Meteorological Modeling of Very High-Resolution Wind Fields and Snow Deposition for Mountains

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ABSTRACT

The inhomogeneous snow distribution found in alpine terrain is the result of wind and precipitation interacting with the snow surface. During major snowfall events, preferential deposition of snow and transport of previously deposited snow often takes place simultaneously. Both processes, however, are driven by the local wind field, which is influenced by the local topography. In this study, the meteorological model Advanced Regional Prediction System (ARPS) was used to compute mean flow fields of 50-m, 25-m-, 10-m-, and 5-m grid spacing to investigate snow deposition patterns resulting from two snowfall events on a mountain ridge in the Swiss Alps. Only the initial adaptation of the flow field to the topography is calculated with artificial boundary conditions. The flow fields then drive the snow deposition and transport module of Alpine3D, a model of mountain surface processes. The authors compare the simulations with partly new measurements of snow deposition on the Gaudergrat ridge. On the basis of these four grid resolutions, it was possible to investigate the effects of numerical resolution in the calculation of wind fields and in the calculation of the associated snow deposition. The most realistic wind field and deposition patterns were obtained with the highest resolution of 5 m. These high-resolution simulations confirm the earlier hypothesis that preferential deposition is active at the ridge scale and true redistribution—mainly via saltation—forms smaller-scale deposition patterns, such as dunes and cornices.

1. Introduction

Snow transport by the wind is a dominant factor influencing the temporal and spatial distribution of snow cover during winter. Especially in very complex terrain, the inhomogeneous snow distribution is the result of wind and precipitation interacting with the snow cover and the local energy balance at the snow surface. The investigation of the variability of the alpine snow cover and snow cover stability on different scales are the main challenges in avalanche forecasting (Schweizer and Kronholm 2007). Furthermore, an inhomogeneous snow cover leads to large uncertainties in estimating snow water storage and therefore the hydrological resources of high alpine catchments (Essery et al. 1999; Hartman et al. 1999; Liston and Sturm 2002; Mernild et al. 2006; Lehning et al. 2008; Michlmayr et al. 2008).

The flow in the atmospheric boundary layer is the driving force responsible for the redistribution of snow and the spatial distribution of precipitation. It is well understood that topographic features induce characteristic wind flow patterns on a local scale, such as crest speedup, flow channeling, flow blocking, updraft and downdraft zones, or flow separation downwind of a ridge crest (Lewis et al. 2008). The modification of the near-surface wind field results in preferred deposition of precipitation in the leeward slopes of mountain ridges, discussed in detail by Lehning et al. (2008). In the case of snow, preferential deposition is defined as the spatially variable deposition of precipitation due to topography-induced near-ground modification of flow field in the absence of local erosion and saltation. The effect of preferential deposition of precipitation on leeward slopes is attributed to reduced deposition velocities for snow on the windward slope, due to higher wind velocities and updraft, and increased deposition velocities leeward of ridges.

If snowfall events are combined with strong wind conditions, the already deposited snow is eroded and redistributed by saltation and suspension processes, causing an

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even more spatially variable snow depth distribution. When snow particles are transported in saltation, the snow grains follow ballistic trajectories within a thin layer close to the ground where particle concentration is highest (Bagnold 1941; Takeuchi 1980). The length scales of the transport distances of saltating grains are considerably smaller than of snow grains transported within suspension. Suspended snow grains are transported over long distances in a two-phase and well-mixed turbulent flow of snow and air (Essery et al. 1999; Gauer 2001; Essery and Pomeroy 2004).

Multiple model approaches exist that try to physically describe saltation and suspension as the main mechanisms for moving snow (Mellor 1965; Male 1980; Schmidt 1986; Pomeroy and Gray 1990). Winstral and Marks (2002) and Winstral et al. (2002) characterize the effect of wind on snow distribution using a series of terrain-based parameters. In previous studies much effort was carried out in trying to reduce the complexity of snow transport by wind to empirical and analytical relationships between topographical parameters, wind speed, and transport rates of snow (Tabler 1975; Uematsu et al. 1991; Liston et al. 2007). More complex models solve complex 3D wind fields with atmospheric models over high-resolution grids (Gauer 2001; Liston et al. 2007; Lehning et al. 2008; Bernhardt et al. 2009).

Atmospheric models are limited by their ability to represent the complex processes in a computationally efficient manner, especially on fine spatial scales (Liston and Elder 2006; Marks et al. 2008). However, the most sensitive parameter for simulating the small-scale distribution of snow is the driving wind field (Lehning et al. 2000; Essery 2001; Dacic et al. 2010). The spatial variability of winds over mountain landscapes produces high spatial variability in mass and energy fluxes. Raderschall et al. (2008) demonstrated that the atmospheric model Advanced Regional Prediction System (ARPS) successfully reproduces characteristic flow features such as the crest separation and flow blocking found in complex terrain. Atmospheric mean flow fields were successfully used for snow-drift modeling in complex terrain (Lehning et al. 2008; Mott et al. 2008; Bernhardt et al. 2009; Dacic et al. 2010).

Liston (2004) suggests that the spatial scales required to represent the driving mechanisms of snow transport should be 200 m or less. Especially in very complex terrain, this scale appears to be too large, although Bernhardt et al. (2009) use a resolution of 200 m to calculate drift and sublimation over very steep terrain. Other studies showed that a horizontal resolution of 25 m is sufficient to represent ridge-scale deposition features (Mott et al. 2008; Dacic et al. 2010). These studies showed that in glacier basins the main seasonal accumulation areas of snow were

found in areas of decreased horizontal wind velocities and in areas of increased downward surface-normal velocities. This is consistent with investigations that suggest the preferential deposition of precipitation is active at the ridge scale while the redistribution of snow by saltation and suspension is active on smaller scales (Lehning et al. 2008). That study pointed out that the small-scale spatial variability of the snow cover on the Gaudergrat ridge, measured after a major drift event in 1999, could not be captured by simulations with a horizontal resolution of 25 m. Raderschall et al. (2008) suggested that an even higher resolution than 25 m would be essential for drift simulations to resolve flow features caused by topographical features smaller than ridge size.

In this paper we address three main points concerning the application of microscale airflow simulations for quantifying snow transport and deposition processes. First, we discuss mean wind fields and embedded flow patterns on different scales since this has never been addressed explicitly. We discuss and show the effect of the numerical resolution on flow features, such as speedup effects at mountain crests. Second, we investigate the effect of the representation of characteristic flow features on the formation of observed snow deposition features as well as on the overall variability of the snow cover. Here we show that the mean flow field, calculated by the simple adaptation of the flow field to the underlying topography, can explain major snowdrift zones observed and modeled at Gaudergrat ridge, Switzerland. Finally the contribution of preferential deposition and pure snow drift processes (saltation and suspension) are shown and discussed on various scales. Flow field and snowdrift simulations are initially carried out in the context of a sensitivity study. Then, to check the accuracy of modeled snow depth patterns gained from the sensitivity runs on different grid resolutions, modeled snow depth patterns are compared qualitatively to snow depth patterns observed after a snowstorm event in January 1999. In the last part of the paper, we model a major snowfall and drift event in October 2003 and compare the results to field measurements.

2. Observed flow features and patterns of snow deposition

A steep mountain ridge, the Gaudergrat ridge (Davos, Switzerland; Fig. 1), was used as a test site for the investigation of wind and snowdrift processes for many years (Föhn and Meister 1983; Gauer 2001; Raderschall et al. 2008; Lehning et al. 2008; Faure 2008; Lewis et al. 2008).

The ridge crest reaches a height of 150 m above the surrounding topography, rising to a southern peak of 2305 m MSL. The ridge is oriented from southwest to northeast. In the Gaudergrat area, cold fronts with

FIG. 1. Location of the wind field and snowdrift modeling at Gaudergrat ridge, southeast Switzerland.

precipitation and increased wind speeds generally come from the northwest.

In summer 2003, the Gaudergrat Experiment (GAUDEX) (Lewis et al. 2008) was conducted at Gaudergrat ridge. There were 33 measurement stations, consisting of 27 automatic weather stations and 6 turbulence stations. The aim of GAUDEX was to provide a detailed description of the flow patterns of the surface boundary layer for the verification of numerical simulations of flow patterns over steep terrain (Lewis et al. 2008). These measurements provide evidence of two main terrain-induced flow patterns that are significant for snowdrift investigations. First, independent of the flow direction into the domain, a pronounced cross-ridge flow is observed together with strong speedup effects at the mountain ridge. This cross-ridge flow was explained by a cross-ridge pressure gradient. Since the hill-induced flow separation is enhanced by steep slopes in the lee of sharp mountain crests, flow separation occurred downwind from the main ridge independent of the flow direction into the domain. Second, air flows parallel to the ridge at lower altitudes in the western and eastern valley were observed for southerly and northwesterly wind conditions. In the eastern valley higher wind speeds were measured. Full details of measured flow patterns at the Gaudergrat ridge are discussed in Lewis et al. (2008) and Raderschall et al. (2008).

The variability of the snow cover at Gaudergrat ridge was measured after two major snowdrift events in 1999 (Fig. 2) and in 2003 by manual measurements of the snow depths (Fig. 12). Snow depth measurements from the drift event in 1999 were made by manual snow probing along transect lines running across the ridge before and after a major drift event spanning 26–31 January 1999.

The spatial resolution of the measurements is between 2 and 15 m. For the 2003 drift event measurements were made with a coarser spatial resolution, but the measurements covered a larger area.

Variability of snow depth on the steep slopes of Gaudergrat ridge can be attributed to both larger-scale (tens to hundreds of meters) and smaller-scale (a few meters)

FIG. 2. Manual measurements of snow depths after a snowstorm event in 1999.

deposition features. The larger-scale patterns of snow deposition are formed by flow features caused by the topography at the ridge scale. These patterns of snow deposition are primarily caused by preferred snow loading on the leeward slopes of the mountain ridge and reduced snow deposition or even snow erosion on the windward slopes.

The smaller-scale snow deposition patterns are formed by local topographic features such as local depressions, local bumps such as moraines, and the local shape/sharpness of the mountain ridge crest. During the course of the winter enhanced snow accumulation is detected in relation to these topographical features, leading to local deposition features such as snow dunes, cornices, and filling of troughs. One of these small-scale deposition features is the snow dune observed on the windward slope of the main ridge of Gaudergrat (Fig. 2), which will be discussed in more detail in the following sections. Additionally, smaller-scale variability is also caused by the high local roughness of the terrain (rocks, cow paths, etc.), which is not present in the digital elevation model.

3. Methods and model setup

In this study all sensitivity analysis runs were carried out over the same geographical domain with horizontal grid resolutions of 50 m, 25 m, 10 m, and 5 m, but with the same initial and boundary conditions. Wind field sensitivity runs were carried out for a free-stream wind velocity of $u_{\infty} = 6 \text{ m s}^{-1}$ and a northwest inflow wind direction. For the drift sensitivity runs, we set a constant meteorological input with a precipitation rate of 2 mm h^{-1} and constant wind parameters. To discuss the individual transport processes, we analyzed both a time series of preferential deposition (Lehning et al. 2008) and the final snow distribution. The relative contribution of saltation/suspension versus preferential deposition was evaluated by switching off the saltation/suspension in the corresponding sensitivity cases, as described in Lehning et al. (2008). For all sensitivity runs, the snow volume of the ridge and the windward/leeward slopes were calculated and analyzed. Furthermore, snow depth distribution was calculated for the first snowfall event in 2003. For this run, the snow transport model was initiated with a total of 10 different wind fields corresponding to the wind direction and velocity conditions. The meteorological data used to force the Alpine3D simulations were collected from the automatic weather station Versuchsfeld Weissfluhjoch.

The initial adaptation of the flow field to the complex terrain of the Gaudergrat ridge was computed using the nonhydrostatic and compressible atmospheric model ARPS (Xue et al. 2000, 2004). We used the wind fields as

an input for Alpine3D so as to quantify the redistribution of snow via saltation and suspension as well as the preferential deposition of precipitation.

An overview of the characteristics of the ARPS model relevant for the present airflow simulations in complex environments is given in Raderschall et al. (2008). Next, a short description of the model setup, especially adjusted to the purpose of this study, is given. A block diagram of all model runs, model modules, and input/validation data is shown in Fig. 3.

Under the prevailing northwesterly wind, flow is perpendicular to the mountain ridge. The domain size of $1.7 \text{ km} \times 1.4 \text{ km}$ in the horizontal direction was chosen to be sufficiently accurate to account for the influence of the surrounding topography. The horizontal grid resolutions are $dx, dy = (50 \text{ m}, 25 \text{ m}, 10 \text{ m}, \text{ and } 5 \text{ m})$. Note that the original grid spacing of the digital elevation model was 10 m and that the 5-m resolution was resampled. The coordinate system is terrain following, and the finest vertical resolution is found close to the ground. In accordance with a constant aspect ratio, the vertical grid spacing dz of the first level aboveground, $nz = 1$, ranges between $dz = 6 \text{ m}$ for the coarsest grid resolution and $dz = 0.8 \text{ m}$ for the finest grid resolution. We initiated ARPS from a horizontally homogeneous field and fixed lateral boundary conditions since a full three-dimensional initialization would require a nesting approach. To assure the comparability of different model runs with different horizontal resolution and to minimize the computational effort, the ARPS model was initialized with an artificial sounding of the atmosphere, which describes a vertical profile of the atmosphere for a characteristic situation during snowfall events. The atmosphere was assumed to be stably stratified and nearly saturated. The horizontal wind was initialized with a standard logarithmic wind profile within the boundary layer (Raderschall et al. 2008). Above the boundary layer, a vertically constant free-stream velocity u_{∞} was defined. To account for a snow-covered surface, a constant aerodynamic roughness length of $z_0 = 0.01 \text{ m}$ was chosen in agreement with measurements over similar terrain (Doorschot et al. 2004; Stössel et al. 2010). For the lateral boundaries, periodic boundary conditions were applied. To compute the characteristics of the mean wind field, which is shaped by the topography, the wind field was calculated until the flow adapted itself to the topography. This first adaptation of the flow field to the topography was reached after a 30 s integration time. Since we were only interested in the mean flow features, this short integration time ensured that the flow was not dominated by turbulent structures, and the model therefore remained stable in the face of periodic boundary conditions. The results of flow adaptation for various integration times

FIG. 3. Overview of models used and input/validation data.

and boundary conditions are discussed in Raderschall et al. (2008).

The modeled wind fields were used as a three-dimensional input for the snow transport model of Alpine3D by assigning a value of wind velocity and wind direction at each grid point in the three-dimensional space for each time step. The Alpine3D model was developed to deal with alpine surface processes and their temporal and spatial variability (Lehning et al. 2006). In this study we are primarily interested in processes that lead to the spatial and temporal variability of the snow cover. A full description of the energy balance (Helbig et al. 2009) at the snow surface is therefore implemented in the model and coupled to the SNOWPACK module (Lehning and Fierz 2008) and the snow transport module (Lehning et al. 2008). The three-dimensional snow transport module calculates the redistribution of previously deposited snow through saltation (Clifton and Lehning 2008) and suspension. Furthermore, the model is able to capture the preferential deposition of precipitation. The model allows the process of preferential deposition of precipitation to be simulated separately from processes related to the erosion and transport of previously deposited snow.

To investigate the influence of numerical resolutions on snowdrift processes a sensitivity analysis was performed

using four different numerical resolutions. This sensitivity analysis was carried out by a resample of the original grid with $dx, dy = 10$ m to grid resolutions of 5 m, 25 m, and 50 m. Note that increasing the spatial resolution from 10 to 5 m does not lead to an increase in topographical

FIG. 4. Simulated wind direction for the first level above ground with applied horizontal grid spacing (dx, dy) of 10 m.

FIG. 5. Modeled distribution of 3D wind speed with applied horizontal grid spacing of (a) 5 m, (b) 10 m, (c) 25 m, and (d) 50 m.

information. The corresponding effect of grid resolution on the wind field is therefore only a numerical effect. A decrease in grid resolution from 10 m to 25 and 50 m, in contrast, led to a considerable loss in topographical information.

4. Results

a. Sensitivity runs

1) MODELED AIRFLOW ON FOUR DIFFERENT SCALES

The mean flow fields gained from sensitivity runs are shown in Fig. 4 with numerical grids of 50 m, 25 m, 10 m, and 5 m. The simulations captured the main

terrain-induced flow patterns measured during GAUDEX. In all simulation setups, that is, at all four resolutions, the pronounced ridge-normal flows, as well as the flow that developed in the eastern and western valleys with a roughly parallel orientation to the ridge axis, are represented in the airflow simulations (Fig. 4). The spatial distribution of wind velocities (u , v , and w components of the wind vector) are shown in Fig. 5 for all numerical resolutions. Three transects (1–3) are plotted in Fig. 6, crossing the main ridge roughly in the direction of the cross-ridge flow during northwest wind conditions. The locations of these transects are shown in Fig. 5a.

The spatial patterns of the three-dimensional wind speed were clearly influenced by the two main terrain-induced flow features and were captured, in principle, at

FIG. 6. Transects of 3D wind speeds and respective topography crossing the main ridge at the (a) northern and (b) southern end of the ridge and in the (c) middle part of the ridge.

all numerical resolutions. The locally driven cross-ridge flow caused crest speedup. The airflow that developed parallel to the crest line induced speedup effects at the cross-slope ridge crests. In finer resolution runs ($dx, dy = 5 \text{ m}/10 \text{ m}$) reduced wind speeds were simulated for local terrain depressions.

The effects of the terrain-induced flows were most pronounced in simulations with sufficiently high resolution to resolve flows at length scales smaller than the ridge scale. With an enhancement of grid resolution the crest speedup increased dramatically. The strongest increase in wind velocity at the ridge crest appeared when grid resolutions were increased from 50 to 25 m. The strong speedup effects led to strong gradients of wind velocities in the direction of the mean flow upwind and downwind of the crest line. The strongest increase in wind velocity was modeled where the slope gradient of the topography was highest (see transect 2, Fig. 6b). At transect 2, flow separation with very low wind speeds in the lee close to the ridge were computed for $dx, dy = 10 \text{ m}$ and 5 m (Figs. 5a,b, 6b). Additionally, reduced wind speeds were computed on the windward side of transect 2. This decrease in wind speed is strongest for $dx, dy = 10 \text{ m}$ and 5 m (Figs. 5a,b, 6b) and exists owing to a small topographical depression. Due to the

modification of the wind speed by the shape and steepness of the ridge, we assumed a strong influence of these terrain parameters on final snow depth patterns, as has been previously observed by Föhn and Meister (1983).

2) FLOW FEATURES AND SNOW DEPOSITION

The results of drift sensitivity runs for different numerical resolutions are shown in Fig. 7. The respective wind fields (Fig. 5) used for these sensitivity runs were discussed in more detail in the previous subsection.

All simulations with different grid resolutions were able to capture the large-scale deposition patterns of snow, that is, reduced (enhanced) accumulation on the windward (leeward) slopes. All sensitivity runs also showed an increased snow loading on the leeside of pronounced secondary cross-slope ridges, albeit of different magnitude in the different spatial resolutions. It is clearly visible in Figs. 7c,d that for coarser grid simulations ($dx, dy = 25 \text{ m}, 50 \text{ m}$), any accumulation of snow on the windward slope did not occur. In contrast, a more adequate representation of smaller-scale topography—and thus flow features—caused a higher spatial variability of the snow depth distribution on both leeward and windward slopes. This also led to the more realistic result that, overall, more snow was deposited on the windward slopes

FIG. 7. Snow depth distribution caused by saltation, suspension, and preferential deposition with $dx, dy =$ (a) 5 m, (b) 10 m, (c) 25 m, and (d) 50 m.

and the completely bare areas on wind-exposed ridges observed at coarser resolutions no longer occurred. This is consistent with observations that bare areas on the windward slope are present only close to the ridge in reality.

Figure 8 shows modeled snow depths for a small strip of the Gaudergrat ridge. Measured snow depths for this strip are shown in Fig. 2. The location of the strip is highlighted in Fig. 7 by the black line. Measurements were made by manual snow probing before and after a major drift event in January 1999. Note that the sub-scale variability in snow depth is also caused by a high local roughness of the terrain that is not present in the

digital elevation model and is therefore not an object of our investigation. Observed new snow depths evidenced the formation of a snow dune on the windward slope (Fig. 2). This snow dune was successfully modeled in the sensitivity runs with finer grid resolutions ($dx, dy = 10$ m, 5 m) (Figs. 8a,b). For coarser grid resolutions the snow dune is missing (Figs. 8c,d). This development of the snow dune at finer grid resolutions is a result of a strong decrease in wind speeds in this area for respective grid resolutions of $dx, dy = 10$ m, 5 m (Figs. 5a,b, 6b). We did not compare absolute values of measured and modeled depths of new snow since modeled snow depth data were derived from sensitivity runs with one wind

FIG. 8. Modeled snow distribution for a small strip over Gaudergrat ridge. The horizontal resolutions (dx , dy) for sensitivity runs are (a) 5 m, (b) 10 m, (c) 25 m, and (d) 50 m.

field only, without an attempt to reconstruct the more complex drift event of 1999. We were mainly interested in the general pattern of snow deposition and in the relationship between mean snow depths (or total snow volume) on windward and leeward slopes. We therefore divided the small strip of the ridge into two subareas: the windward slope and leeward slope. We calculated the snow volume of the strip and its two subareas for the measurements and all sensitivity runs (Fig. 9). The analysis reconfirmed that more snow was deposited on the windward slope for higher grid resolutions (dx , $dy = 10$ m, 5 m) and that for the higher resolutions the relative distribution of snow mass across the ridge was very close to the measurements.

Observations and simulations at the finest grid resolution of 5 m showed that about 30% of the total ridge snow volume accumulated on the windward slope. In simulation runs with coarse grid resolutions only 10% of the total ridge snow volume was deposited on the windward slope. Figure 9b shows the mean snow depths of the windward and leeward subareas.

The final snow depths caused by drifting and blowing snow along three transects marked by solid lines in Fig. 7a

FIG. 9. Distribution of (top) snow volume and (bottom) mean snow depths for each subarea of a strip crossing the ridge.

are presented in Figs. 10a–c; transects 1–3 are identical to those discussed for the respective wind fields, shown in Fig. 6. Measured snow depths for transect 2 are shown in Fig. 10d, collected during the measurement campaign in 1999. The cross-ridge transects demonstrate that an increase of the horizontal grid resolution to 10 m provoked a maximum snow loading close to the ridge, as indicated by the distinct peaks of snow depth in Fig. 10. For the finest grid resolution a further (less pronounced) peak of snow depth was modeled downwind from the crest. By contrast, the snow loading became equally distributed tens of meters downwind of the ridge crest for the runs with a spatial resolution of 25 m. The snow cover was then much more uniformly distributed over the leeward slope. As already discussed in section 4a(1) and now apparent in the cross-ridge transects, the local shape and steepness of the mountain crest (as modified by the grid resolution) had a strong effect on the magnitude of snow redistribution by the wind. Measured snow depths also show the first peak of snow depth close to the ridge and a second peak at some distance from the ridge, similar to the modeled snow depths with the finest grid resolution (Fig. 10d). The snow dune in the windward slope is also visible in the first measurement points with a snow depth of about one meter (highlighted in Fig. 10d by a circle).

Finally, we want to reference the very distinct and narrow peaks of snow depth in transect 2 for grid resolutions of 10 and 5 m. These peaks correspond to pronounced and narrow peaks of wind speed gradients

FIG. 10. Transects of snow depths distributed by saltation, suspension, and preferential deposition with numerical resolutions of (a) 5 m, (b) 10 m, and (c) 25 and 50 m. (d) Transect 2 of measured snow depths (snowdrift event 1999).

(Fig. 11). Very low wind speeds in the lee and high wind speeds at the ridge indicate flow separation at this part of the ridge (Figs. 5a,b) and lead to local maxima of snow deposition. Flow separation is primarily responsible for the formation of cornices at the ridge crests. From the physics implemented in Alpine3D (Lehning et al. 2008), saltation processes are particularly sensitive to high gradients in wind speed. We may therefore interpret the simulated deposition peak as an approximation to a numerical simulation of a ridge cornice.

3) RELATIVE CONTRIBUTION OF PREFERENTIAL DEPOSITION AND TRANSPORT VIA SALTATION AND SUSPENSION AT VARIOUS SCALES

The relative contributions of saltation/suspension and preferential deposition were evaluated and analyzed with respect to the mass balance of the ridge. The final distribution of preferential deposition of precipitation for all numerical grids is presented in Fig. 12. In Fig. 13 the final snow depths along transects 1 and 2 caused by preferential deposition are compared with snow depths caused by both blowing/drifting snow and preferential deposition.

With preferential deposition of precipitation as the only mechanism, leeward-slope snow loading was enhanced but the snow distribution downwind from the crest was fairly uniform. On the leeward slope three pronounced

accumulation zones could be identified that were obviously influenced by flow features connected to the cross-slope ridges (Fig. 12). The windward slope, by contrast, only experienced a considerable amount of deposited

FIG. 11. Comparison of snow depths (sd) caused by drifting and blowing snow and wind speed (ws) for corresponding grid resolutions of transect 2.

FIG. 12. Patterns of snow depths caused by preferential deposition of snow alone (other drift processes are switched off) for dx , dy = (a) 5 m, (b) 10 m, (c) 25 m, and (d) 50 m.

precipitation if we force the simulation with resolved flow features at the finest grid scale (Figs. 12a,b). Subsequently, enhanced deposition of snow was encountered in a local depression on the windward slope, indicated in transect 2 (Fig. 13b) by a minor peak in snow depth.

Sensitivity runs for the full range of processes were already discussed in the previous section (Figs. 7 and 13). The simulations of preferential deposition, presented in this section, only confirm that with an increase of crest speedup, that is, runs with high grid resolution of 10 and 5 m, saltation as a transport process becomes more important. Consequently small-scale deposition features such

as cornicelike features and snow loading in the local depression on the windward slope are enforced owing to drifting and blowing snow.

From the various sensitivity runs, we conclude that the homogeneous snow accumulation patterns in the middle and lower parts of the leeward slope should be attributed to the preferential deposition of precipitation (Figs. 12 and 13b). In contrast, the formation of a cornicelike feature near the ridge (Fig. 13b) was clearly caused by snow redistribution processes.

To gain some quantitative information about the mass balance of the whole simulation area around the ridge, we again arbitrarily divided the ridge into two subareas,

FIG. 13. (a) Transect (a) 1 and (b) transect 2 of snow depths caused by preferential deposition alone vs snow depths caused by drifting and blowing snow for grid resolutions of $dx, dy = 10$ m, 25 m.

the windward and leeward slopes, outlined in Fig. 7d by green (windward slope) and red (leeward slope) lines. Then we calculated the snow volume on the ridge and its two subareas for all sensitivity runs. The fraction of subarea snow volume to total snow volume on the ridge and the mean snow depths gave an indication of the amount of snow deposited on the two slopes. We used the standard deviation of the snow depths as a simple measure of the variability of snow depth within the subareas (Fig. 14); however, note that this offers only a very limited view of snow depth distribution.

This analysis, now for a larger area across the ridge, again showed that considerably higher amounts of snow were deposited on the windward slope for higher grid resolutions ($dx, dy = 10$ m, 5 m). The distribution of the snow depths showed a normal distribution for finer resolution runs in both subareas (not shown). For coarser grid resolutions, however, the snow depth distribution on the windward slopes was a heavily skewed-right distribution, indicating the predominance of (almost) snow-free areas (not shown).

It is interesting to note that for a coarser grid resolution ($dx, dy = 25$ m/50 m) the fractions of subarea snow volumes were similar for runs with only preferential deposition and sensitivity runs including drifting and blowing snow (preferential deposition, saltation, and suspension). This indicated that snow redistribution

processes were not very active at coarser resolution. In contrast, for finer grid resolutions, the fractions of subarea snow volumes changed between sensitivity runs with drift enabled and those where only preferential deposition was enabled. The final snow distributions caused by drifting and blowing snow also showed an enhanced accumulation of snow on the windward slope compared to the snow distribution caused by preferential deposition. This can be explained by the existence of a saltation layer near the ground. Saltation may be a continuous additional source of snow for the windward slope. Due to the additional mass available in the saltation layer, the filling of topographical depressions on the windward slope was possible, and the snow became more evenly distributed between the leeward and windward portions of the study area.

The standard deviations calculated for snow depth distributions caused by drift processes were higher than those for preferential deposition only. This emphasizes, once more, the higher variability of the snow cover caused by snow redistribution processes.

b. Deposition features developed during the first snowfall event in October 2003

Snow transport processes were simulated for the drift and snowfall period from the 4–6 October 2003. This

FIG. 14. Distribution of snow volume for each subarea. Standard deviation and mean snow depths in the subareas for all sensitivity runs are given.

period was selected for study because in early fall a single snowfall event rarely deposits a meter of snow onto bare ground. Manual snow sampling and proxy snow depth estimation were made directly after the end of the snowfall event. During the snowfall event the prevailing wind was northwesterly. For some hours the wind backed toward the southwest.

Modeled and measured snow depths are plotted in Fig. 15. Additional information including the snow cover variability was provided by mapping zones of avalanche activity and the magnitude of erosion of the snow cover. The spatial resolution of measurement points was rather coarse, and sampling was not regularly distributed over

FIG. 15. Simulated snow depths for the snowfall and drift period in October 2003 compared to manual snow and proxy snow depths estimation (zones of avalanches and wind erosion).

the area. The observed snow depths, however, allow rough mapping of zones of erosion and accumulation.

The simulation results showed similar patterns of snow deposition and snow erosion to those for the 1999 event. The formation of the small-scale snow dune in the local depression is even more pronounced for the 2003 drift period than shown in the sensitivity runs. Note that no measurement points were available for this area. The more dense measurement points of snow depths after the storm event in 1999 (Fig. 2), however, clearly showed formation of the snow dune in this area and could be seen as an indirect verification of our model results. Observed snow depths showed the formation of cornices at the ridge crest. In model results a strong erosion of the snow cover was still indicated near the ridge crest. Owing to the limited resolution of the topography within the model, the maxima of snow accumulation, which could be interpreted as a formation of cornices, were modeled directly behind the ridge.

At the southern end of the ridge (location of transect 2) heavy snow loading was measured close to the ridge (within 25 m distance). A second maximum of snow loading was observed about 60 m away from the ridge. Snow loading to the lee of cross-slope ridges could also be detected from some measurement points. Note that small avalanches occurred in most areas of preferential deposition of precipitation and drifting snow, located on

FIG. 16. Photograph taken after the snowdrift event in October 2003.

the lee of the cross-slope ridges (also visible in Figs. 15 and 16). No measurement points were available for the windward sides of the cross-slope ridges. In this area the modeled erosion of the snow could, therefore, only be verified by a visual comparison of model results from photographs taken after the snowfall event (Fig. 16). The deposition/erosion patterns at the cross-slope ridges, both observed and simulated, indicated a strong degree of snow transport from the windward to the leeward side of the cross-slope ridges.

5. Conclusions

We have presented results from sensitivity studies of atmospheric and snow transport modeling as well as a case study simulating a snowdrift and snowfall event lasting three days. Our simulations further support the hypothesis that at the Gaudergrat ridge preferential deposition is active at the ridge scale and true redistribution—mainly via saltation—forms smaller-scale deposition patterns such as dunes and cornices. We demonstrated that a horizontal resolution of 10 m or less is necessary to calculate small-scale deposition patterns such as snow dunes and cornices, which can be attributed to snow redistribution by the wind. The analysis of total snow volume on the ridge and the snow volumes of the subareas windward/leeward showed that for coarser grids, more snow was deposited on the leeward slopes, while for finer grid resolutions a considerable amount of snow also accumulated on the windward slopes, in good agreement with field observations of this site.

The main purpose of this paper was to give a description of modeled mean flow fields and associated snow deposition. The flow fields are created by a common Reynolds-averaged Navier–Stokes (RANS) approach, and snow deposition modeling is based on a detailed physical representation of the snowdrift processes and preferential deposition as well as an investigation of the

effect of grid resolution on these processes. Also presented is a limited dataset for model verification. The detailed physical process representation is, however, still based on an average description of flow and transport processes. The saltation model (Clifton and Lehning 2008) is based on the assumption of equilibrium saltation, and we only use the adaptation of the mean flow field to topography to solve a stationary form of the advection–diffusion equation to describe preferential deposition and suspension (Lehning et al. 2008).

We will explore the implications of assuming that time-averaged equilibrium descriptions are able to represent the highly nonlinear instantaneous fluxes in future. This will involve measurements as well as Lagrangian simulations of flow–particle interactions in a large-eddy simulation (LES) frame (Bou-Zeid et al. 2007; Weil et al. 2004).

On the other hand, the detailed physical representation attempted here can also serve as a basis for simpler parameterizations of snow deposition (e.g., Winstral et al. 2002; Tabler 1975). A comparison with and development of simpler schemes is now especially valuable since the detailed representation provides an ability to distinguish processes that are—or are not—captured by parameterizations. For example, Dadic et al. (2010) have already shown that larger-scale mean velocities are well correlated with snow deposition and have suggested a parameterization based on mean wind speeds. In addition, the technology is now available to acquire high-resolution snow depth distribution maps via laser scanning (Prokop et al. 2008; Grünwald et al. 2010). These maps will give further insight into snow deposition processes and will help to establish simple parameterizations, needed for a variety of applications.

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