1. Introduction

Blowing and drifting snow is one of the crucial factors influencing the development of a spatially and temporally inhomogeneous snow cover over high-alpine terrain during the winter season. Avalanche activity is directly linked to the inhomogeneous snow distribution in slopes of different orientations. Furthermore, drifting snow can cause major problems for infrastructure such as roads and houses. Therefore, a quantitative assessment and forecast of snow transport by wind is of great practical value. Small-scale boundary layer flow features such as flow separation, local speed-up of the flow and channelling effects, which are typical for flow over steep and complex topography, are the driving mechanisms behind the inhomogeneous snow distributions [Lehning et al., 2008]. Thus, understanding and modeling the boundary layer wind field over steep and complex topography is necessary for a successful modeling of snow drift but also for understanding land surface exchange in such terrain.

The boundary layer flow over steep and complex terrain is highly turbulent and therefore strongly nonlinear. Experimental studies of turbulence in wind tunnels [e.g., Shuyang et al., 2001] and over real topography [e.g., Taylor and Teunissen, 1987; Vosper et al., 2002] have contributed a large amount of wind and turbulence data. The data collected during Riviera-MAP project in 1999 allow the analysis of mean flow and turbulence structures in a steep Alpine Valley [Rotach et al., 2004; Weigel et al., 2007]. However, the understanding of the boundary layer wind field and the turbulent flux of momentum over steep topography is still limited. While linear analysis has provided a detailed understanding of the main physical processes that are important for flow over gentle hills [e.g., Hunt et al., 1988], this approach is not applicable over steep alpine terrain.

Direct numerical simulation (DNS), where all eddy scales are solved by dynamic equations, is an attractive method to compute turbulent flows on the basis of first principles. However, the computational demand makes DNS impossible for any real atmospheric flow. This is due to the fact that the range of eddy scales within a turbulent motion increases dramatically with the Reynolds number [Wood, 2000]. The usage of Reynolds-averaged Navier-Stokes (RANS) models or large-eddy simulation (LES) techniques are two possible ways to overcome this problem. With the RANS approach, the equations of motion are solved for the Reynolds-averaged flow variables only, whereas the LES approach explicitly resolves the scales of turbulence greater than some spatial filter [Mason and Brown, 1999]. Therefore, a careful choice of the length scale of the spatial filter allows resolving the energetically most important motions within the LES approach, while the smaller scales are parameterized with a subgrid-scale (SGS) model.
[5] Pioneering work on mesoscale modeling of atmospheric flows using RANS models was undertaken in the mid-1970s, [e.g., Tapp and White, 1976; Taylor et al., 1976; Clark, 1977; Klemp and Wilhelmson, 1978]. Since the 1980s, simulations of flows over isolated hills of moderate slope, e.g., the Askervein study [Walsmsley and Taylor, 1996], were performed. Wood [1995] has undertaken numerical studies of flow separation over hills. During the past few years, an rapidly increasing computer power has made high-resolution simulations of atmospheric flows over complex terrain possible. Kim and Patel [2000] have tested different turbulence models and were able to simulate flow separation and recirculation of the boundary layer flow over two- and three-dimensional model topography applying the RANS approach.

[6] An adequate SGS closure model is a key issue for successful LES simulations of turbulent boundary layer flows, especially over complex terrain. Close to the bottom boundary, where the turbulent scales decrease in proportion to the distance to the surface, the energy-containing eddies are no longer resolved by the LES model. Rogallo and Moin [1984] mentioned the importance of the energy transfer between the resolved and parameterized motions within the LES approach, to assure a realistic generation of near surface turbulence. The numerical formulation of an appropriate SGS model is a matter of current research [e.g., Kumar et al., 2006; Chamecki et al., 2006]. While dynamic SGS models are very important in creating a correct near-surface flow over flat terrain [Porté-Agel et al., 2000] or help in finding correct roughness parameterizations for heterogeneous terrain [Bou-Zeid et al., 2007], the current study focuses on the reconstruction of mean flow characteristics in very steep terrain using a standard 1.5-order TKE closure. The same approach has successfully been used in other studies involving steep terrain [e.g., Chow et al., 2006]. In addition to the SGS model, grid resolution, numerical mixing and the turbulence scheme have been found to significantly influence the results of the computations of LES models [e.g., Zängl et al., 2004; Gohm et al., 2004]. Including topographic shading into the numerics of the Advanced Regional Prediction System (ARPS) model, [Chow et al., 2006] could improve the modeling of convective air flows in the Riviera Valley using a horizontal grid resolution of 150 m.

[7] In the current study, we apply the “Submeso” version [Anquetin et al., 1998] of the ARPS model to the computation of high-resolution wind fields over steep and complex terrain. Since our goal is to resolve flow features with scales below 100 meters, we have applied a very fine grid resolution. The compressible and nonhydrostatic ARPS model has been developed at the Center for Analysis and Prediction of Storms (CAPS) at the University of Oklahoma [Xue et al., 2000, 2004]. Preparing the Submeso version [Anquetin et al., 1998] modified the original ARPS code, enhancing the surface layer parameterization.

[8] Anquetin et al. [1998] used the Submeso model to calculate thermally driven circulations in deep mountain valleys, i.e., the formation and destruction of the thermal inversion layer using a horizontal resolution of a few hundred meters. In the current approach, we use the Submeso model to compute the development of boundary layer wind fields for storm events in the Alpine region. In contrast to Anquetin et al. [1998] and Chow et al. [2006], we use the ARPS model under strong wind conditions. In this setting, the formation of a turbulent boundary layer, the generation of air flow recirculation, the adjustment of the large-scale wind field and the momentum exchange with the ground are the most influential factors, whereas thermal effects are less important.

[9] The results of the fine-scale modeling of the wind field over steep and complex topography are verified by comparison against wind measurements at our experimental site, the Gaudergrat ridge near Davos in the southeastern part of Switzerland. The Gaudergrat ridge is a slightly concave-shaped mountain ridge in the Weisshüfl area near Davos. The sharp crest with slopes partly steeper than 45° reaches a height of approximately 150 m above the surrounding topography. The current study is a continuation of work by Föhn and Meister [1986], who investigated wind profiles and associated snow transport over Gaudergrat, and Gauer [2001], who initiated the study of the wind field over Gaudergrat with the current equipment. At five meteorological stations, wind speed and wind direction have been measured at three heights using a temporal resolution of 1 Hz.

[10] In a companion study, the wind fields computed with the ARPS model are used to drive the three-dimensional model of Alpine surface and snow drift processes, ALPINE3D [Lehning et al., 2006]. Detailed results of the snow drift modeling with ALPINE3D are presented in a companion paper by Lehning et al. [2008].

[11] The paper is structured as follows: In section 2, the configuration of the ARPS model for the current simulations is explained. In section 3, simulations of the wind field over Gaudergrat and the influence of initial as well as boundary conditions on the development of the wind fields in the interior of the domain are described. In section 4 the results of the simulations over Gaudergrat are compared to measurements. Section 5 contains summary and conclusions.

2. Model Description and Configuration

2.1. ARPS Model

[12] The ARPS model has originally been developed to study and predict storm-scale weather phenomena such as supercell storms, squall lines and tornados. The physics and parameterizations of the model incorporate processes ranging from the mesoscales down to the micrometres. Comprehensive verification of the ARPS model has been performed by comparing the model results against analytical solutions of linear and nonlinear mountain waves in flows past idealized topography. Observed small-scale weather phenomena such as thunderstorms and windstorms downstream of mountains have also been considered [Xue et al., 2000]. At the end of the 1990s, the “Submeso” version of the ARPS model has been developed at the University of Grenoble [Anquetin et al., 1998] to study thermally driven flows in steep Alpine valleys.

[13] In the following description of the ARPS model, we focus on the characteristics of the model that are important for the current application. A detailed and more comprehensive description of the ARPS model including the dynamic and numerical framework of the model as well as the different subgrid-scale turbulence and computational
mixing parameterizations can be found in work by Xue et al. [2000, 2004].

[14] ARPS uses a terrain-following coordinate system to solve the nonhydrostatic, compressible Navier-Stokes equations [Gal-Chen and Somerville, 1975]. In the ARPS implementation, the vertical coordinates and the associated Jacobians are defined numerically and can thus be arbitrary and time-dependent [Fiedler and Trapp, 1993; Fiedler et al., 1998]. Smoothing of the coordinate surfaces above a certain height helps to reduce errors associated with the deformation of the computational mesh. This property of ARPS is attractive as the truncation error due to grid transformations has the same leading order as that associated with the solution of the dynamical equations [Schär et al., 2002].

[15] The governing equations include the conservation of momentum, heat, mass, water and turbulent kinetic energy, as well as the equation of state of moist air. The model variables \( \Psi \) are defined as the sum of the base state variables \( \Psi = \Psi(x, y, z, t) \)

\[
\Psi(x, y, z, t) = \Psi(x, y, z) + \Psi'(x, y, z, t).
\]

[16] The ARPS model equations describe the time development of the deviations \( \Psi' \) from the horizontally homogeneous, time-invariant and hydrostatically balanced base state \( \Psi \). The horizontal homogeneity of the base state assures that the horizontal pressure gradient terms of the base state vanish. This reduces the computational error associated with the terrain-following coordinate system [Janjic et al., 2003].

[17] Since the ARPS model has been set up as an LES model, the deviations from the base state are divided into resolved scales \( \Psi \) and subgrid scales \( \psi \):

\[
\Psi(x, y, z, t) = \tilde{\Psi}(x, y, z, t) + \psi(x, y, z, t).
\]

Thereby, the SGS model acts as a filter operation of the simulation [Mason and Brown, 1999]. In the ARPS model, the size of the grid mesh \( \Delta \) defines the characteristic length scale \( l \) of the filter.

[18] In the current study, a 1.5-order turbulent kinetic energy-based SGS closure is used to complete the set of model governing equations [Deardorff, 1980; Moeng, 1984]. The components of the turbulent mixing coefficient for momentum \( K_m \) are herein related to the mixing length scales \( l_j \) of the SGS model and to the SGS turbulent kinetic energy \( e \) both coming from the SGS model,

\[
K_m = 0.1 \frac{e}{l_j}.
\]

For anisotropic turbulence, the length scales \( l_j \) are defined as

\[
l_1 = l_2 = \Delta h \quad \text{and} \quad l_3 = \begin{cases} 
\Delta v & \text{for unstable or neutral conditions} \\
\min(\Delta v, l_s) & \text{for stable conditions},
\end{cases}
\]

where \( l_s = 0.76C_1^3 N^{-1} \); \( \Delta h \) and \( \Delta v \) are the horizontal and vertical grid spacing, respectively. \( N \) is the Brunt-Väisälä frequency.

[19] When calculating surface fluxes of momentum, heat and moisture over nonflat terrain, the effective slope angle of the topography is considered in the Submeso version of the ARPS model [Anquetin et al., 1998]. The surface fluxes are computed orthogonal to the slope of the topography and then transformed into the Cartesian coordinate system. This avoids artificial sources and sinks across the lower boundary in the presence of steep terrain.

[20] As ARPS retains compressibility effects, meteorologically negligible acoustic waves are supported by the model. A split-explicit time integration scheme [Klemp and Wilhelmson, 1978] is used to improve the efficiency of the model. The acoustic terms are calculated every small time step, while all other terms are calculated only every big time step.

2.2. Model Setup

[21] Figure 1 shows an overview over the topography that surrounds the Gaudergrat ridge. Figure 2 shows a zoom of the computational domain of the Gaudergrat simulations along with the slope angle (gray shading). In order to account for the important topographic features influencing the air flow over the Gaudergrat ridge, a model domain of \( 1.5 \times 1.5 \text{ km} \) in the horizontal has been chosen. The setup is optimal for northwesterly inflow, when the Gaudergrat reaches a height of about \( H = 150 \text{ m} \) above the slightly sloped terrain in the inflow region. The prevailing wind direction during the passage of strong cold fronts is northwest and thus perpendicular to the ridge line. The Gaudergrat has a typical characteristic length scale of \( l = 500 \text{ m} \), here defined as the width of the obstacle perpendicular to the inflow. The average terrain slope in the direction of the mean flow amounts to \( 31^\circ \) (Figure 5), the maximum slope angle to \( 45^\circ \).

[22] The standard setup of the simulations presented uses a horizontal grid spacing of 25 m and 30 terrain-following levels between the surface and an altitude of 5000 m above sea level. The vertical spacing in the lowest layers amounts to 3 m. The small and large time steps are set to 0.001 and 0.1 s, respectively. At the lowest model level we use a kinematic lower boundary condition. At the top of the model domain the zero gradient formulation was applied, together with a 1500 m thick upper level Rayleigh damping layer. Regarding the lateral boundary conditions, we tested different setups, namely zero-gradient, fixed inflow, radiation open and periodic boundary conditions (see below for further details).

[23] The ARPS model was initialized with a single atmospheric profile, defining a horizontally homogeneous base state. Since boundary layer flows over cold surfaces are the focus of our study, the vertical profiles of the atmosphere used to initialize the ARPS model were assumed to be stably stratified. Model simulations for dry as well as humid upstream profiles were carried out. In the simulations including the Gaudergrat topography (section 3) a humid atmosphere was chosen, because nearly saturated conditions are typical for snowstorm conditions in the Alpine region.

[24] Analyzing measured as well as modeled vertical profiles of cold frontal passages in the Alps, an averaged vertical potential temperature gradient

\[
\frac{\partial \theta_s}{\partial z} = 0.5 \text{ K/100 m}
\]
and a boundary layer height of \( \delta = 500 \text{ m} \) were found to be representative. The averaged vertical potential temperature gradient leads to the Brunt-Väisälä frequency of

\[
N = \sqrt{\frac{g}{\Theta}} \frac{\partial \Theta}{\partial z} = 0.014 \text{ s}^{-1}.
\]

For the simulations including humidity, the model was initialized with a uniform profile of 90% relative humidity. This yields a vertical gradient of the equivalent potential temperature of

\[
\frac{\partial \Theta_e}{\partial z} \approx 0.35 \text{ K/m}.
\]

Therefore, the stratification is reduced if saturation occurs. However, as the vertical displacements are comparatively small, dry and moist simulations yield very similar flow structures. In the humid case the humidity profile quickly adjusts after initialization, resulting in typical values of the relative humidity in the range between 80% and 95%. In none of the simulations did a significant cloud cover develop.

The horizontal wind was initialized with a standard logarithmic wind profile within the boundary layer:

\[
\frac{u(z)}{u_*} = \frac{\ln(z/z_0)}{\kappa}
\]

Here \( z \) defines the vertical distance to a horizontal plane crossing the lowest grid point of the topography, \( u_* \) is the friction velocity and \( \kappa = 0.4 \) is the von Karman constant. On the basis of the measurement of Doorschot and Lehning [2002] on the roughness length during snow fall events in similar terrain, an aerodynamic surface roughness of \( z_0 = 0.01 \text{ m} \) was chosen.

Above the boundary layer, a layer with vertically constant free-stream velocity \( u_\infty \) was assumed. Choosing
the free-stream velocity $u_\infty$, the friction velocity $u_*$ can be calculated (equation (5)). The solid line of Figure 10a shows the initial wind profile for a simulation using a westerly wind of $u_1 = 6 \text{ m s}^{-1}$. Since the initialization uses a horizontally homogeneous vertical profile, a standard logarithmic wind profile creates strong initial winds for the grid layers close to the summit of the topography.

2.3. Dynamical Considerations

[28] Before proceeding, we briefly discuss the dynamical setting in the context of the available literature. In particular, we would like to isolate the key factors that determine the nature of the flow regime, and to know whether boundary layer separation is likely to occur. We start by considering the dimensionless mountain height $H_{\text{Smith and Gronas}, 1993}$, which provides a measure of the role of stratification in orographic flows. Using the dimensional values listed above, we estimate it as

$$H = \frac{NH}{u_\infty} \approx 0.35.$$  \hfill (6)

Although most of the literature discusses flows over substantially broader topography than considered here [Schär and Durran, 1997; Bauer et al., 2000], the comparatively low dimensionless mountain height suggests that stratification effects do not play a major role in determining the flow regime. Indeed, as we have used the dry value of the Brunt-Väisälä frequency, and as we have neglected the presence of the boundary layer with vertical mixing and spatially varying height, the stratification effects are likely even smaller than indicated by the dimensionless mountain height (6).
Next we estimate the role of nonhydrostatic effects using the dimensionless parameter

\[ \mathcal{S} = \frac{u_\infty}{N L} \approx 0.86. \]  

(7)

It implies that the hydrostatic wavelength of the flow is similar as the dimension of the topography. The flow will thus be strongly affected by nonhydrostatic effects. The dimensionless parameters (6) and (7) together suggest that, in absence of a boundary layer (i.e., with a free-slip lower boundary condition), we would expect vertically evanescent gravity waves, possibly with some trapped lee waves [Smith, 1979].

The crucial question thus relates to the effects of the boundary layer. Several studies have addressed the role of boundary layers on neutral flows over small hills. On the basis of theoretical considerations and numerical simulations, Wood [1995] found that boundary layer separation will occur provided the terrain slope \( \alpha \) exceeds some critical slope \( \alpha_{crit} \) somewhere on the hill, and that \( \alpha_{crit} \) will approximately only depend upon a dimensionless length scale

\[ L = \frac{L}{z_0} \approx 1.5 \times 10^4 \]  

(8)
determined by the width of the hill and the surface roughness length. Using simulations for \( L = 10^4 \), Wood found a critical slope angle of \( \alpha_{crit} = 17^\circ \). The analysis of Wood [1995] has later been extended by Ross et al. [2004] to include cases with stratification. His study showed that stratification reduces the strength of the turbulence and the scale turbulence is weaker. The presence of this small-scale turbulence as well as a large roughness length was found to significantly alter the structure of the whole flow regime.

Although the quoted value of \( \alpha_{crit} = 17^\circ \) applies to a somewhat smaller dimensionless scale \( L \), the above discussion suggests that for the Gaudergrat, with slopes up to \( 45^\circ \), we should anticipate boundary layer separation, at least in cases where the incident flow is close to perpendicular to the ridge.

2.4. Idealized Simulations

The main message from the previous subsection is that the Gaudergrat is likely to promote boundary layer separation for the flow settings considered. It is of key importance to properly represent this aspect of the flow. In the absence of boundary layer separation, the flow away from the topography will only weakly be affected by the boundary layer, as most of the horizontal vorticity will be concentrated in shallow layer along the topography. As soon as boundary layer separation occurs, however, vorticity will be advected into the interior of the flow, and this will fundamentally alter the structure of the whole flow regime.

To find an appropriate model setup and to investigate the ability of ARPS to reproduce the characteristic flow features, a series of numerical experiments for homogenous flows past two-dimensional triangular ridges with similar dimensions as the topography of the Gaudergrat ridge has been undertaken. The slope ranged from \( 15^\circ \) to \( 45^\circ \). Because fully three-dimensional studies on a very fine grid including a model domain of \( 1.5 \times 1.5 \) km are extremely computer time consuming most of the idealized studies were undertaken using a setup of only four grid points perpendicular to the flow direction, hereafter called quasi two-dimensional studies.

The experiments with idealized topography indicate that a horizontal resolution of at least \( 25 \) m and a stretched vertical grid ranging from \( 3 \) m near the ground to \( 300 \) m near the top of the model domain are required to realistically simulate important flow features such as flow separation in the lee or speed-up over the ridge. The resulting grid aspect ratio remains moderate (between \( 0.1 \) near the top of the model domain and \( 10 \) near the ground). Near the topography the grid levels are terrain following while close to the top of the domain the grid layers are horizontal. This fine-scale mesh implies a small time step of \( 0.001 \) s and a big time step of \( 0.1 \) s in our simulations.

Figure 3 shows the time evolution of the recirculation downwind of an idealized ridge with slopes of \( 30^\circ \). The model setup described above has been used together with zero gradient boundary conditions. The model was initialized with a free stream velocity of \( u_\infty = 13 \) m s\(^{-1}\) in combination with a dry atmosphere.

The snapshot after \( 600 \) s integration time (Figure 3a) shows a fully developed separation zone downstream of the ridge. A recirculating eddy has been detached from the ridge and is moving downstream with the mean flow (position \( 1250 \) m). Concurrently, the formation of a new recirculating eddy starts over the downwind slope (position \( 750 \) m). In the region directly upwind of ridge, blocking of the flow leads to significantly reduced wind speeds. The near surface flow upwind of the ridge shows additional small-scale fluctuations. After \( 690 \) s (Figure 3b), the first recirculating eddy has been moved further downwind (position \( 1500 \) m) and the formation of the second recirculating eddy is still ongoing. Also after \( 780 \) s (Figure 3c), the source of near surface vorticity over the downwind slope leads to a further growth of the recirculating eddy. The life cycle of such a recirculating eddy has a timescale of approximately \( 300 \) s.

An additional series of idealized experiments has been conducted using the above setup with different lateral boundary conditions. Table 1 shows the variation of the wind speed at three different points along a model ridge for different boundary conditions. We found that the intensity of the resolved small-scale turbulence is strongest for zero-gradient and periodic lateral boundary conditions, whereas for fixed inflow boundary condition the intensity of small-scale turbulence is weaker. The presence of this small-scale turbulence as well as a large roughness length was found to favor the development of flow separation downwind of the ridge. Our idealized simulations suggest that for a perpendicularly incoming flow, a critical slope angle of the topography of \( 20^\circ \) is necessary for the formation of recirculation. This result is consistent with the discussion of the literature in the previous subsection. As the choice of the lateral boundary conditions appears crucial, we present in section 3 simulations for Gaudergrat using different approaches.

3. Gaudergrat Wind Field Simulations

Computations of the wind field over the Gaudergrat ridge for different initial and lateral boundary conditions are presented in this section. Using a horizontal grid size of
25 m, we focus on the simulation of the main boundary layer flow features caused by the ridge. An even higher resolution would be essential for our drifting snow simulations in order to resolve flow features caused by topographic features smaller than ridge size, but such simulations cannot be implemented for our terrain at present because of limited computational resources. An even higher grid resolution would additionally cause slopes steeper than 45° in our model terrain, which cannot be handled properly by the terrain-following coordinates of ARPS resulting in unstable flow simulations.

3.1. Fixed Inflow Boundary Conditions

In Figures 4 and 5, results of ARPS computations initialized with a southwesterly wind of $u_{\infty} = 6 \text{ m s}^{-1}$ free stream velocity are presented. Since we intended to study

Table 1. Mean Wind Speed and Variance of the Mean Wind Speed for an Integration Time of 10 min for Different Combinations of Boundary Conditions Based on 1 s Model Output

<table>
<thead>
<tr>
<th>Boundary Conditions</th>
<th>$u_{w}$ (m s$^{-1}$)</th>
<th>$\sigma_{u_{w}}$ (m s$^{-1}$)</th>
<th>$u_{r}$ (m s$^{-1}$)</th>
<th>$\sigma_{u_{r}}$ (m s$^{-1}$)</th>
<th>$u_{d}$ (m s$^{-1}$)</th>
<th>$\sigma_{u_{d}}$ (m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zero gradient</td>
<td>3.6</td>
<td>0.9</td>
<td>14.0</td>
<td>2.3</td>
<td>$-1.9$</td>
<td>3.8</td>
</tr>
<tr>
<td>Periodic</td>
<td>3.9</td>
<td>0.3</td>
<td>14.0</td>
<td>0.3</td>
<td>1.6</td>
<td>0.9</td>
</tr>
<tr>
<td>Fixed inflow/radiation open outflow</td>
<td>3.9</td>
<td>0.1</td>
<td>14.8</td>
<td>0.2</td>
<td>1.5</td>
<td>0.9</td>
</tr>
<tr>
<td>Fixed inflow/zero-gradient open outflow</td>
<td>2.9</td>
<td>0.6</td>
<td>13.1</td>
<td>1.0</td>
<td>1.2</td>
<td>1.0</td>
</tr>
<tr>
<td>Radiation open boundaries</td>
<td>3.7</td>
<td>0.3</td>
<td>13.6</td>
<td>0.6</td>
<td>1.6</td>
<td>0.9</td>
</tr>
</tbody>
</table>

*Here $u_{w}$ defines the wind speed over the middle part of the upwind slope, $u_{r}$ defines the wind speed on the ridge line, and $u_{d}$ defines the wind speed 125 m downwind of the base of the downwind slope.
the adaptation of the boundary layer wind field to the terrain for specific inflow conditions, fixed inflow boundary conditions in combination with radiation open outflow boundary conditions were used. The combination of fixed inflow with zero-gradient-outflow boundary conditions was found to cause a significant damping of the turbulence in the interior of the domain for the three-dimensional simulations.

In Figure 4, a snapshot of the computed wind field over the central part of the Gaudergrat ridge for a southwesterly flow using fixed inflow boundary conditions after 150 s integration time is presented. A region of
strong speed-up of the flow has formed along the ridge line. The averaged fractional speed-up for the lowest model level is

$$\tau_{\text{hill}} = \frac{\bar{u}_L - \bar{u}_a}{\bar{u}_a} = 1.4.$$  

(9)

Here $\bar{u}_a$ is the undisturbed wind speed in the inflow region and $\bar{u}_L$ the wind speed at the ridge line. A maximum horizontal wind speed of $10.2 \text{ m s}^{-1}$ is reached over the southern part of the ridge.

[42] The inflow is not perpendicular to the ridge line and the modeled fractional speed-up of 1.4 is significantly smaller than the theoretically expected value for gentle two-dimensional ridges [Taylor and Teunissen, 1987; Stull, 1988]:

$$\Delta U_{\text{hill}} = \frac{2H}{L_{(1/2)}} \approx 2.4,$$  

(10)

where $H \approx 150 \text{ m}$ is the height of the ridge and $L_{(1/2)} \approx 125 \text{ m}$ is the half width of the ridge.

[43] The speed-up is associated with a $10^\circ$ change of the wind direction to the west. This is probably the result of the acceleration of the flow perpendicular to the ridge line. A region of flow separation has formed over the downwind slope of the ridge. On the bottom of the downwind slope, channel effects cause a southerly flow parallel to the ridge line. The channel effects are caused by the mountain ridge southeast of Gaudergrat (see Figures 1 and 2). The channel effects are strongest near the position of the wind mast $M_o$ on the bottom of the downwind slope ($M_o$ is marked in Figure 2). In the middle part of the ridge, where the cross section of Figure 6 has been taken, the wind has a small component directed toward the ridge. This weak flow reversal is also visible in the cross section of Figure 5.

[44] The situation illustrated in Figures 4 and 5 characterizes a quasi steady state of the flow in the interior of the domain. For longer integration times (more than 10 min) the flow field starts to diverge toward an unstable state because of intensifying perturbations which result from the inhomogeneity of the topography on the southern and northern boundaries, but no new significant flow features develop.

[45] In Figures 6 and 7, the same model setup as for the previous simulation was used but the inflow was now a westerly wind of $u_\infty = 6 \text{ m s}^{-1}$ free stream velocity. The situation after 8 min integration time shows again a region of strong speed-up near the crest line. The maximum wind speed of $12.9 \text{ m s}^{-1}$ is reached on the crest line, slightly to the north of the cross section. The averaged fractional speed-up for the lowest level of the computational grid is $\tau(z)_{\text{hill}} = 1.8$. Since the inflow is now almost perpendicular to the ridge line, this value is now higher but still lower than the one expected for gentle hills. Over the northern part of the downwind slope, a sheltered region with distinct reduction of the wind speed exists and a flow reversal has formed. The small recirculating zone is nicely visible in the cross section of Figure 7. During the time evolution of the model simulation, the intensity of the recirculating region varies slightly but the position stays the same. Downwind of the
ridge the flow direction changes because of channel effects from west to northwest. This effect is not as strong as for the situation with southwesterly flow.

Experiments with different initialization wind profiles suggest that the main flow features as well as the onset of separation strongly depend on the wind speed and wind direction at the inflow boundary upwind of the ridge. If the flow is perpendicular to the crest line, a critical wind speed exceeding 4 m s\(^{-1}\) is necessary to generate separation of the boundary layer flow. If the wind direction upwind of the ridge is not perpendicular to the ridge, more air flows around the ridge and the formation of a recirculating eddy becomes more and more unlikely. A deviation of approximately 30° from perpendicular was found to be critical.

Also the fractional speed-up shows a strong dependency on the wind direction (Table 2). For this analysis a free stream velocity \(u_\infty = 6\) m s\(^{-1}\) was used. The highest speed-up of 1.8 was reached for westerly inflow. If the deviation from the perpendicular incidence exceeded more than 45°, the fractional speed-up decreased rapidly. For a range of \(u_\infty\) between 3 m s\(^{-1}\) and 10 m s\(^{-1}\), the simulated fractional speed-up was almost independent of the wind speed.

### 3.2. Periodic Boundary Conditions

As mentioned in the subsection 2.4, the choice of boundary conditions is a crucial factor for the transient development of the turbulence structure and the onset of separation in the interior of the model domain. Figure 8 shows a snapshot of the computed wind field over the central part of the Gaudergrat ridge on the lowest level after 8 min integration time using periodic lateral boundary conditions. All other model parameters were the same as for the previous simulation shown in Figures 6 and 7. The turbulence of the wind field is significantly stronger than for the case study with fixed inflow boundary conditions. This is due to the fact that turbulence, created over the Gaudergrat ridge, is advected back into the domain by the periodic boundary conditions.

The adaptation of the wind field to the steep terrain results in a complex three-dimensional flow pattern. Over the ridge line, again a speed-up region has developed. Downwind of the crest two vortices with vertically oriented axes have formed as a result of the interaction between the air streams flowing over and around the ridge, respectively. The vertical cross section in Figure 9, corresponding to the simulation of Figure 8, shows a strong rotor with a horizontally oriented axis which occurred in the separation region downwind of the crest. The cross section was taken in the northern part of the Gaudergrat ridge, where the separation eddy is strongest. Together the two pictures nicely visualize the complex three-dimensional structure of the flow.

### 3.3. Time Development of Typical Flow Features

In Figures 10a and 10b, vertical profiles of the west-east wind component are illustrated for the simulation with periodic boundary conditions after 30 s and 300 s integration time for five locations across the ridge. The five wind profiles were taken in a cross section at locations similar to the locations of the wind masts (Figure 2).

Starting with the initial wind field described in section 2.2, the adaptation of the mean wind field to the terrain takes only a few seconds. After 30 s, the profile in the inflow region of the model is identical to the initialization profile (thick solid line). Negative pressure perturbations at the ridge cause an acceleration of the flow over the ridge.

![Figure 7. Vertical cross section of the west-east wind component over the northern part of the Gaudergrat ridge for a westerly flow using fixed inflow boundary conditions after 8 min integration time. The position of the cross section is marked in Figure 6.](image)

### Table 2. Dependency of the Modeled Fractional Speed-Up on the Wind Direction for 6 m s\(^{-1}\) Free-Stream Velocity

<table>
<thead>
<tr>
<th>Wind Direction</th>
<th>Speed-Up</th>
</tr>
</thead>
<tbody>
<tr>
<td>Southwest</td>
<td>1.4</td>
</tr>
<tr>
<td>West-southwest</td>
<td>1.7</td>
</tr>
<tr>
<td>West</td>
<td>1.8</td>
</tr>
<tr>
<td>West-northwest</td>
<td>1.2</td>
</tr>
<tr>
<td>Northwest</td>
<td>3.0</td>
</tr>
<tr>
<td>North-northwest</td>
<td>0.5</td>
</tr>
</tbody>
</table>
upper part of the upstream slope. Therefore, on both slopes and on the ridge top, a typical speed-up wind profile has developed. The speed-up in both slopes is symmetric which indicates that no hydrostatic wave formation occurs downwind of the crest. In the profile on the ridge top, the highest wind speed of $u_{c} = 10.5 \text{ m s}^{-1}$ is reached at the lowest grid level 3 m above the topography. The modeled maximum fractional speed-up is $S_{\text{hill}} = 1.9$. In this state of the simulation, flow separation but no clear recirculation is observed.

**Figure 8.** Wind field of the first grid level over the central part of the Gaudergrat ridge ($\times 2246$, 46°51'25"N, 9°47'52"E) for a westerly flow using periodic boundary conditions after 8 min integration time. The position of the cross-section of Figure 9 is marked with two small lines at the west and east sides.

**Figure 9.** Vertical cross section of the west-east wind component over the northern part of the Gaudergrat ridge using periodic boundary conditions after 8 min integration time. The position of the cross section is marked in Figure 8. The cross section was taken slightly north of the cross section of Figure 7.
During the progress of the simulation, positive pressure perturbations downstream of the ridge decelerate the flow over the downstream slope. This adverse pressure gradient leads to the production of near surface vorticity of opposite sense to the boundary layer vorticity. Triggered by resolved, small-scale turbulent eddies, a recirculation zone begins to form. After 300 s, because of the periodic boundary conditions, the inflow profile is slightly modified.

In the inner sections of the domain, the flow regime has changed significantly. Over the downwind slope a separation eddy has formed which leads to a strong negative horizontal wind component close to the topography. Between the lowest grid level and the fifth grid level, a region of strong wind shear has formed. The separation eddy goes along with modified speed-up profiles over the upwind slope and on the top of the ridge. The fractional speed-up

**Figure 10.** Vertical profile of the west-east wind component after (a) 30 s and (b) 300 s integration time.
is reduced to a value of \( \tau_{\text{hill}} = 1.3 \). As a function of time, the speed-up of the flow at the ridge undergoes variations, which are dependent on the strength of the recirculating eddy. The two steps of the time evolution give an impression about the influence of the development of the recirculating eddy on the vertical wind profiles.

[54] During the time evolution of the simulation, the intensity of the separation eddy varies because parts of the turbulence are detached from the ridge. But, because of a strong source of vorticity over the downwind slope, the recirculating eddy builds up again at the same location (see also subsection 2.4).

4. Comparison With Measurements

[55] In this subsection a qualitative comparison of the model simulations with wind measurements during a 5-day-long snow drift period at the end of January 1999 is shown. The snow drift event was associated with strong westerly to northerly winds. The same drift event is treated in the companion paper [Lehning et al., 2008]. During the whole

Figure 11. Measured time series of 10 s averages of (a) wind direction and (b) wind speed approximately 1 m above the snow cover on the wind masts upwind (circles) and downwind (crosses) of the Gaudergrat ridge. Measurements start at 26 January 1999, 1846 UTC.
snow drift event, wind measurements were performed approximately 1 m, 3 m and 5 m above the snow cover at the five wind stations described above. The wind masts are situated in a line that is approximately perpendicular to the crest line (Figure 2). The masts upwind and downwind of the crest are positioned at a horizontal distance of approximately 500 m from the crest line. The masts on both slopes have a horizontal distance of 30 m from the crest line. At the stations upwind and downwind of the ridge, the horizontal wind speed and wind direction were measured, whereas at the stations near the ridge line, all three orthogonal wind components were measured at the three heights.

In this verification of the model results, we focus on the characteristic flow features over the Gaudergrat ridge. There are several limitations that make a strict verification of the ARPS model results difficult: (1) The exact initial conditions of the atmosphere upstream of the ridge are unknown. The assumption of a profile with a vertically constant Brunt-Väisälä frequency and a logarithmic wind profile is a reasonable guess but will never be true in detail. (2) The model domain does not include the topography of the larger-scale upstream environment that might influence the wind field in the interior of the model domain. (3) The coarse grid resolution does not resolve all topographical features in the domain that might be important for the local wind. (4) The model grid points do not exactly correspond to the measurement locations. Therefore, only a qualitative comparison between model results and measurements is possible.

4.1. Channel Effects

Because of the complex topography east of the Gaudergrat ridge, the flow regime on the bottom of the downwind slope changes rapidly, dependent on the wind direction in the inflow region. Channel effects on the bottom of the downwind slope result in two preferential wind directions approximately parallel to the ridge line. Figures 11a and 11b illustrate the jump of the flow direction on the wind mast on the bottom of the downwind slope. For westerly inflow (270°) there is southerly flow downwind of the ridge. Once the flow in the inflow region turns more to the north, the wind direction at the wind mast on the bottom of the downwind slope swaps toward northeast.

The phenomenon of two predominant, opposite flow directions over the downwind slope cause wave structures within the accumulation patterns of the freshly fallen snow (Figure 12). The dunes on the snow surface on the bottom of the lee slope indicate the two predominant opposite flow directions. The wind direction is perpendicular to the dunes with the steep front of the waves on the downwind side [Kennedy, 1969]; that is, the wind has been blowing from right to left in the foreground of the picture and from left to right in the background near the weather station.

As shown in Figures 4 and 6, the model simulations reproduce parts of the phenomenon of two predominant but opposite flow directions over the downwind slope. The channel effects are weaker in the model simulations especially over the southern part of the Gaudergrat ridge. Additionally, the jump in wind direction over the downwind slope occurs for a slightly different wind direction for the measurements and the simulations. While the jump is for northwesterly inflow for the measurements, it occurs for westerly inflow for the simulations.

The main reason for the partial reproduction of the channeling effects is that only parts of the surrounding topography downwind of the ridge could be included within
the model domain. This is necessary to make the model topography as homogeneous as possible at the boundaries of the model domain to avoid numerical perturbation of the flow caused by the lateral boundaries.

4.2. Fractional Speed-Up

Another important feature of the wind field over Gaudergrat is the speed-up of the flow over the ridge line. Analyzing 10 s wind data during the 5-day-long snow drift period at the end of January 1999, a measured mean fractional speed-up of 0.6 has been found. This result confirms former studies on the Gaudergrat by Gauer [2001]. The maximal fractional speed-up is reached 1 m above the snow cover. On the anemometer 3 m and 5 m above the snow cover the speed-up is 0.5 and 0.4, respectively.

An example of a time series of the measured wind speed on the stations upwind of and on the Gaudergrat ridge as well as the associated fractional speed-up 1 m above the snow cover. On the anemometer 3 m and 5 m above the snow cover the speed-up is 0.5 and 0.4, respectively.

Splitting the wind data of the complete drift period into different wind directions, a strong dependency of the fractional speed-up on the wind direction (Table 3) becomes evident. The comparison between measured and modeled fractional speed-up indicates that the slightly asymmetric dependency of the speed-up on the wind direction can be reproduced by the model. However, the model overestimates the speed-up in general, which partly may be due to the smoothed model topography based on the digital elevation model of 25 m resolution. Both measured and modeled speed-up reach their maximum for westerly inflow.

Keeping the wind direction fixed, the occurrence of recirculation downwind of the ridge line is the most influencing factor for the intensity of the speed-up (Figure 14). In the presence of separation downwind of the ridge, the mean speed-up is 0.6. As soon as the recirculation breaks down the speed-up increases to 1.2.

Analyzing 30 min averaged data for a 14 h period of strong northwesterly winds and snowfall, no significant dependence of the fractional speed-up on the snow fall intensity was found (not shown). This is qualitatively consistent with the considerations presented in section 2.

4.3. Recirculation

The most challenging task for the ARPS model is the reproduction of the formation of recirculating eddies over the downwind slope of Gaudergrat. The onset of recirculation is dependent on the wind direction and wind speed upwind of the ridge. In the measurements, the recirculation on the wind station on the downwind slope of the Gauder-

![Figure 13. Time series of 1 min averages of the measured wind speed on the wind masts upwind of the Gaudergrat ridge (triangles) and on top of the Gaudergrat ridge (circles) as well as the fractional speed-up (crosses) 1 m above the snow cover. Measurements start at 27 January 1999, 1600 UTC.](image)

<table>
<thead>
<tr>
<th>Wind Direction</th>
<th>Measured Speed-Up</th>
<th>Modeled Speed-Up</th>
</tr>
</thead>
<tbody>
<tr>
<td>Southwest</td>
<td>0.9</td>
<td>1.4</td>
</tr>
<tr>
<td>West-southwest</td>
<td>1.1</td>
<td>1.7</td>
</tr>
<tr>
<td>West</td>
<td>1.5</td>
<td>1.8</td>
</tr>
<tr>
<td>West-northwest</td>
<td>0.9</td>
<td>1.2</td>
</tr>
<tr>
<td>Northwest</td>
<td>0.7</td>
<td>1.0</td>
</tr>
<tr>
<td>North-northwest</td>
<td>0.6</td>
<td>0.5</td>
</tr>
</tbody>
</table>
grat ridge occurs for a range of wind directions between 300° and 330° as observed at the wind station in the inflow region. Figure 15 shows how the recirculating eddy in the downwind slope collapses when the wind direction upwind of the ridge turns from northwest to the north, and reappears when the wind direction turns back to northwest. A wind speed of approximately 2 m s⁻¹ at a height of 3 m above the snow surface is necessary for the onset of recirculation.

As described in section 3, the ARPS model is able to create a recirculation zone downwind of the crest line of the Gauergrat ridge. Recirculation was found for the simulations with wind directions northwest, west and southwest. It takes approximately 150 s integration time until an organized separation eddy is formed. The intensity of the turbulence in the recirculation zone is strongly dependent on the turbulence status of the inflow wind field. The recirculating eddy is positioned over the center of the downwind slope. In our simulations, only strong recirculating eddies reach the position of the available weather station on the downwind slope. Because of the lack of additional weather stations, no exact investigation of the size of the recirculation zone is possible.

5. Conclusions and Outlook

In this study, the Submeso version of the ARPS model has been used for the fine-scale modeling of the wind field over steep and complex terrain. Applying the model to an idealized ridge, we found that a grid resolution of at least 25 m is necessary to resolve the overall structure of the wind field as well as characteristic flow features such as speed-up of the flow over the ridge line, flow separation and recirculation downwind of an obstacle with horizontal and vertical scales of 500 m and 150 m, respectively.

A strong dependency of the simulated flow field in the interior of the domain with respect to the lateral boundary conditions and grid resolution has been found. The advection of turbulent wind perturbations into the inflow region supports the generation of recirculation downwind of the ridge. Periodic boundary conditions lead to a recycling of such turbulent structures into the inflow region of the domain. For fixed inflow boundary conditions, the generation of adequate turbulence in the inflow region is missing. A deviation of approximately 30° of the incoming flow from the perpendicular direction was found to be critical for the onset of recirculation downwind of the ridge, for stronger deviations no boundary layer separation takes place.

When using realistic topography, difficulties arise regarding the formulation of reasonable boundary and initial conditions for the model simulations. So far, we are able to calculate the adaptation of the wind field to steep and complex terrain for fixed or periodic boundary conditions, but no computation using a time-dependent wind field has been performed yet. To run the ARPS model in a forecast mode, three-dimensional time-dependent boundary and initial conditions are necessary. The most reasonable way to provide such boundary conditions is a grid nesting technique. Since a grid nesting starting with a resolution of a few kilometers down to a resolution of a few meters is expensive to realize, a downscaling method has been developed at the SLF that allows the generation of vertical atmospheric soundings using the output of the operational weather forecast model of MeteoSwiss [Kaufmann et al.,...
Acknowledgments. The authors would like to thank all the people who helped to prepare and conduct the field experiments on Gaudergrat, some of them spending many hours outside in a snow storm. A lot of effort from Peter Gauer has gone into the site. Thanks also to the GIS (Andy Stoffel) and IT specialists for their support. The work has been funded partly by the Swiss National Science Foundation. The numerical simulations with observations, on 24 October 1999 (MAP IOP 10): Verifications of high resolution numerical simulations with observations, Mon. Weather Rev., 122, 2879–2888.


References


2003; Spreitzer and Raderschall, 2004]. Initializing the ARPS model with such a sounding, it can be used in a simplified forecast mode.

The wind fields characteristics computed with the ARPS model have been successfully used to drive a numerical three-dimensional snow model which has been developed at the SLF [Lehning et al., 2006]. The application of the wind fields over Gaudergrat computed by the ARPS to the modeling of a snow drift event are described in the companion paper by Lehning et al. [2008].

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References


