, Alpine3D now includes the feedbacks of drifting snow sublimation on snow concentration, air temperature, and humidity [*Groot Zwaaftink et al*., 2011, hereafter referred to as *GZ11*]. The implemented feedbacks significantly limit drifting snow sublimation, resulting in a 2% difference in deposition reduction in a lee slope. In the current study we use this model to

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estimate the amount of drifting snow sublimation in an alpine catchment over the course of an entire season, which is a great challenge because of the associated computational effort. This study contains the results of this simulation and will show that sublimation is only locally significant on longer time scales at our study site. After an introduction to our study site and Alpine3D, we will start with a brief validation of the wind fields used as input for the simulation. In order to show that the snowdrift module of Alpine3D can adequately describe the snow cover not only on short time scales, such as the simulations by *Mott et al*. [2010], but also over the course of a season, we will compare the modeled snow depth to measurements (both spatial and time series) at our study site. We will show time series of area-averaged sublimation and the spatial variability by means of the sublimation summed over one season. The spatial variability will be analyzed with respect to the weather conditions, in particular the wind direction. Finally, we present a simple drifting snow sublimation parameterization for our small domain based on the domain-averaged values of wind speed and relative humidity over ice.

2. Study Site

[6] Our study site is the Wannengrat near Davos, Eastern Swiss Alps (Figure 1). The modeled area is about 2.4 km^2 and the altitude ranges from about 2100 to 2650 m above sea level (asl). The study site is characterized by steep slopes with varying expositions, as well as by flatter terrain. The wind therefore causes several kinds of snow deposition patterns, such as homogeneous lee-slope loading, cross-slope loading and cornices, as described by *Mott et al*. [2010]. The area is equipped with seven automatic weather stations (AWSs, WAN1–7) to provide us with the necessary input and validation data for model simulations. In winter 2008–2009 the hourly mean temperature at WAN3 was -2.8°C, the relative humidity 75%, the wind speed 2.2 m s^{-1}, the incoming shortwave radiation 150 W m^{-2} and the incoming longwave radiation 247 W m^{-2}.

[7] The seasonal snow distribution in the catchment was measured by airborne laser scanning (ALS). In this paper we will verify the results of Alpine3D with an ALS measurement conducted in spring 2009. The accuracy of ALS was determined by a comparison with tachymeter data by *Grünewald et al*. [2010]. They maintain that there was a mean deviation of about 5 cm and a standard deviation of 6 cm.

3. Methods

3.1. Model Description

[8] We use Alpine3D for the snow cover simulations. Alpine3D is an alpine snow surface processes model [*Lehning et al*., 2006]. It is based on the 1-D physical snow cover model SNOWPACK [*Lehning and Fierz*, 2008]. SNOWPACK describes snow stratigraphy, snowpack settlement, surface energy exchange, and mass balance of a seasonal snow cover. In Alpine3D, SNOWPACK is applied distributed over the domain and is coupled to a snow transport module based on a saltation module [*Doorschot and Lehning*, 2002; *Clifton and Lehning*, 2008] and a 3-D suspension module [*Lehning et al*., 2008], including a drifting snow sublimation routine [*GZ11*].

[9] Within the suspension module the snow particles are modeled as passive heavy tracers, i.e., particles do not influence the flow, follow the mean flow, but have a settling velocity. We assume that snow transport is steady over 1 h and calculate the snow concentration with an advection-diffusion equation. As the model allows for sublimation of drifting snow and the effect thereof on temperature, humidity, and snow concentration, we do not only calculate the snow concentration, but also the 3-D temperature and humidity field. Sublimation is represented by a sink in the snow concentration and temperature conservation equations and by a source in the humidity conservation equation according to

$$\frac{\partial}{\partial \mathbf{x}}\left(K\frac{\partial c}{\partial \mathbf{x}}\right) - \mathbf{u}_p\frac{\partial c}{\partial \mathbf{x}} = -\mathrm{S}. \quad (1)$$

$$\frac{\partial}{\partial \mathbf{x}}\left(K\frac{\partial q}{\partial \mathbf{x}}\right) - \mathbf{u}\frac{\partial q}{\partial \mathbf{x}} = \frac{S}{\rho_{\mathrm{air}}}. \quad (2)$$

$$\frac{\partial}{\partial \mathbf{x}}\left(K\frac{\partial \theta}{\partial \mathbf{x}}\right) - \mathbf{u}\frac{\partial \theta}{\partial \mathbf{x}} = -\frac{SL_s}{\rho_{\mathrm{air}}c_p}. \quad (3)$$

[10] Here c is the snow concentration, q is the specific humidity, θ is the potential temperature, ρ_{air} is the density of the air, c_p is the specific heat capacity, and L_s is the latent heat of sublimation. u and u_p denote the wind speed and particle velocity, respectively. K is the diffusivity coefficient (attained from the Advanced Regional Prediction System (ARPS)), and S is the sublimation rate. Calculation of the sublimation rate is based on the equation of *Thorpe and Mason* [1966] that describes the sublimation of a single ice sphere. We omit the influence of solar radiation. This calculation for a single particle is then applied to an ensemble of particles. This approach has principally been validated in a wind tunnel study [*Wever et al*., 2009]. The set of conservations equations (1)–(3) and S are solved iteratively until they converge. Please see *GZ11* for further explanation of the sublimation routine in Alpine3D.

[11] A more accurate mass balance scheme was introduced in Alpine3D compared to previous publications. For drifting snow calculations the older versions of the drift module only checked whether there is snow and if snow transport is possible (by means of a threshold friction velocity depending on snow properties). However, the model did not check whether there was enough snow available for transport to last for the full time step (1 h). Practically, this meant that when there was little snow left at some grid points, snow transport could be occasionally overestimated. This problem occurred mainly in simulations where snow was exposed to pure drift conditions over longer periods, and this model deficiency is now corrected.

3.2. Model Input and Initialization

[12] The input of Alpine3D contains modeled wind fields and measurements of air temperature, relative humidity, incoming shortwave and longwave radiation, and precipitation. The sources of the input data and the initialization in Alpine3D are all consistent with *GZ11*.

[13] Except for precipitation, all measurements were taken from the AWS WAN3, or occasionally from WAN2 if no data were available at WAN3. We used precipitation measurements from a heated rain gauge at the nearby flat field site Weissfluhjoch (WFJ, see *Marty and Meister* [2012] for an overview of long-term operational measurements) as no precipitation measurements were available at the study site. WFJ (see Figure 1) is located about 4 km NE from Wannengrat at a similar altitude (2540 m asl). *Wirz et al*. [2011] have compared the snow depth at WFJ in winter 2008/2009 to the mean snow depth in the Albertibach catchment. This catchment, as defined in their study, is an area mostly overlapping with the model domain considered in the current study and covers an altitude range of roughly 2300–2650 m asl. *Wirz et al*. [2011] showed that the mean snow depth in the Albertibach catchment during the accumulation period is about 15%–40% lower than the snow depth at WFJ. However, a comparison study over a period of 4 years showed that the WFJ rain gauge underestimates the local snow water equivalent (SWE) at WFJ by about 32%, mainly due to wind influence [*Egli et al*., 2009]. Therefore, we find that the (uncorrected) WFJ precipitation data are an adequate estimate for a long-term simulation of the Wannengrat study area. This will be further discussed in section 4.2 based on a comparison of measured and modeled snow depths at WAN3.

[14] Wind fields, used as input for Alpine3D, were calculated with ARPS [*Xue et al*., 2001]. As described by *Mott et al*. [2010] and *GZ11*, we use mean flow fields as an input for Alpine3D. Rather than simulating a new flow field for each time step, we generated a library of wind fields to limit computing time. We classify the prevailing wind direction, which is either SE or NW, based on the observations at WAN1–3 (Figure 1). For both wind directions we have a set of wind fields from simulations initialized with a varying velocity such that we cover the range of measured wind speed at WAN3. For the SE wind direction we simulated four different strengths of wind speed, for the NW direction seven. We thus have a library of 11 wind fields. Based on the hourly mean of the observed wind speed at WAN3 and wind direction at WAN1–3 we choose the most suitable wind field from this library for each simulated hour. This of course decreases the variability in wind compared to the measurements, as will be discussed in section 4.1.

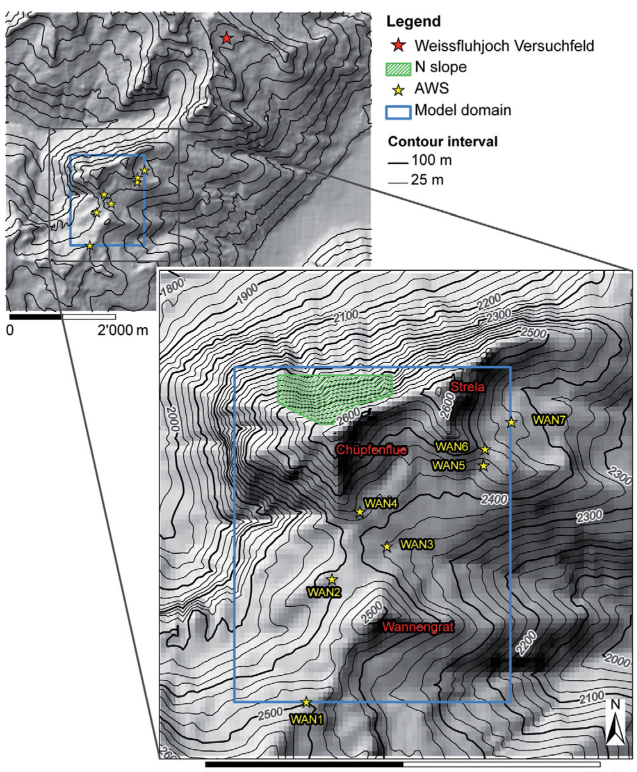

Figure 1. Map of study site Wannengrat and the surrounding area, including the WFJ Versuchsfeld. Names of mountains are labeled in red. The background shows the digital elevation model (*Source*: swisstopo) with standard hill shading.

[15] The initial temperature field has a dependency on elevation according to a dry-adiabatic lapse rate, consistent with our choice to only consider the sublimation of drifting and blowing snow during nonprecipitation (dry) periods. We assume that the specific humidity is initially constant throughout our domain.

3.3. Model Simulation

[16] We simulate winter 2009 (1 October 2008 to 28 June 2009) as this period has already been investigated by previous studies on snow depth variability [e.g., *Mott et al*., 2010; *Schirmer et al*., 2011; *Wirz et al*., 2011] and provides sufficient input and validation data. This winter was shown to have typical high wind speeds from NW and to be representative regarding snow climatology based on measurements of wind speed and snow depth at WFJ [*Schirmer et al*., 2011]. In our overall mass balance, drifting snow sublimation is included. We additionally calculate the erosion and deposition that would occur if drifting snow sublimation is not included in the mass balance. These erosion and deposition values are then used to calculate how deposition and erosion are changed due to sublimation. Drifting snow sublimation is only calculated when it is expected to be relevant, i.e., the wind is strong enough to initiate drifting snow and the air is not saturated in the regions where drifting snow occurs. This setup saves computing time and excludes all processes other than drifting snow sublimation (such as differences in saltation occurrence due to a threshold value of snow amount) that might cause differences between two simulations. The simulation is done in hourly time steps and with a horizontal resolution of 10 m. As we use terrain following coordinates the vertical resolution varies; the lowest grid level is at 0.7 m on ridges and increases to 1.1 m on flatter terrain.

4. Results and Discussion

4.1. Validation of Wind Fields

[17] The wind fields are important for snow transport simulations. We therefore evaluate the simulated wind fields using the measurements at the meteorological stations. The measured and modeled means and standard deviations of wind speed are given in Table 1. The observations and model estimates cover the period October 2008 to June 2009, except at WAN2 (March–June 2009) and WAN4

Table 1. Observed and Modeled Means of Wind Speed (m s^{-1}) and Standard Deviations of Wind Speed (m s^{-1}) at the AWS Indicated in Figure 1

	Mean Wind Speed (m s^{-1})		Standard Deviation of Wind Speed (m s^{-1})	
AWS	Obs	Model	Obs	Model
WAN1	3.1	3.9	2.5	1.7
WAN2	3.3	2.5	2.3	1.2
WAN3	2.2	2.6	1.7	1.4
WAN4	1.8	2.6	2.4	1.4
WAN5	2.1	2.1	1.5	1.0
WAN6	2.9	2.3	2.2	1.3
WAN7	1.8	2.1	1.3	1.0

(October 2008 to February 2009) where sensor failures occurred. The modeled estimates are interpolated values, based on the wind speed at wind field grid points surrounding the location of the AWS. Both observed and modeled wind speeds were scaled to a height of 2 m above the surface, assuming a logarithmic wind profile (roughness length is 1 cm). The modeled mean wind speed corresponds reasonably to the measured mean wind speed. The standard deviation or the variability of the wind is lower for the model estimates due to the use of 11 mean flow wind fields to cover the wind during one winter season.

[18] Wind roses of weather station observations and the simulated wind enable us to evaluate both the representations of the wind direction and wind speed (Figure 2).

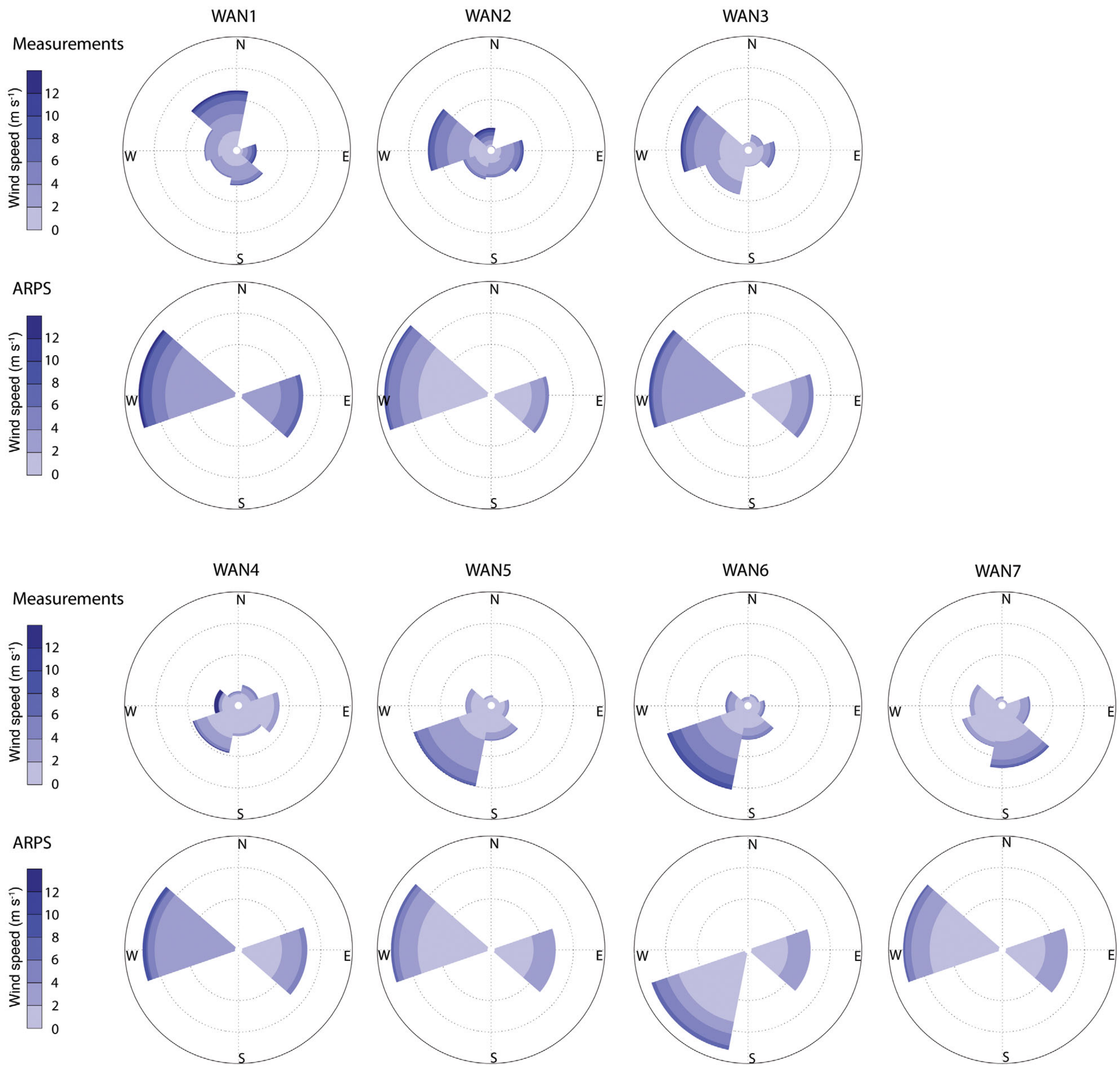

Figure 2. Wind roses of observations and ARPS simulations at WAN1–7. Colors indicate the hourly mean wind velocity (m s^{-1}), the frequency of occurrence is represented in percentage circles, the dotted inner circle represents <30%, the dotted mid circle represents <50%, and the filled outer circle represents <70%.

Figure 2 illustrates that the model cannot reproduce the complex wind patterns fully but approximately represents the dominant wind at each station. In the current model setup of ARPS [see *Mott et al*., 2010], the model is only able to represent topographically induced flow patterns of the mean flow. Thermal flow patterns, for example, are thus not captured by the model. Also, the model is not able to fully reproduce the flow separation observed at WAN4 and WAN7 (Figure 2). Therefore, both wind direction and wind speed at WAN4 and WAN7 diverge from the observations. At WAN5 the wind direction is not represented well. According to our observations, the wind is deflected around the SE ridge of Strela (see Figure 1); this results in a SW wind at WAN5 and WAN6. In the simulated wind fields the wind turns somewhat too far downstream. At WAN5 the model therefore still predicts west wind, but at WAN6 the wind has turned and ARPS represents the prevailing wind direction quite well. Periods with very weak winds are difficult to model as is seen at stations WAN1, WAN3, and WAN4, where ARPS overestimates low wind speeds. However, once the modeled wind speed is weak enough to not exceed the threshold friction velocity, the effects on drifting snow and sublimation thereof are limited. Consequences are more likely to be seen in, for instance, the estimates of turbulent heat fluxes from the surface, which also affects our results as discussed later.

4.2. Snow Depth

[19] *Mott et al*. [2010] showed that Alpine3D can provide an accurate representation of the changes in snow distribution due to snow transport during storm periods lasting up to 2 weeks at Wannengrat. Before looking at estimates of sublimation amounts during a season, we want to confirm that Alpine3D is able to give an adequate representation of the snow cover on a longer time scale of one winter season. We therefore compare the modeled snow depth around the time of peak accumulation to measurements obtained by ALS (Figure 3).

[20] Alpine3D is able to capture some of the main drift patterns in this area, as we will demonstrate with a selection of patterns indicated by the arrows in Figure 3. For example, the location of snowdrifts in relatively smooth terrain (Figure 3, arrow 1) and cornices in steep terrain (arrow 2) is represented properly. Also, the observed lee-slope loading of snow (arrow 3) is seen in the model estimates. Alpine3D captures the persistent cornice-like drifts in the cross-loaded NE slope of Wannengrat (arrow 4) and the snow accumulation near WAN5 (arrow 5). Alpine3D cannot capture the decreased snow depth in avalanche release zones (arrow 7), as avalanches are not included in the model. Furthermore, the resolution of the digital elevation model restricts the spatial variability. Small features in the snow cover such as the filling of narrow gullies (arrows 9 and 10) are not present in the simulation results due to the resolution of the wind field. Finally, especially the patterns that are related to steep slopes are mostly exaggerated. The model overpredicts the extent and depth of cornices (arrows 6 and 8). These cornices become unrealistically large due to a speed-up of the wind, as described by *Mott et al*. [2010]. The speed-up of the wind thus causes errors that accumulate during the season in restricted areas.

[21] Besides the drift patterns we also study the snow depth distributions by means of box plots (Figure 4). We measure the central tendency using the median, a robust measure, since the model largely overestimates the snow depths in confined areas. The median values are 1.39 (observations) and 1.79 m (Alpine3D). Alpine3D underestimates the spatial variability in snow depth as can be inferred from the range

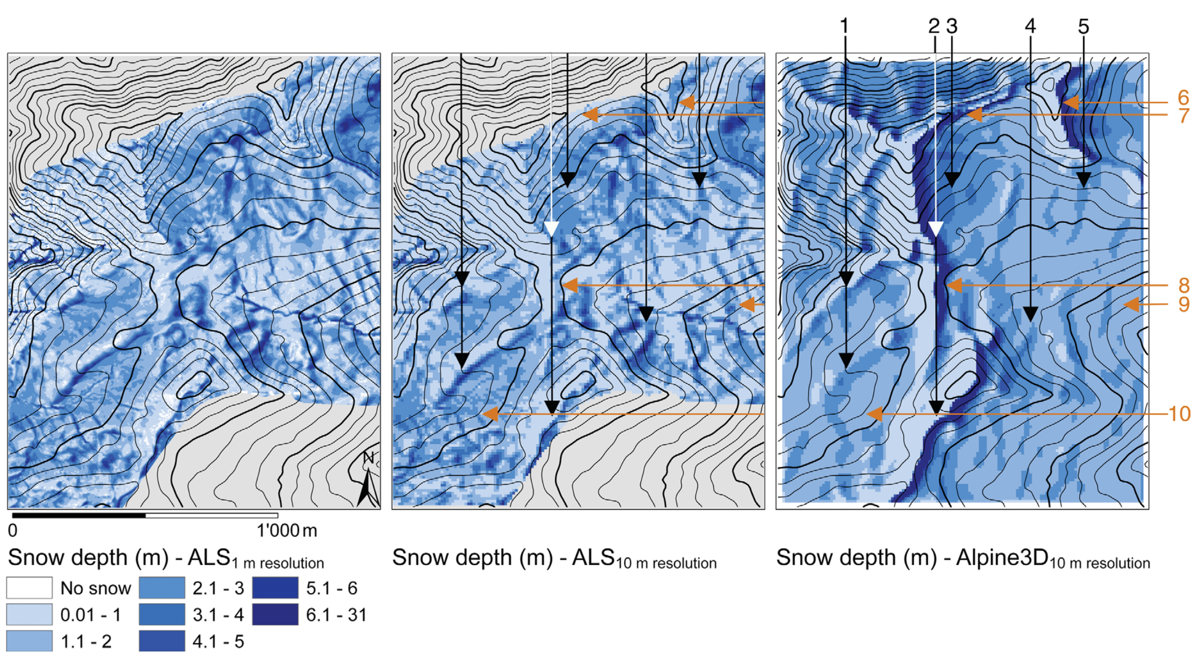

Figure 3. Observed (1 m resolution and averaged to 10 m resolution) and simulated (10 m resolution) snow depth, 9 April 2009. The lines show the topography: the intervals between the thin lines represent 25 m and between the thick lines 100 m. The arrows indicate examples of snowdrift patterns discussed in section 4.2.

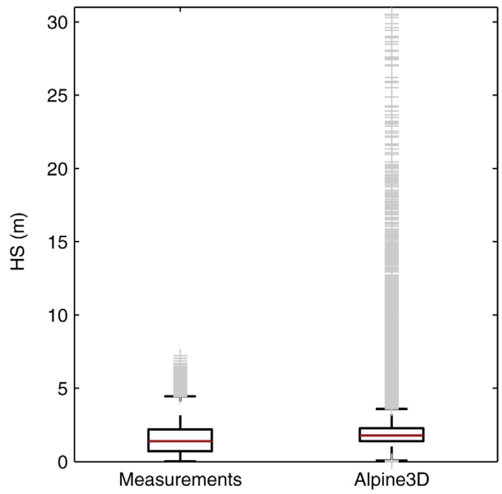

Figure 4. Snow depth (m) as measured with ALS and estimated by Alpine3D on 9 April 2009. Boxes span the interquantile range from first (Q1) to third (Q3) quartile. The red horizontal line shows the median value. The whiskers show the range of observed values that fall within 1.5 times the interquantile range above Q3 and below Q1. Values outside this range are shown by gray crosses.

between the first and third quartiles (Figure 4). This range is smaller in the simulation results than in the measurements. Likely causes of this underestimation of variability are (i) the use of steady state wind fields, (ii) the limited number of wind fields, (iii) the assumption of hourly steady state snow transport, and (iv) the resolution of the simulation.

[22] Apart from the spatial distribution, we also compare the snow depth estimates to measurements from all AWS within our model domain equipped with a sonic snow depth sensor during the simulation period (Figure 5). Especially at WAN3, snow depth of Alpine3D estimates correspond well to the measurements. WAN2 and WAN6 are located on windward sides of snowdrifts. In Alpine3D, however, the snowdrifts extend up to the AWS location, and snow depth is therefore overestimated. WAN4 is located in a cornice, and measurements only cover the period until it is

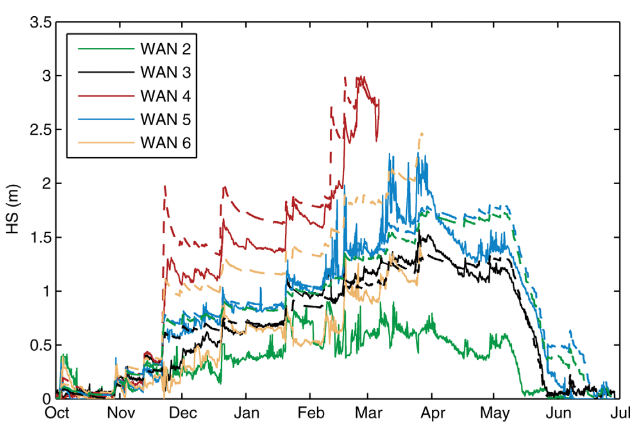

Figure 5. Snow depth (m) as measured and modeled at AWS WAN2–6 (see Figure 1) during winter season 2008/2009. Solid lines represent measurements and dashed lines represent model estimates.

buried in snow, and WAN5 is positioned at a wind exposed site. Even though at both locations ARPS showed difficulties to capture the wind speed and direction (see section 4.1), the modeled snow depth corresponds reasonable to the measurements.

[23] Generally, the snow depth is overestimated. Several aspects may be responsible for this overestimation. One possible reason is an overestimation of precipitation amounts. However, the good correspondence between measured and modeled snow depths at the relatively flat field area of AWS WAN3 (a site dominated by precipitation rather than by drifting snow) indicates that precipitation used for this simulation was adequate. A second reason for overestimation of snow depth is the underestimation of density by the model. This is a likely reason since the settlement routines have been developed for flat field sites with limited influence of drifting snow. From a comparison to eight snow profiles (measured in the period January–May 2009; not shown), we conclude that Alpine3D underestimates the averaged snow density of the snowpack by about 10%. This causes an overestimation of snow depth. A third reason can be an underestimation of drifting snow sublimation. Plausible factors causing underestimation of sublimation may be underestimation of dry-air entrainment through errors in the vertical moisture profile and neglecting solar radiation in the mass exchange between a particle and its surrounding air. Underestimation of drifting snow sublimation, however, would cause a local rather than a general overestimation of snow depth and is therefore not supported by our results. Moreover, due to overestimated snow transport at the ridges we expect that we overestimate rather than underestimate sublimation.

[24] An accurate representation of drifting snow is, apart from other factors, essential to a reliable estimate of drifting snow sublimation. The results presented here suggest that Alpine3D can represent drifting snow during a season over large areas. At the same time we expect that drifting snow sublimation is overestimated in small regions where snow transport is overestimated. We expect that these errors are of minor importance compared to other improvements of Alpine3D, in particular the high resolution of this simulation compared to previous studies and the inclusion of thermodynamic feedbacks of sublimation. We thus assume that Alpine3D is able to give a reliable estimate of seasonal drifting snow sublimation, and we note potential for further improvement through advances in wind field modeling in steep terrain.

4.3. Drifting Snow and Surface Sublimation: Time Series

[25] *GZ11* estimated, based on simulations, that sublimation could reduce snow deposition by 2.3% (domain average) at Wannengrat during a Föhn storm. As a Föhn storm has advantageous conditions for sublimation, i.e., strong wind to bring snow in warm and dry air, we expect that the total effect over a season is smaller. In Figure 6 we show the cumulative mean reduction of deposition (black line) averaged over the domain. This is 0.8 mm water equivalent (mm w.e.) at the end of the season. Compared to the precipitation, 755.5 mm, observed at WFJ in the same period, this value is only 0.1% and thus negligible. This estimate is lower than a previous estimate for drifting snow sublimation

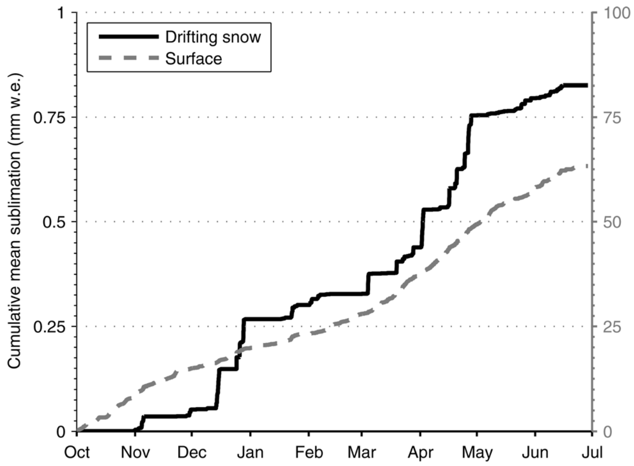

Figure 6. Cumulative mean reduction of SWE (mm w.e.) due to sublimation of drifting snow (black, solid) and from the snow surface (gray, dashed) during the complete simulation.

at an alpine site given by *Bernhardt et al*. [2012]. Reasons for this will be discussed in section 4.4.

[26] Other than the total seasonal effect, the plot shows that most drifting snow is sublimated in short storm periods of several days. During such storms deposition in the north slope of Chüpfenflue (further referred to as N slope and indicated in Figure 1) was reduced by up to 25%. In spring time, especially March and April, there are more periods with strong drifting snow sublimation. Their effect compared to winter snowfall, however, remains negligible.

[27] *GZ11* compared drifting snow sublimation to surface sublimation and model results showed that these are on the same order of magnitude. Therefore, we also show the cumulative surface sublimation in Figure 6. We selected only sublimation (no deposition) and included evaporation during snow melt. Drifting snow sublimation is much smaller than surface sublimation. Besides in the order of magnitude, we also see a difference in the timing of sublimation events. Unlike drifting snow sublimation that occurs in short intervals only, surface sublimation is more evenly distributed over the season. For both processes dry and warm air are needed. In case of surface sublimation, strong wind is favorable as it increases the exchange, but for drifting snow sublimation strong winds are a prerequisite. Drifting snow sublimation therefore only occurred in specific events, whereas surface sublimation occurs more frequently. However, also for surface sublimation we can distinguish periods of relatively weak sublimation (December, January) and stronger sublimation (March, April). These periods may be related to air temperature and the energy available from shortwave radiation.

4.4. Drifting Snow and Surface Sublimation: Spatial Distribution

[28] Drifting snow sublimation is not only limited to occasional storm periods, but also restricted to certain areas. This can be inferred from the reduction of snow amount due to sublimation summed over a season (Figure 7). The model results in Figure 7 show that there is a difference in the pattern of drifting snow sublimation and surface sublimation. Significant effects of drifting snow sublimation (Figure 7, left) are limited to the N slope of

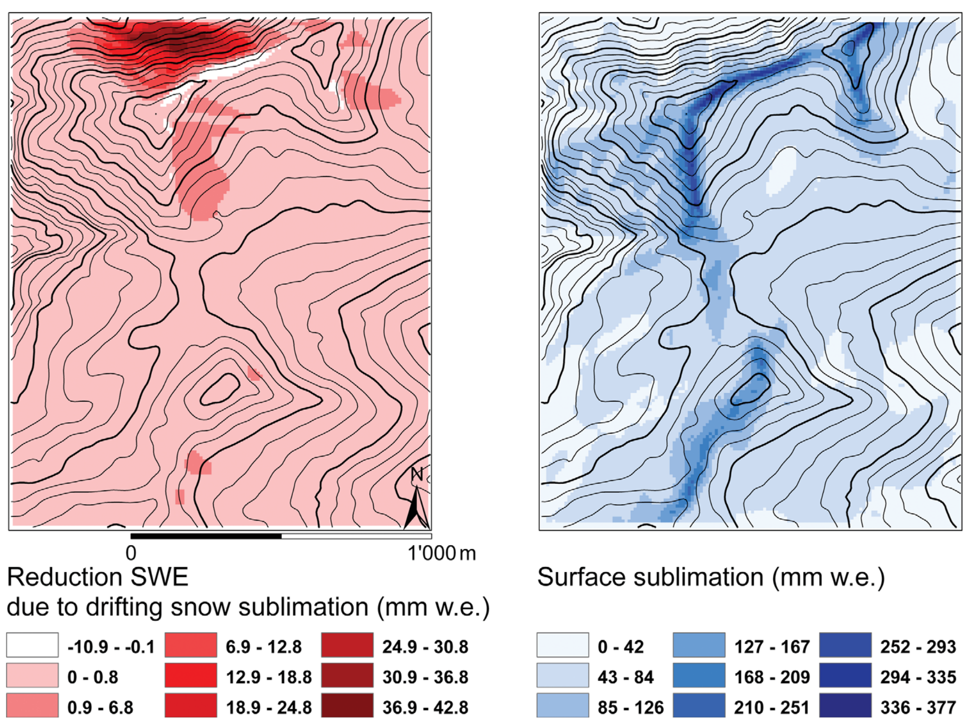

Figure 7. Seasonal reduction of snow amount due to (left) drifting snow sublimation and (right) surface sublimation in mm w.e. The lines show the topography: the intervals between the thin lines represent 25 m and between the thick lines 100 m. The color scale in the left panel is such that all above-averaged values stand out.

Chüpfenflue (see Figure 1) and the east slopes of Chüpfenflue and Strela, whereas surface sublimation (Figure 7, right) occurs throughout the model domain. This can be explained as surface sublimation can occur at every site where snow is present. Drifting snow sublimation however can only occur in plumes of drifting snow. Its effect can consequently only be seen in the deposition zones of these plumes. The consideration that surface sublimation affects a much larger area than drifting snow sublimation helps us to further explain the difference in the order of magnitude of the mean reduction of snow amount at the end of the season discussed in section 4.3.

[29] Apart from a difference in occurrence of the two processes, there is also a difference in their local relevance. Whereas surface sublimation shows peak values at ridges and no clear dependence on exposition, drifting snow sublimation is much larger in the N slope than in any other drifting snow deposition zone. In the N slope, drifting snow sublimation reduced the snow amount with values up to 42 mm w.e. and an average reduction of 15.9 mm w.e. Note that we concluded in section 4.2 that drifting snow sublimation estimates in such slopes might be overestimated due to an overestimation of snow transport at steep ridges. The sublimation effect relative to the snow amount in this slope may therefore be of more value. Compared to the amount of snow at the time of peak accumulation, which is 888 mm w.e. averaged over this slope, 1.8% is lost to drifting snow sublimation. An explanation for the relatively large effect of drifting snow sublimation in this particular slope could be that there are more occasions where snow is transported to the N slope than to other slopes. Such increased transport is possible if SE winds are prevailing or are stronger than NW winds. The wind rose of WAN3 (Figure 2) however shows the contrary, i.e., NW winds are prevailing and generally stronger. Consequently, we can conclude that even though SE winds occur less frequently, they may be more effective for sublimation of drifting snow. This is possible due to the dependency of sublimation on the relative humidity and temperature of the surrounding air. During periods with strong SE winds the relative humidity is lower and temperature is higher than during periods with strong NW winds, as we can infer from Table 2. SE winds thus bring favorable conditions for drifting snow sublimation and therefore cause an enhanced reduction of snow amount due to drifting snow sublimation in the N slope, relative to other deposition zones.

[30] The peak values of surface sublimation at ridges, on the other hand, are due to the locally very strong winds. As discussed by *Mott et al*. [2010], the wind is overestimated at the ridges. Therefore, the estimates of surface sublimation at the ridges are probably also overestimated.

Table 2. Mean Values of Air Temperature and Humidity at WAN3 During the Complete Winter, During Periods With Strong SE Wind and During Periods With Strong NW Wind

	T (°C)	RH (%)
Season	−2.8	75
SE wind >4.5 m s^{-1}	−2.4	76
NW wind >4.5 m s^{-1}	−7.5	87

[31] The estimated large spatial variability of drifting snow sublimation and its small seasonal influence on the snow distribution compared to snowfall (<0.1%) in this model simulation suggest that in this catchment or regions with a similar climate and topography, drifting snow sublimation needs only to be accounted for on short time scales or if the spatial variability of the snow cover is of main interest. Compared to the studies addressed in section 1 Alpine3D gives a very low estimate of drifting snow sublimation at Wannengrat even though the maximum local value of hourly sublimation rate (0.58 mm w.e. h^{-1}) is larger than the values reported by *Déry and Yau* [2001] (0.16 mm h^{-1}) and *Bernhardt et al*. [2012] (0.13 mm h^{-1}) but is on the same order of magnitude. Explanations for the large difference in sublimation estimates can be found both in differences in model setup and in the climate of the studied regions. First of all, a reason for the low mean sublimation is the inclusion of the feedbacks in Alpine3D [*GZ11*]. The feedbacks on snow concentration, humidity, and temperature have a limiting effect on sublimation of drifting snow. In the case study of *GZ11*, the reduction of deposition due to drifting snow sublimation in a lee slope changed by 2% due to these feedbacks. The inclusion of the feedbacks can thus partly explain why the current drifting snow sublimation estimates at Wannengrat are lower than other sublimation estimates. Another plausible explanation is the resolution of the simulation, as other studies were based on models with a coarser resolution. This effect needs to be investigated in more detail in the future, however. These differences in model setup are likely to explain the current low value of 0.1% sublimation compared to the estimates of *Bernhardt et al*. [2012] for an alpine site (1.6%). Comparing to studies such as *Pomeroy et al*. [1997] and *MacDonald et al*. [2010] on the other hand, the difference in climate and topography between the test sites is likely to be the most important factor, albeit model differences are larger yet difficult to quantify without a careful comparison study. These studies were done for catchments that differ greatly from Wannengrat, as far as we could infer from the climate data given in the respective studies. *Pomeroy and Gray* [1995] studied a site in the Canadian Prairies with relatively strong winds (mean velocity of 4.5–6.6 m s^{-1}) and probably more snow transport. The mean relative humidity recorded at the study site located in the Canadian Rocky Mountains [*MacDonald et al*., 2010] was lower (seasonal mean of 67% compared to 75% at Wannengrat). A lower relative humidity suggests that drifting snow sublimation could be larger. Finally, at the site of *Pomeroy et al*. [1997], relative humidity was somewhat larger than at Wannengrat, but the winds were much stronger (monthly mean 10 m wind speeds ranging from 4.2 to 7 m s^{-1} compared to 1.7–4 m s^{-1} at Wannengrat). Snow transport was thus likely more frequent and relevant. Furthermore, the terrain considered is described as level and consisting of larger fetches. This may be relevant as *Pomeroy et al*. [1997] argue that sublimation becomes more important over longer fetches. At Wannengrat, fetches are small, snow plumes do not extend far, and snow may reach the ground before significant sublimation occurs. A better comparison of the sublimation estimates from Alpine3D relative to other snow transport models may be reached by applying several models to the same study site, but this is out of the scope of the current study.

4.5. Parameterization of Drifting Snow Sublimation

[32] The drifting snow sublimation estimates discussed in the previous sections are a result of a time-consuming model simulation. Especially in cases where there is less interest in spatial variability on slope scales or smaller, a parameterization of the effect of sublimation on the snow mass balance could be valuable. *Pomeroy et al.* [1997] gave a parameterization to estimate drifting snow sublimation. They related monthly sublimation amounts to monthly means of wind speed, daily minimum air temperature, daily minimum relative humidity, snowfall, and snow amount. However, this parameterization was made for source areas (1000 m fetch) of drifting snow in level terrain and is not applicable to complex terrain such as Wannengrat.

[33] In our attempt to make a parameterization we used multiple linear regressions. First, we identified the main parameters to be the wind speed (at 2 m) and the relative humidity over ice, both averaged over the domain. We then made a linear regression between the parameters and the hourly (domain-averaged) reduction of deposition due to sublimation calculated with Alpine3D. We found a parameterization of the form $\sqrt{\Delta \mathrm{SWE}_{\mathrm{subl}}} = 5.24 \times 10^{-4}\overline{u}^3 - 4.64 \times 10^{-6}\mathrm{RH}_i\overline{u}^3$. A comparison between the reduction of SWE calculated with Alpine3D and estimated with this parameterization is given in Figure 8. This parameterization could describe the reduction of sublimation with a root-mean-square error (RMSE) of 8.3×10^{-4} and a coefficient of determination (R^2) of only 0.30. Even though temperature also affects sublimation, the inclusion of temperature in the parameterization could not improve its performance, possibly because temperature already has an indirect effect through the relative humidity.

[34] In Figure 8 we can distinguish two groups with a different behavior. We found the main difference between these groups to be the prevailing wind direction, suggesting that a parameterization may only be valid for a subset of specific conditions investigated. Therefore, in the insets in Figure 8, we also show the results of a similar parameterization but based only on data from periods with either SE or NW wind. The regression coefficients for these subsets of NW winds (2.86×10^{-4}, -2.7×10^{-6}) and SE winds (11×10^{-4}, -9.7×10^{-6}) vary strongly. The estimates of deposition reduction based on these parameterizations are much more adequate, as can be inferred from the RMSE and R^2 values for both SE winds (5.3×10^{-4}, 0.82) and NW winds (1.99×10^{-4}, 0.96). The results in Figure 8 show that such a parameterization can potentially give a good estimate of sublimation, while the large difference in regression coefficients indicates that a simple parameterization is only valid for very specific conditions. The current parameterization is thus not applicable in general and merely a first step toward a simple drifting snow sublimation parameterization for alpine terrain. The parameterization is, however, a confirmation of the most significant parameters, namely, wind speed and relative humidity. A problem for retrieving a more general parameterization could be the rather small domain in which single drifts have a too large effect on the average sublimation. A more consistent parameterization might be retrieved based on the simulations of a larger area where all wind directions influence a similar sized region and a single slope will not affect the domain mean value, but this remains to be shown in the future. This may also be the reason why we failed to express drifting snow sublimation simply as a fraction of surface sublimation, which might work for larger areas (not shown).

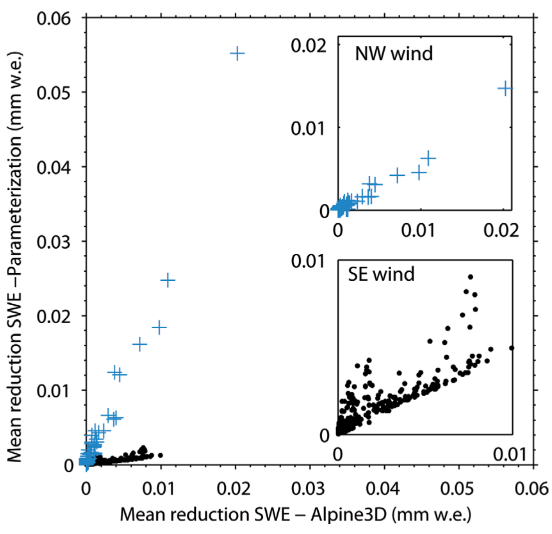

Figure 8. Mean reduction of snow amount (mm w.e.) due to drifting snow sublimation according to Alpine3D and the simple parameterization discussed in section 4.5. Symbols indicate the prevailing wind direction which is either NW (blue cross) or SE (black dot). Insets: same as main figure but only for NW wind or SE wind.

5. Conclusion

[35] The snow surface process model Alpine3D was used to simulate the seasonal snow cover in a small alpine catchment with particular aim to quantify the amount of drifting snow sublimation. We showed that we are able to reasonably represent the mean flow field over a season using a library of wind fields obtained with ARPS, especially considering the steep terrain we are dealing with. The use of mean flow fields generally results in a satisfactory snow depth distribution capturing most of the snowdrifts measured by ALS. Especially in steep terrain, however, potential for improvement is recognized as snow drifts and cornices, and therefore probably also drifting snow sublimation, are overestimated.

[36] Model results showed that drifting snow sublimation is particularly strong in single storms and in spring months. Furthermore, drifting snow sublimation is shown to have a large spatial variability at our study site. Strongest reduction of deposition due to sublimation is estimated in the leeward slopes during SE wind, as these flows tend to be warmer and drier. Furthermore, snowfall usually occurs with NW wind, leading to saltation and preferential deposition of precipitation. This means that after snowfall, there is already less snow available for transport in the windward slopes of the prevailing wind direction NW.

[37] Averaged over the entire domain and considering the complete winter season, drifting snow sublimation sums up to about 0.1% of snowfall and thus appears to be negligible at this site. However, in case the snow distribution on smaller scales is of interest, drifting snow

sublimation may not be negligible. In the slope most affected by this process (N slope, Figure 1), drifting sublimation reduces the amount of snow by 1.8%.

[38] We lack measurements to confirm our estimate as there are no techniques to measure drifting snow sublimation directly and independently of other processes. Our estimate is lower than other estimates for alpine terrain, such as *Strasser et al*. [2008] and *Bernhardt et al*. [2012], and much smaller than estimates given for other regions [e.g., *MacDonald et al*., 2010]. The feedbacks among sublimation, snow concentration, humidity, and temperature of the air limit the sublimation process and are not accounted for in similar model studies. We therefore expect a lower estimate with Alpine3D than with other 3-D snow transport models. However, there are other factors that likely have a larger influence on drifting snow sublimation estimates. First, the resolution of model studies influences snow transport amounts and therefore sublimation amounts. Second, the local climate is crucial as may also be inferred from the large influence of SE or NW wind and the accompanying conditions at the current study site. At Wannengrat, the air is relatively moist compared to other study sites, for instance, the Canadian Rocky Mountains [*MacDonald et al*., 2010], and the wind is weaker than at the study site of *Pomeroy et al*. [1997]. The latter suggests that snow transport may also be weaker, although other factors such as snow properties should also be accounted for. If snow transport is small relative to snowfall, drifting snow sublimation must also be small. The conclusion that drifting snow sublimation appears to be negligible over a season therefore only applies to the studied field site or sites with a comparable climate.

[39] Key issues that need to be improved to enable even more accurate sublimation estimates include the wind field simulations. In the current simulations, ARPS overestimates the wind at steep ridges, and the highly variable wind is limited to a library of wind fields. Moreover, alternatives to the representation of drifting snow as hourly steady state need to be explored as, especially in complex terrain, drifting snow is of highly turbulent nature.

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