Atmospheric Flow Development and Associated Changes in Turbulent Sensible Heat Flux over a Patchy Mountain Snow Cover

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ABSTRACT

In this study, the small-scale boundary layer dynamics and the energy balance over a fractional snow cover are numerically investigated. The atmospheric boundary layer flows over a patchy snow cover were calculated with an atmospheric model (Advanced Regional Prediction System) on a very high spatial resolution of 5 m. The numerical results revealed that the development of local flow patterns and the relative importance of boundary layer processes depend on the snow patch size distribution and the synoptic wind forcing. Energy balance calculations for quiescent wind situations demonstrated that well-developed katabatic winds exerted a major control on the energy balance over the patchy snow cover, leading to a maximum in the mean downward sensible heat flux over snow for high snow-cover fractions. This implies that if katabatic winds develop, total melt of snow patches may decrease for low snow-cover fractions despite an increasing ambient air temperature, which would not be predicted by most hydrological models. In contrast, stronger synoptic winds increased the effect of heat advection on the catchment's melt behavior by enhancing the mean sensible heat flux over snow for lower snow-cover fractions. A sensitivity analysis to grid resolution suggested that the grid size is a critical factor for modeling the energy balance of a patchy snow cover. The comparison of simulation results from coarse (50 m) and fine (5 m) horizontal resolutions revealed a difference in the spatially averaged turbulent heat flux over snow of 40%–70% for synoptic cases and 95% for quiescent cases.

1. Introduction

Mountainous terrain typically induces two types of winds, diurnal mountain wind systems and terrain-forced flows (Defant 1949). Diurnal mountain winds are local thermally driven wind systems that form over mountainous terrain (Zardi and Whiteman 2013). Slope flows, the smallest-scale diurnal wind system, respond rapidly to temporal and spatial variations in sensible heat fluxes, air

temperature, and surface temperature caused by variations in the surface energy budget (Prandtl 1952; Haiden 2003; Serafin and Zardi 2010; Nadeau et al. 2013). Terrain-forced flows develop when large-scale winds are modified by the underlying topography. While diurnal mountain winds are strongest during clear sky conditions and when the synoptic wind is weak, terrain-forced flows become most pronounced if synoptic winds are strong (Whiteman 2000).

Segal et al. (1991) numerically investigated the effect of high land-cover heterogeneity caused by patchy snow cover on the development of a complex thermal wind system consisting of katabatic and anabatic winds at the same time. They could show that at the mesoscale (20–200 km), the existence of snow-covered areas adjacent to snow-free areas produces horizontal gradients in air temperature and pressure during the daytime, possibly resulting in

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a snow breeze. The numerical analysis further demonstrated that a patchy snow cover is associated with the modification of the surface thermal fluxes and a suppression of the daytime thermally induced upslope flow.

Because the temperature of a melting snow surface is always near freezing (0°C), a temperature inversion is present above snow whenever the air is above freezing (Whiteman 2000), causing stable internal boundary layers. In the presence of turbulence generated by wind shear close to the ground, the positive temperature gradients over snow-covered surfaces typically induce a downward sensible heat flux, causing a warming of the snow surface. The daytime heating of snow-free areas leads to the development of convective boundary layers and heating of near-surface atmospheric layers. In the presence of considerably strong winds, the warm air from the bare ground is advected toward the snow, resulting in an increase of the atmospheric stability above the snow cover (Segal et al. 1991). The local advection of sensible heat (horizontal transport of sensible heat with the mean flow across a step change in surface temperature) can significantly contribute to snow ablation (Mott et al. 2011a), especially for situations with high wind velocities and strong mechanical turbulence (Mott et al. 2013). For calm to moderate wind conditions, boundary layer decoupling over snow typically causes the reduction of vertical momentum transfer between the surface and the air above. This is caused by the suppression of turbulence near the surface. For those situations, vertical boundary layer decoupling counteracts the process of local advection of sensible heat (Mott et al. 2013). The efficiency of local advection of sensible heat also appears to be a function of the snow-cover fraction. Marsh and Pomeroy (1996) developed a simple model, which is based on an empirically developed efficiency parameter for the transfer of available heat from bare to snow-covered areas as a function of snow-cover fraction. A similar approach was used by Neumann and Marsh (1998) and Pohl and Marsh (2006) in order to model the spatial variability of spring snowmelt. Liston (1995) investigated the effect of local advection of sensible heat as a function of varying snow-cover fraction by applying a numerical atmospheric model. They ran the Reynolds-averaged Navier–Stokes (RANS) model over a flat two-dimensional domain and for a variety of different snow-cover fractions. Similar to other studies, Liston (1995) could demonstrate that local advection of sensible heat increased with a decreasing patch area. In contrast to those earlier studies, this investigation aims to resolve the larger eddies (most important for turbulent mixing) through large-eddy simulation (LES) in a mesoscale framework, the Advanced Regional Prediction System (ARPS), in which turbulence produced

that is larger than the numerical grid size is fully resolved and the subgrid-scale effect is parameterized. We further use real, complex terrain and snow-cover fraction data with a coupled land surface model.

Even though advective heat transport and vertical boundary layer decoupling were shown to significantly affect the energy balance over patchy snow covers, these processes are largely neglected in energy balance and hydrological studies so far. Typically, most studies on snowmelt dynamics assume spatially constant wind speeds or a constant degree-day factor and do not account for the small-scale ($\sim 100\text{ m}$ and below) patchiness of the melting snow surface. Furthermore, most numerical studies on snowmelt are based on the constant flux layer assumption and simply integrate between air temperatures and the surface temperatures measured or modeled at one or several points in the area. That approach typically results in an overestimation of the turbulent heat flux transported toward a melting snow surface (Mott et al. 2011a) because the development of shallow internal thermal boundary layers with heights less than 3–5 m (typical measurement height for air temperature) are not captured. This investigation represents a sensitivity study with the aim to quantify the effect of a gradually decreasing snow-cover fraction on the heat exchange at the melting snow surface by considering the development of thermal internal boundary layers at each grid using an atmospheric model coupled to a land surface model. Since the simulations of turbulent fluxes are based on the Monin–Obukhov theory, very high vertical grid resolutions are applied in order to adequately capture small-scale temperature profile adaptations over the patchy land cover. We particularly distinguish between terrain-forced and thermally driven flows that evolve over different snow-cover fractions and investigate how different flow features affect the development of thermal internal boundary layers above a patchy snow cover. The main aim of this work is to show how much of the turbulent heat flux over a melting and patchy snow cover is controlled by atmospheric boundary layer (ABL) processes that can only be captured by running atmospheric models on a very high spatial resolution, which is computationally very expensive. The broader purpose of the study is to work toward a parameterization for sensible heat fluxes over patchy snow cover for simpler model approaches or for the case where grid spacing is too coarse compared to snow patch size and terrain features.

This paper is organized as follows. Section 2 presents a description of methods and data. In section 3, we present and discuss numerical results. First, we show and discuss the evolution of the atmospheric temperature profiles as a function of grid resolution, snow-cover fraction, and

FIG. 1. (a) Overview of the study site Wannengrat and (b) snow depth map obtained from ALS at time of peak accumulation in spring 2009. The red polygon marks borders of the area of interest A1 and the black polygon marks the area of interest A2. [Base map: Pixelkarte PK25 swisstopo (dv033492)].

wind forcing. Second, flow features are presented as a function of the initialized snow-cover fraction and wind forcing. Third, we discuss differences in the heat exchange at the snow surface associated with the development of thermally versus terrain-induced flow features after a given integration time. The driving mechanisms that control the heat exchange at the snow cover, such as boundary layer decoupling and local advection of sensible heat, are examined for the different flow fields. Finally, the conclusions include a brief summary of the findings.

2. Methods

a. Study site

The study site is an Alpine catchment, located in the Wannengrat area (Davos, Switzerland; Fig. 1a), where

several studies on snow depth, snowfall variability, and snow hydrology have been completed in recent years (Mott et al. 2010, 2014; Grünwald et al. 2010; Grünwald and Lehning 2011; Groot Zwaftink et al. 2011; Schirmer et al. 2011; Egli et al. 2011; Scipión et al. 2013). The elevation ranges from 1940 to 2658 m MSL and is above the local tree line. The Wannengrat area is equipped with seven automatic weather stations. The peak accumulation is typically reached in April and the complete melt of the seasonal snow cover is typically observed in June (Mott et al. 2010). Area-wide snow depths at time of peak accumulation (9 April 2009) in winter 2008/09 were measured using airborne laser scanning (ALS; Fig. 1b). For this measurement campaign, a helicopter-based technology (Skaloud et al. 2006) was used.

b. Model setup

We calculated daytime flow fields with the atmospheric prediction model ARPS using full model physics (Xue et al. 2001). The ARPS code was modified and set up in order to run the model over a patchy snow cover accounting for differences of thermal properties and roughness lengths of snow and vegetation. This model setup includes a force-restore land surface soil–vegetation model (Noilhan and Planton 1989) and a longwave and shortwave radiation package (Chou 1990, 1992; Chou and Suarez 1994) that calculates the radiative forcing at the ground surface, as well as topographic shading and longwave radiation, including radiative cooling. For the subgrid-scale turbulence parameterization, the 1.5-order turbulent kinetic energy closure scheme was used (Deardorff 1972). The time-dependent state of the land surface is predicted by the surface energy and moisture budget equations (Xue et al. 2001). Surface turbulent flux of sensible heat calculations are based on the similarity theory of Monin and Obukhov (1954), and stability functions for unstable conditions are based on Byun (1990) and for stable conditions on Deardorff (1972). We used the zero-gradient formulation for the upper boundary. For the lateral boundaries, wave-radiation open boundaries were used. To minimize boundary effects on the flow field that is further analyzed in this study, the model was run for a larger domain size of $2.9\text{ km} \times 2.8\text{ km}$ with a vertical extent of 4000 m above ground level. In the following discussion of flow field characteristics and heat exchange processes, the analyses were applied to two smaller inner domains with domain sizes of $1.3\text{ km} \times 1.2\text{ km}$ (A1) and $1.0\text{ km} \times 0.5\text{ km}$ (A2) (Fig. 1). The area A1 was chosen for all illustrated maps of the flow field and turbulent fluxes of sensible heat and for the variogram analysis. The smaller area A2 was chosen for the calculation of the average potential temperature profiles and the change of the mean sensible heat flux and potential air temperature as a function of fetch distance. The smaller area allowed us to limit the analysis of the energy balance at the patchy snow cover to those areas, which were not affected by strong topographic shading (i.e., northeast slope).

To examine the effect of grid resolution on the boundary layer development over patchy snow covers, we applied horizontal grid spacing ($\Delta x, \Delta y$) of 100, 50, 25, and 5 m over the same model domain. The vertical grid spacing Δz of the first numerical level above ground averaged over A2 was 0.4 m for the smallest horizontal grid spacing of 5 m and 12.4 m for the coarsest horizontal grid spacing of 100 m (Fig. 2). Because of vertical grid stretching, the vertical grid spacing increased with increasing numerical level. The aspect ratios $\Delta x/\Delta z$ close to the ground varied between

FIG. 2. Spatially averaged vertical grid resolutions of the first five numerical levels above ground (z_1 – z_5) for horizontal grid resolutions of 100, 50, 25, and 5 m.

12 and 8 for the different horizontal resolutions. The small values of the aspect ratios ensure a limitation of numerical errors (Chow et al. 2006), which is especially important for steep terrain (De Wekker 2002). The small time step was set to 0.001 s in order to integrate acoustic wave modes. The larger time step for model integration was set to 0.01 s.

All simulations start at 1200 UTC 8 May 2011 and were run for an integration time of 700 s. The results on boundary layer dynamics over patchy snow covers only represent the development of the boundary layer flow and the associated sensible heat exchange after the integration time of 700 s. All results are therefore a snapshot in time and do not cover the temporal variability in turbulent fluxes of heat and momentum that is connected to the larger eddies. Considerably longer integration times generate numerical stability problems likely related to insufficient vertical resolution when very stable layers develop. The study of Raderschall et al. (2008), however, could demonstrate that turbulent flow field characteristics fully developed after an integration time of about 600 s, when running idealized simulations with ARPS with similar domain size, model resolution, and the very small integration time step of 0.01 s. Although the absolute values in wind velocity and turbulent fluxes change with regard to the integration time step, the effects of the snow-cover fraction on the patterns of the flow field and the turbulent fluxes of sensible heat are assumed to be well captured by the applied methods. Recent progress in the implementation of the immersed boundary method in mesoscale models (e.g., Lundquist et al. 2012) may in the future allow atmospheric mesoscale models to simulate boundary layer dynamics over steep terrain and fine resolutions.

Although we are performing idealized simulations, we chose this specific date, because a field experiment took place in the Wannengrat area, where turbulent fluxes of sensible heat were measured above a melting snow surface (Mott et al. 2013), allowing a comparison to our simulation results. We chose noon because approximately peak radiative forcing should highlight the contrast in heating between snow-covered and snow-free patches. Since we are interested in the effect of radiative forcing on boundary layer development, we assumed clear sky conditions, which were also observed during the experiment conducted by Mott et al. (2013).

For all simulation runs, ARPS has a horizontally homogeneous initial condition. A base state profile consisting of neutral static stability was utilized. Such an initial state was chosen to allow the study to focus on cooling and heating processes controlled by the heterogeneous land cover. The potential air temperature was initialized with 300 K for all initialization cases. The potential temperature of 300 K corresponds to an air temperature of 5°C at an elevation corresponding to the 770-hPa level. These values were chosen to approximately match measurements made at the meteorological station Wannengrat 3 (WAN3) on 8 May 2011 (Fig. 1b). The model was initialized with a range of different base state wind profiles in order to separate synoptic effects from thermal effects on the flow. Note that V_{initial} defines the initial wind velocity throughout the flow field, as well as the lateral boundary forcing at the west end of the domain. The variable dd_{initial} defines the initial wind direction in degrees from north. We distinguish between two model setups: 1) the quiescent cases without synoptic wind forcing ($V_{\text{initial}} = 0 \text{ m s}^{-1}$ with $dd_{\text{initial}} = 0^\circ$) or very weak synoptic wind forcing ($V_{\text{initial}} = 0.5 \text{ m s}^{-1}$ with $dd_{\text{initial}} = 270^\circ$) and 2) forced cases with a westerly synoptic wind ($V_{\text{initial}} = 3$ and 5 m s^{-1} with $dd_{\text{initial}} = 270^\circ$). The wind direction of 270° and the range of initialized wind velocities correspond to typical northwest wind situations measured at the six permanent meteorological stations Wannengrat 1–6 (WAN1–6) on 8 May 2011 (Fig. 1). Northwest wind situations were also shown to be the predominant wind situation that mainly controls the snow-cover distribution in that area (Mott et al. 2010; Schirmer et al. 2011).

Recent investigations demonstrated that the patchiness of the mountain snow cover is largely caused by inhomogeneous deposition during the accumulation phase rather than differences in melt (Luce et al. 1998; Liston et al. 2007; Grünwald et al. 2010). The snow hydrological study of Egli et al. (2011) that was conducted at the Wannengrat area in spring 2011 showed that the development of the total snow-covered area at the catchment scale can be estimated by spatially constant melt

rates, if the initial snow distribution at the end of the accumulation season is known. Furthermore, Schirmer et al. (2011) found a very strong interannual consistency of snow depths at the time of peak accumulation at the Wannengrat area. Those findings allowed us to define the different initial snow-cover distributions, assuming a constant melting rate of snow heights measured by ALS at time of peak accumulation in spring 2009 (Mott et al. 2011a; Egli et al. 2011). Simulations were performed for six snow-cover fractions (SCF) that cover the range between complete snow cover and complete melt in that catchment in the ablation period from April to June: 100%, 65%, 50%, 37%, 23%, and 15%. These maps of snow-cover distribution were produced by applying a constant melting rate to the measured snow depths at the time of peak accumulation (Egli et al. 2011). Therefore, we subtracted 0 m (SCF = 100%), 1 m (SCF = 65%), 1.3 m (SCF = 50%), 1.6 m (SCF = 37%), 2 m (SCF = 23%), and 2.3 m (SCF = 15%) from the measured snow depth at each grid point. The resulting snow-cover distribution maps are shown in Fig. 3. The mean perimeters of snow patches decreased from 149.5 m for SCF = 65% to 67 m for SCF = 15% (Table 1). Please note that only simulations with $V_{\text{initial}} = 0$ and 3 m s^{-1} were run for all snow-cover fractions and horizontal grid spacings. For each snow-cover fraction, the soil–vegetation model was initialized using two land surface types, snow and alpine meadow. Differences of surface roughness between snow-free and snow-covered areas were considered using a roughness length z_0 of 0.005 m for snow-covered areas (Doorschot et al. 2004) and 0.01 m for snow-free areas, assuming alpine meadow with isolated obstacles (mainly rocks; WMO 2008). Since no detailed information on the initial surface temperatures of snow-covered and snow-free areas were available, the initial surface temperature was assumed to be spatially homogeneous and set to 270 K, which is the measured snow-surface temperatures at WAN3.

c. Geostatistical analysis: (Semi)variograms

We calculated variograms of the three-dimensional wind velocity at the first numerical level above the ground in order to determine the spatial continuity of the flow field over a distinct range of scales (Goodchild and Mark 1987; Mott et al. 2011b) and to see how the spatial continuity of the flow field might be affected by the snow-cover fraction and topography. Variogram analysis was performed on the wind velocity datasets to quantify the spatial correlation between pairs of samples as distances vary.

Scaling parameters were derived from the semivariance plot (Webster and Oliver 2007):

FIG. 3. Maps of snow cover initialized in ARPS for a gradually decreasing SCF of (a) 65%, (b) 50%, (c) 37%, (d) 23%, and (e) 15%. Maps of snow-covered areas are derived by applying a constant melting rate to measured snow depth map (see Fig. 1). The direction and end point of transects T1 and T2 are marked by arrowheads. Swiss coordinates are given (m).

$$\hat{\gamma}(h) = \frac{1}{2|N(h)|} \sum_{(i,j) \in N(h)} (V_j - V_i)^2, \quad (1)$$

$$D_{s,l} = 3 - \frac{\alpha_{1,2}}{2}. \quad (3)$$

where N is the number of available point pairs (i, j) of the three-dimensional wind velocity V in each distance class h , using 50 logarithmic width bins. We investigated the variance on a resolution of 5 m (lag distance, distance between distance classes), which corresponds to the horizontal resolution of the modeled flow fields.

A logarithmic least squares fit was applied to the estimated variogram in order to determine the scale break L , the slope α before and after the scale break, and the ordinal intercept β (Schirmer and Lehning 2011):

$$\log[\hat{\gamma}(h)] = \begin{cases} \alpha_1 \log(h) + \beta_1, & \text{if } \log(h) \leq \log(L) \\ \alpha_2 \log(h) + \beta_2, & \text{if } \log(h) > \log(L) \end{cases} \quad (2)$$

with the continuity constraint

$$\alpha_1 \log(L) + \beta_1 = \alpha_2 \log(L) + \beta_2.$$

Based on the slope, the fractal dimensions before (short range D_s) and after (long range D_l) the scale break were calculated following Sun et al. (2006):

The fractal dimension D is a measure of the irregularity of the field in consideration (Sun et al. 2006). In a general sense, large D_s indicates a very rough surface (i.e., small spatial continuity or coherence of the flow field) and small fractal dimensions are linked to smooth surfaces (i.e., high spatial continuity or coherence of the flow field; Burrough 1981, 1993; Klinkenberg and Goodchild 1992; Pentland 1984; Sun et al. 2006). Furthermore, D can also be applied to identify the scale at which different processes dominate (Emerson et al. 1999; Sun et al. 2006). The scale break has been identified

TABLE 1. Mean perimeter and area of snow patches for different SCFs.

SCF	Perimeter (m)	Area (m ²)
15%	67	372
23%	82	552
37%	101	921
50%	120	1359
65%	149.5	2139

FIG. 4. Spatially averaged profiles of potential air temperature that developed after an integration time of 700 s for snow-covered cells of the area A2. Profiles are shown for (a) four different horizontal grid resolutions (100, 50, 25, and 5 m, given by different line types), (b) six different initialized SCF (100%, 65%, 50%, 37%, 23%, and 15%, defined by different marker types), and (c) forced wind velocities (0, 0.5, 3, and 5 m s⁻¹, given by different colors of the line). Data points follow grid stretching and are only shown for the first 13 grid points (5-m resolution) above the land surface. Note that the number of data points is smaller for coarser grid resolution simulations.

as an indicator for a change of the dominating process at a respective scale (Deems et al. 2006; Trujillo et al. 2007). Recent studies demonstrated that the spatial characteristics of the flow field are similar to those of the snow depth distribution (Deems et al. 2006; Mott et al. 2011b; Schirmer and Lehning 2011; Trujillo et al. 2007; 2009) and are mainly governed by the roughness of the local topography (Schirmer and Lehning 2011).

3. Results and discussion

a. The evolution of the atmospheric temperature profile over patchy snow cover as a function of wind forcing, snow-cover fraction, and grid resolution

Figure 4 presents profiles of the average potential air temperature θ above snow (planar averages at each vertical level) as a function of initialized grid resolution ($\Delta x, \Delta y$), SCF, and V_{initial} .

The temperature profiles that were calculated for the quiescent case ($V_{\text{initial}} = 0$ m s⁻¹; black lines) and for ($\Delta x, \Delta y$) = 25, 50, and 100 m, respectively, feature very small deviations from the base state neutral profile (Fig. 4a). A

high grid resolution of 5 m appears to be necessary to capture the strong cooling of the atmospheric layers adjacent to the snow surface that is caused by a loss of sensible heat transferred toward the snow surface. Consequently, well-developed stable internal boundary layers (SIBLs) were formed above snow for ($\Delta x, \Delta y$) = 5 m and the air temperature became significantly cooler than the initial air temperature of 300 K. The depth of the well-developed SIBL decreased as the SCF decreased (Fig. 4b) and ranged between 17 and 27 m.

For the forced case ($V_{\text{initial}} = 3$ m s⁻¹; red lines), temperature profiles of the low-resolution simulations [($\Delta x, \Delta y$) = 50 and 100 m] also demonstrate a low sensitivity of the air temperature to the thermal conditions of the surface (Fig. 4a). Temperature profiles reveal a slight warming of the lower atmosphere with a slightly unstable stratification over the first tens of meters above the snow-covered ground. The vertical grid resolution appears to be too coarse to account for the cooling of the atmospheric layers adjacent to the snow cover induced by the loss of sensible heat toward the cooler snow surface. If the horizontal resolution was increased to

25 m, an approximately 3-m thick stable layer was simulated close to the surface with a slightly unstable layer above. A further increase of the horizontal resolution to 5 m resulted in the development of a strong temperature gradient close to the snow surface.

For a continuous snow cover (SCF = 100%), a deep stable layer evolved. For a patchy snow cover (SCF = 65%–15%), the near-surface atmosphere above snow-covered areas was heated and caused an increase in the temperature gradient above snow. We suggest that the heating of the near-surface atmosphere above snow is the result of the local advection of sensible heat from the bare ground toward the snow-covered area with the ambient wind that will be further discussed in a later section. As a consequence of the horizontal transport of sensible heat toward the snow-covered areas, shallow stable layers adjacent to the snow surface developed with depths ranging between 1 (SCF = 15%) and 2 m (SCF = 65%), overlain by pronounced convective layers (Fig. 4b). These numerical results are consistent with measurements presented by Mott et al. (2013), which suggested the occurrence of stable internal boundary layers over snow patches with depths ranging between centimeters and meters that are typically overlain by convective atmospheric layers.

The formation of those shallow SIBLs appeared to be not only sensitive to (Δx , Δy) and to SCF, but also to V_{initial} (Fig. 4c). The SIBL is deepest for the highest SCF (Fig. 4b) and for the quiescent wind case, $V_{\text{initial}} = 0 \text{ m s}^{-1}$ (Fig. 4c). A weak forcing ($V_{\text{initial}} = 0.5 \text{ m s}^{-1}$) resulted in the formation of a deep SIBL similar to those calculated for the quiescent cases. A stronger synoptic wind enhanced the increase in air temperature above snow-free and snow-covered areas. Considering that the snow-cover temperature cannot exceed its melting point, the heating of the near-surface atmospheric layers resulted in an increase in the static stability close to the snow surface. The static stability near the ground was thus highest for high wind velocities and for low SCF.

The analysis of the sensitivity of the boundary layer development to grid resolution demonstrated that a high horizontal grid resolution of 5 m with a vertical grid resolution of 0.4 m (first vertical numerical grid level) is necessary to resolve the vertical development of thermal internal boundary layers over snow patches that have mean perimeters ranging between 67 m for an SCF of 15% and 149 m for an SCF of 65%. For simulations with a horizontal grid resolution of 25 m, the mean vertical grid resolution of the first vertical numerical level is 2.3 m (Fig. 2), which appears to be too coarse to capture the shallow stable layers that develop over the small-scale snow patches. In the following discussion on boundary layer dynamics, we focus our analysis on the very high-resolution simulations.

b. The evolution of the atmospheric flow field over patchy snow cover

1) THE QUIESCENT CASE: THERMALLY DRIVEN FLOWS

Daytime atmospheric flow fields are shown for the quiescent case and for a range of different SCF in Fig. 5. The wind velocity was derived from the first numerical grid level that corresponds to a mean height of 0.4 m above the surface (Fig. 2). Simulation results show that the complex thermal wind system is characterized by small-scale anabatic and katabatic flows and a strong local variation in wind speed and direction (Fig. 5). In the absence of a synoptic wind forcing and assuming clear sky conditions, the thermally driven flow features were mainly caused by differences in the land surface temperatures as a result of the complex terrain (slope and exposition) and the land surface heterogeneity (snow–vegetation). The strong local differences in the land surface temperatures over snow-free and snow-covered areas (not shown) cause the formation of thermal internal boundary layers.

The vertical evolution of the potential air temperature and the wind velocity is shown in Fig. 6. Depending on the SCF, the transect crossed a snow-covered (Figs. 6a,b), a partly snow-covered (Figs. 6c,d), and a snow-free slope (Figs. 6e,f). Over slopes largely covered by snow (Figs. 6a–d), negative buoyancy caused the development of stable internal boundary layers (Figs. 6a,c). The temperature difference between the colder air above snow and the warmer air at the same height above snow-free areas resulted in strong daytime downslope drainage flows (katabatic winds), especially for simulations with high snow-cover fractions (Figs. 5a–d and 6b,d). In contrast, solar radiation caused surface heating over snow-free slopes (Fig. 6e). The surface got warmer than the overlying atmosphere and the sensible heat flux acted to warm the near-surface atmosphere, causing the evolution of a convective thermal internal boundary layer (CIBL) above snow-free patches. Positive buoyancy within the convective layer initiated a daytime upslope wind (Figs. 5,6f).

Numerical results reveal the strong dependence of the evolution of thermally driven flow features on the SCF (Fig. 5). For simulations with a high SCF (100%–50%), pronounced drainage flows evolved on the widely snow-covered southeast slopes. The drainage flow developed over the large snow field at the top of the slope and converged in gullies and the valley bottom leading to relatively high wind velocities at those locations (Figs. 5a–c). The wind velocity of the drainage flow increased with increasing fetch distance over the snow field (Figs. 5b, 6b)

FIG. 5. Atmospheric flow fields for the quiescent cases ($V_{\text{initial}} = 0 \text{ m s}^{-1}$) that developed after an integration time of 700 s, initialized with SCF of (a) 100%, (b) 65%, (c) 50%, (d) 37%, (e) 23%, and (f) 15%. Flow fields are shown for A1 and on a horizontal resolution of 5 m.

FIG. 6. Cross sections of (left) potential air temperature and (right) three-dimensional wind velocity that developed after an integration time of 700 s over a slope transect (see Fig. 3, T1) for the quiescent case ($V_{\text{initial}} = 0 \text{ m s}^{-1}$), initialized with SCFs of (a),(b) 65%; (c),(d) 37%; and (e),(f) 15%. The flow fields were calculated on a horizontal resolution of 5 m. The transects (T1 in this case) were chosen in the direction of the flow so that the wind vectors u and w are aligned with the slice plane. The color bar shows the three-dimensional wind velocity and potential air temperature. Note that we show only the first 13 grid points above the land surface.

and was further enhanced by the channeling effect along the valley floor. Peak wind velocities over snow of approximately 4 m s^{-1} are found where the air trajectories over snow are longest (Fig. 5b). On the southwest slopes, strong katabatic winds developed over large and isolated snow patches and also converged toward the valley.

The dominance of snow-covered areas for simulations with $\text{SCF} = 65\%$ led to a suppression of the daytime thermally induced upslope flow over the sun-exposed slope (Fig. 5b). As the snow-covered area decreased, the katabatic wind system got increasingly perturbed by small-scale upslope flows that were generated over snow-free patches (Figs. 5d–f). The evolution of upslope winds that were characterized by lower wind velocities and the attenuation of the katabatic wind system resulted in a decrease in the average wind velocity over snow patches. For simulations with low SCF of 23% and 15% (Figs. 5e,f), drainage flows over the small snow patches on southwest slopes were mostly eliminated by the positive buoyancy evolving over large snow-free areas. The distribution of patches of bare ground and snow and the associated heterogeneity of sensible heat fluxes are thus the main forcing related to the intensity of the small-scale thermal wind system.

Segal et al. (1991) concluded from their numerical analysis of flows above a uniformly distributed patchy snow cover that the thermal forcing of the daytime drainage flow is offset by the upslope flow if the snow-cover fraction is below 90%. They state, however, that a less uniform distribution of bare ground patches will lead to a higher areal fraction of snow-free areas that is required to suppress the drainage flows. This is confirmed by our study, where suppression of the drainage flows only occurred for small SCFs below 37%.

The geostatistical analysis of the spatial coherence of the wind velocity distribution for the quiescent wind case (Fig. 7a) demonstrates a rather constant short-range fractal dimension D_s for high SCF between 100% and 65%, but a strong decrease in D_s with decreasing SCF below 65% (Table 2). The low values of D_s for a high SCF of 65% and 100% ($D_s = 2.38$ and 2.39) indicate a low-frequency variation of wind velocity and a dominance of spatially coherent flow features at scales smaller than 49.6 m, at which the scale break was found. A higher short-range fractal dimension of 2.63 at $\text{SCF} = 15\%$ indicates a higher-frequency distribution of wind velocity for low SCF and, thus, less coherent flow features than for high SCF at scales smaller than the scale break of 67 m. The higher D values after the scale break (e.g., D_l for $\text{SCF} = 65\% = 2.82$; Table 2) suggest less spatially coherent flow features at larger scales independent of SCF. Thus, thermally induced flow features are spatially more coherent at smaller scales and for high

SCF. These spatially coherent flow features are katabatic winds that evolve over larger and more continuous snow patches at high SCF. The locally enhanced wind velocities caused by the katabatic wind system cause a high variance of wind velocity that is highest for the highest SCF. Thus, the development of thermally induced secondary flow features is mainly driven by the distribution of the patchy snow cover. The scale break that ranges between 67 and 49 m (Table 2) can be interpreted as the upper-scale limit at which the land surface (SCF) persistently influences the flow field.

2) THE FORCED CASE: TERRAIN-FORCED FLOWS

Terrain-forced flows developed if the flow fields were initialized with a synoptic westerly flow (Fig. 8). In the case of a coupling to a synoptic flow of 3 m s^{-1} , thermal winds as described for the quiescent case were opposed by a strong westerly flow, featuring enhanced wind velocities (Fig. 8). The terrain-forced wind system is less complex than the purely thermally driven wind system that was discussed above. The flow field is characterized by less pronounced secondary flow features that are restricted to the largest snow patches on the southeast slope for high SCF (Figs. 8a–d) and large snow-free and sun-exposed slopes on the southwest slope for very low SCF (Figs. 8e,f). Wind velocity variations were mainly caused by the local complexity of the topography. The highest wind speeds were modeled for ridges due to local speedup effects and for the pass and valley due to channeling. Flow separation zones on leeward slopes caused patterns of low wind speeds.

The wind velocities at the first numerical level above ground (about 0.4 m above the surface) are lower for higher SCFs (Fig. 8). The vertical evolution of the wind velocity and the potential air temperature along well-developed thermal internal boundary layers (Fig. 9) shows that the high static stabilities of SIBLs over snow were associated with reduced wind velocities near the ground. Above the shallow SIBL, the flow field is characteristic of the upstream conditions despite the detachment of the strong synoptic flow from the snow surface and its displacement to higher atmospheric levels. A low-level jet (LLJ) evolved above the locally developed SIBLs, featuring a wind velocity maximum several meters (8–15 m) above the ground (Fig. 9b). Spatially averaged normalized profiles of wind velocity above snow show that the magnitude of the velocity maximum in the low-level jet increased with decreasing SCF, but the height of the wind velocity maximum decreased with decreasing SCF from 15 (SCF = 65%) to 8 m (SCF = 15%; Fig. 10a). The adjacent air above snow patches features lower values of turbulent kinetic energy for high than for low SCF (Fig. 10b), indicating a stronger vertical decoupling from

FIG. 7. Semivariogram of wind velocity calculated for flow fields of the (a) quiescent ($V_{\text{initial}} = 0 \text{ m s}^{-1}$) and (b) forced cases ($V_{\text{initial}} = 3 \text{ m s}^{-1}$), initialized with a range of different SCFs. Semivariance plots were calculated for wind velocities (horizontal resolution is 5 m) within the area A1. (c) The fractal dimensions below the scale break (D_s) for all snow-cover fractions and for both cases of wind velocity initialization. Values of scale breaks and fractal dimensions are summarized in Table 2.

the upper-level flow for simulations with high SCF. For low SCF (SCF = 50% and below), the turbulence and momentum were less destroyed by negative buoyancy, leading to higher wind velocities above snow-free areas, higher wind velocity maxima of the LLJ over snow-covered areas, and less decoupling of the atmosphere adjacent to the snow cover from the synoptically induced flow aloft (Figs. 9d,f).

Variogram analysis of the spatial coherence of flow features that evolve over patchy snow covers for the forced case reveal an increase in the short-range fractal dimension with a decreasing SCF (Fig. 7b, Table 2). Flow features are thus spatially more coherent at high SCF than for low SCF. In comparison to the fractal behavior of thermally driven flow fields, the values of D_s

are generally higher than those calculated for the quiescent case (Fig. 7d, Table 2), indicating a lower spatial continuity of wind velocity if forced by a synoptic wind than for quiescent cases. Only for very small SCF (SCF below 37%), forced flow fields are characterized by longer-range flow features than for the quiescent case, which can be explained by the much smaller effect of the snow cover on the flow adjacent to the surface caused by the minor presence of the snow cover (see discussion above). The relatively small effect of the snow cover on the flow for a very small SCF of 15% also explains the very low variance for the forced flow for low SCFs. Furthermore, the lower increase of D_s values with decreasing SCF (Fig. 7c) suggests a weaker control of the snow patch size distribution on the flow patterns if

TABLE 2. Break distances and fractal dimensions.

$V_{\text{initial}} \text{ (m s}^{-1}\text{)}$	SCF	$L \text{ (m)}$	D_s	D_l
0	15%	67.3	2.63	2.92
0	23%	76.7	2.59	2.90
0	37%	58.7	2.49	2.82
0	50%	50.2	2.44	2.80
0	65%	49.6	2.38	2.82
0	100%	58.2	2.39	2.71
3	15%	33.6	2.48	2.86
3	23%	44.1	2.53	2.88
3	37%	39.0	2.51	2.85
3	50%	39.1	2.48	2.84
3	65%	30.1	2.43	2.82
3	100%	34.4	2.39	2.89

initialized by a strong synoptic wind. Only for continuous snow covers (Fig. 7c), the spatial coherence of the flow field patterns are similar for both cases, despite the smaller-scale break that is found for synoptically forced flows (Table 2). These results are similar to the results found by Mott et al. (2011b), who showed that modeled flow fields forced by a synoptic wind over a continuous snow cover are mainly governed by the terrain roughness. Furthermore, the scale breaks, ranging from 30 to 40 m, are of the same order of magnitude as those found by Mott et al. (2011b). Similar to Mott et al. (2011b), we interpret the length scale that is expressed by the scale break as the upper scale at which the roughness of the terrain and temperature of the land surface persistently influence the flow field.

c. Heat exchange processes over a patchy snow cover

1) THE SPATIAL DISTRIBUTION OF TURBULENT FLUXES OF SENSIBLE HEAT

The spatial distribution of turbulent fluxes of sensible heat that evolved with thermally driven flows (quiescent case) over a gradually decreasing SCF is shown in Fig. 11. Figure 12 presents the spatial patterns of turbulent fluxes of sensible heat (QH) that evolve with the terrain-forced flows (forced case). Positive values of the sensible heat flux indicate a downward heat flux (cooling of air above the land surface) and negative values indicate an upward heat flux (heating of air above the land surface).

Spatial patterns of daytime turbulent fluxes of sensible heat follow the patterns of available radiation energy that are governed by differences in exposition, slope angle, and the heterogeneous land surface consisting of snow-covered and snow-free areas. In addition to the radiative energy, wind velocity fluctuations due to thermally and dynamically driven flow features strongly control the spatial patterns of the turbulent fluxes of sensible heat. The

strong land-cover heterogeneity leads to the small-scale coexistence of downward and upward heat fluxes for both the quiescent and the forced cases (Figs. 11, 12).

For the quiescent case, the spatial variation in the surface heat flux is strongly coupled to the complex thermal wind system (Fig. 5). Fair weather conditions and high air temperatures induce daytime downward turbulent fluxes of sensible heat over snow-covered areas (Fig. 11). On the southeast slope, peak wind velocities, driven by the katabatic wind system, strongly enhance the turbulent sensible heat flux toward the snow cover (Fig. 11). A second pronounced area that is characterized by significant downward heat fluxes is the shaded northeast slope and the northwest slope (upper left-hand corner, Fig. 11a), where the near-surface air is cooled by the shadowing effect of the topography and downward heat fluxes are promoted by high wind velocities.

Areas characterized by upward heat fluxes that heat the air above the surface are associated with the formation of CIBLs above snow-free areas (Figs. 6, 11). The highest values of negative (upward) heat fluxes are found for small snow-free areas at the bottom of the southeast slope (high wind velocities induced by the upwind drainage flow) and at the southwest slopes (Fig. 11).

For the quiescent case, peak magnitudes of downward heat fluxes (cooling the air) are controlled by the strength of the katabatic wind system. Since the katabatic wind gradually attenuates with a decreasing SCF, downward sensible heat fluxes also decrease with a decreasing SCF. The downslope wind system on the northeast slope is independent from the SCF because the air is cooled by the strong shadowing effect of the topography on this slope. The magnitudes of the turbulent sensible heat fluxes thus remain constant for simulations of different SCF (Fig. 11). On the southwest slopes, upward heat fluxes that are associated with the upslope wind system increase with a decreasing SCF (Fig. 11). The magnitude of upward sensible heat fluxes (warming the air) at the bottom of the southwest slopes are mainly controlled by the high wind velocities of the drainage wind system induced by the upwind snow fields at higher parts of the slope.

Similar to the quiescent case, the coexistence of upward and downward heat fluxes that are calculated for the forced case ($V_{\text{initial}} = 3 \text{ m s}^{-1}$; Fig. 12) is associated with local temperature differences between the land cover and the near-surface atmosphere, but also by wind velocity variations of the terrain-forced flow (Figs. 8, 12). Both positive and negative heat fluxes are highest for those slopes, where wind velocities are highest because of the channeling effect of the flow (Fig. 12). Peak upward heat fluxes are simulated for lower SCF on the southeast slopes (Fig. 12e), when snow-free areas get larger and the

FIG. 8. As in Fig. 5, but for the forced cases ($V_{\text{initial}} = 3 \text{ m s}^{-1}$).

FIG. 9. As in Fig. 6, but for the forced case ($V_{\text{initial}} = 3 \text{ m s}^{-1}$) along transect T2 (Fig. 3) that is directed along the local wind fetch in the west–east direction so that the wind vectors u and w are aligned with the slice plane. The horizontal resolution is 5 m. Note that we show only the first 13 grid points above the land surface.

upslope winds are no longer suppressed by the secondary drainage flows over snow patches (Fig. 8f).

A comparison between modeled and measured turbulent fluxes of sensible heat at the location of the

turbulence tower (Fig. 12) is shown in Fig. 13. We compare simulated turbulent fluxes of sensible heat at the surface obtained from the forced and quiescent cases with fluxes measured with an ultrasonic anemometer at

FIG. 10. Spatially averaged profiles of (a) wind velocity and (b) turbulent kinetic energy above snow normalized by the free-stream wind velocity. Profiles are shown for an integration time of 700 s for the forced case and for SCFs of 100%–15%. The corresponding flow field was calculated on a horizontal resolution of 5 m.

0.3 m above the ground for one ablation day on 8 May 2011 (Mott et al. 2013). Note that the boundary conditions for model initialization, such as wind and temperature (not shown) were chosen to approximately match the meteorological conditions measured at the meteorological stations (WAN1–6, Fig. 1) during that day. The observed snow-cover fraction for the investigation area at the corresponding ablation day approximately matched the snow-cover fraction of 37%. Measured turbulent fluxes of sensible heat increased with wind velocity and measured fluxes ranged between 20 and 141 W m^{-2} . The peak sensible heat flux of 141 W m^{-2} was explained to be caused by a temporary increase in friction velocity (Mott et al. 2013). In comparison, simulated fluxes ranged between 14 and 28 W m^{-2} for the lower wind velocity cases and up to 95 W m^{-2} for the higher wind velocity case. While the modeled fluxes obtained from the lower wind velocity cases appear to match the measured turbulent sensible heat fluxes observed in the morning hours with calm winds, the simulated flux for the forced case appears to represent the measured heat exchange during midday and afternoon observed with strong winds that promoted

the turbulent sensible heat flux over snow. The overestimation of the turbulent heat flux for the forced case points to the fact that the model is not able to capture the strong suppression of turbulence (including sensible heat) close to the snow cover (Mott et al. 2013). The qualitative comparison at one point in the investigation area, however, indicates that modeled sensible heat fluxes at the surface are of the same order of magnitude as measured fluxes for one ablation day with similar meteorological conditions.

2) THE EFFECT OF BOUNDARY LAYER DYNAMICS ON THE SPATIALLY AVERAGED TURBULENT SENSIBLE HEAT EXCHANGE AT THE CATCHMENT SCALE: THE QUIESCENT CASE

The strong interaction between the wind system and the magnitude of the turbulent flux of sensible heat is reflected in the spatially averaged turbulent sensible heat flux and the spatially averaged air temperature (Fig. 14) that are also shown as a function of the distance to the upwind edge of snow patches (Fig. 15).

The spatially averaged turbulent sensible heat flux at the snow cover and the spatially averaged potential air

FIG. 11. Maps of QH calculated for the quiescent case ($V_{\text{initial}} = 0 \text{ m s}^{-1}$) after 700 s and for an SCF of (a) 100%, (b) 65%, (c) 50%, (d) 37%, (e) 23%, and (f) 15%. The horizontal resolution is 5 m. Please refer to Fig. 5 for the corresponding snow-cover distribution.

FIG. 12. As in Fig. 11, but for the forced case ($V_{\text{initial}} = 3 \text{ m s}^{-1}$).

FIG. 13. Measured QH acquired with an ultrasonic anemometer at a height of 0.3 m (Mott et al. 2013) and modeled QH at the snow surface are plotted against the respective wind velocity. The modeled wind velocity was derived from the first grid level above the surface at a height of 0.4 m. All values are shown for the location of the turbulence tower in the Wannengrat area (Fig. 1). Measured fluxes were acquired from 0900 to 1630 UTC 8 May 2011. Simulated surface fluxes are shown for initialized wind velocities of 0, 0.5, and 3 m s⁻¹ and for an integration time of 700 s. All simulations results are shown for an SCF of 37%, which approximately fits the SCF during the measurement period.

temperature at the first computational level above land surface were calculated for all snow-covered cells (QH_{snow} , θ_{snow}) and for all cells including snow-covered and snow-free cells (QH_{all} , θ_{all}) within the area A2. All spatially averaged values are shown for high- [$(\Delta x, \Delta y) = 5$ m; $\Delta z(1) = 0.4$ m] and coarse-resolution simulations [$(\Delta x, \Delta y) = 50$ m; $\Delta z(1) = 6$ m].

The analysis of the high-resolution simulations of the quiescent cases ($V_{\text{initial}} = 0$ and 0.5 m s⁻¹) show that QH_{snow} decreased with a decreasing SCF (Fig. 14a). The mean values of QH_{snow} are very similar for both quiescent cases, because of the same main processes driving the energy balance for both simulations (Fig. 14a). If the SCF decreased from 100% to 15%, the mean turbulent sensible heat flux over snow decreased by 44%. This positive correlation between QH_{snow} and SCF is surprising given the strong warming of the atmosphere (increase in θ_{all}) when the SCF decreased to 15% (Fig. 14c). In the absence of any secondary boundary layer processes, we would expect an increase in QH_{snow} with an increase in θ_{all} .

The main process that drives the higher QH_{snow} for high SCF and a smaller θ_{all} is the well-developed katabatic wind system that promoted the turbulent sensible heat exchange toward the snow surface because of the approximately linear dependence of the turbulent sensible heat flux on the wind velocity (Dadic et al. 2013). Furthermore, strong katabatic winds enhance the mechanical turbulence close to the surface and remove the very stable shallow layers close to the surface that typically promote boundary layer decoupling and a shut-down of turbulent fluxes. Less pronounced katabatic winds for low SCF resulted in lower wind velocities over snow and, thus, a decrease in the QH_{snow} .

The negative correlation between QH_{snow} and θ_{all} evidence a weak response of QH_{snow} to the enhanced atmospheric heating above snow-free ground for high SCF (Fig. 14c) and is thus a clear sign for the lateral decoupling (isolation) of the atmosphere adjacent to the snow cover from the ambient air temperature. For calm wind conditions, the mitigation of the advective heat flux from bare toward snow-covered ground and the development of strong katabatic wind systems facilitated the isolation of the temperature above snow patches from the warmer surrounding air temperature. Consequently, well-developed stable layers characterized by cooler air temperatures formed above snow patches. These findings are consistent with the results of field experiments presented by Mott et al. (2013) that indicated that for calm wind conditions well developed stable layers could form that isolated the air above snow from the ambient air temperature.

The effect of lateral atmospheric decoupling (isolation) on the mean energy balance is also visible in Fig. 15a, which shows the change of the θ_{snow} from the base state as a function of the distance to the upwind edge of snow patches. Simulation results of quiescent cases showed a cooling of air adjacent to the snow cover with the strongest cooling effect above the center of snow patches (Fig. 15a). As described above, low synoptic winds facilitate the lateral decoupling of the atmosphere adjacent to the snow cover from the surrounding warmer air above snow-free ground leading to a cooling of the air by the cold snow surface that was most effective in the center of respective snow patches (Fig. 15a).

The positive sign of QH_{all} for all SCF (Fig. 14b) indicates that downward heat fluxes over snow that are associated with downslope winds were stronger than the upward heat fluxes over snow-free ground that are associated with upslope winds. The QH_{all} is almost zero for a low SCF of 15%, indicating a balance between downward and upward heat fluxes when the katabatic wind system is largely suppressed by the weaker upslope wind system. These results of energy balance calculations for

FIG. 14. Spatial averages of QH_{snow} and QH_{all} (snow-covered and snow-free cells). Mean values are shown as a function of a decreasing snow-cover fraction from SCF = 100%–15% and wind forcing ($V_{initial} = 0, 0.5$, and $3 m s^{-1}$). The potential air temperature for the first level above ground that was averaged over all cells is shown as a function of SCF and wind forcing. Results are shown for simulations with horizontal grid resolution ($\Delta x, \Delta y$) of (a)–(c) 5 and (d)–(f) 50 m. All values were calculated as spatial averages over the area A2.

quiescent wind cases show that if the patchiness of a mountain snow cover is considered in atmospheric modeling, the daytime katabatic wind can become a driving factor for the energy balance at the catchment scale.

Results of the coarse-resolution simulations (Fig. 14d) do not show any relation between QH_{snow} and SCF, because katabatic wind systems were not sufficiently resolved. Results suggest that underestimating the temperature gradients close to the snow surface (Fig. 4a) and the effect of thermal wind systems resulted in a difference in QH_{snow} of up to 95% between fine (5 m) and coarse (50 m) grid resolution simulations, especially when the SCF is high (Fig. 14d). The QH_{all} is positive (downward flux) for SCF 100%–65%, reflecting the dominance of snow-covered cells and negative (upward, warming the air adjacent to the surface) for SCF 37%–15% because of the dominance of snow-free areas. Upward and downward heat fluxes balance at SCF of 50% ($QH_{all} = 0$, Fig. 14e), which indicates that no thermal wind systems affected the energy balance if simulations were run on ($\Delta x, \Delta y$) = 50 m. This has an important consequence for mesoscale simulations of flow fields above any heterogeneous land cover.

3) THE EFFECT OF BOUNDARY LAYER DYNAMICS ON THE SPATIALLY AVERAGED TURBULENT SENSIBLE HEAT EXCHANGE AT THE CATCHMENT SCALE: THE FORCED CASE

For the forced case, QH_{snow} increased when the SCF decreased from 65% to 15% and thus shows an opposite trend to the quiescent case (Fig. 14a). The QH_{snow} is directed downward (cooling the atmosphere) and increased by 68% if the snow-cover fraction decreased from 100%–15%. The decrease in QH_{snow} between SCF = 100% and 65% can be explained by the fact that areas affected by strong wind velocities got snow-free first, because strong winds not only enhanced the turbulent fluxes of sensible heat but also caused low snow deposition rates during winter. Consequently, by applying a constant melting rate to the initial snow cover (see methods section), these pixels of lowest snow depths that show at the same time peak turbulent fluxes of sensible heat were already snow-free for the SCF = 65% and did not add to the statistics anymore. The low values of QH_{snow} for high SCF (SCF = 100%–65%) that are rather similar to those calculated for the quiescent

FIG. 15. (a) Mean change of near-surface air temperature (first numerical level above ground) above all snow-covered cells within area A2 (Fig. 3) as a function of distance of the upwind (western) edge of snow patches compared to the initial temperature of 300 K modeled for flow fields initialized with synoptic wind velocities (V_{initial}) of 0 and 3 m s^{-1} and for a gradually changing snow-cover fraction from 65% to 15%. (b) Mean turbulent sensible heat flux of all snow-covered cells within area A2 as a function of distance of the upwind edge of snow patches modeled for flow fields with synoptic west wind $V_{\text{initial}} = 3 \text{ m s}^{-1}$.

case reveal the impact of the detachment of the synoptic flow at upper levels from the near-surface air (Fig. 9). Furthermore, stronger synoptic wind promoted the advective heat flux from bare ground toward snow patches that increased with a decreasing SCF because of the following process interactions: QH_{all} (Fig. 14b) was negative in sign (directed upward, heating the atmosphere) for all simulations except for a full snow coverage (SCF = 100%). The upward heat flux over the snow-free area that acted to warm the overlying atmosphere was, thus, much stronger than the downward heat flux over the snow-covered area that acted to cool the overlying atmosphere. As larger snow-free areas produced stronger positive buoyancy, QH_{all} increased with a decreasing SCF (Fig. 14b), leading to a net warming of near-surface atmosphere (Fig. 14c). The warming of the atmosphere for low SCF enhanced the downward heat flux over snow because of the increase in the local near-surface

temperature gradient over snow (Fig. 14a). At this point, the large differences of the driving forces between the quiescent and forced cases are revealed. For the quiescent case, the katabatic winds dominated the energy balance over snow and led to a peak downward heat flux for high SCF (early in the ablation season). For the forced case, a strong decrease in vertical boundary layer decoupling with decreasing SCF and the strong advective winds together with the increasing source of available energy led to peak downward heat fluxes for low SCF (late in the ablation).

The effect of local advection of sensible heat that caused the strongest near-surface atmospheric heating effect above snow at the upwind edge of snow patches is shown for the forced case in Fig. 15b. Performing a similar analysis for the quiescent case is complicated by the nonuniformity of wind directions in the coupled katabatic–anabatic system, so we present this analysis

for the forced case only. As a result of the local advection of sensible heat forced by the synoptic wind, the mean downward heat flux at the upwind edge of the snow patches was about 30% higher (Fig. 15b) than 50 m downwind and 19%–25% higher than the spatially averaged turbulent heat flux over snow. For the given wind conditions, the local advection of sensible heat had the strongest effect over a distance of about 25 m (Fig. 15b), but considerably affected the turbulent sensible heat flux at the snow cover over a fetch distance of 50 m. The fetch distances over which local advection of sensible heat was active are of the same order of magnitude as field measurements presented by Mott et al. (2011a) for similar wind conditions. They observed enhanced snow ablation rates due to local advection of sensible heat at the first 20 m downwind of the leading edges of single snow patches and ablation rates that were 30% higher than the spatially averaged ablation rate. The mean heat fluxes at the upwind edge of the snow patches increased with a decreasing SCF (Fig. 15b), which can be related to an increase in air temperature with a decreasing SCF (Fig. 15a). Following these results, synoptic winds caused a significant redistribution of sensible heat from snow-free toward snow-covered areas and thus significantly contributed to snow ablation as soon as the snow cover got patchy in spring.

The presented result on the increasing turbulent sensible heat flux with decreasing SCF is very sensitive to the clear sky assumption that allowed the strong warming of the atmosphere due to radiative heating of the bare ground and thus the local advection of sensible heat from bare ground toward the snow-covered areas. In case of cloudy conditions, the heating of the atmosphere with a decreasing SCF would be much smaller, leading to a much smaller effect of local advection of sensible heat on the mean turbulent sensible heat flux over snow.

In contrast to the fine-resolution simulations, results of the coarse-resolution simulations show rather constant values of QH_{snow} for all SCF (Fig. 14d) and no significant warming of the atmosphere when the SCF decreased from 100% to 15% (Fig. 14f). The QH_{all} is positive for high SCF and negative for low SCF (Fig. 14e), reflecting the relative contribution of snow-free and snow-covered cells to the catchment's energy balance. Thus, no local advection of sensible heat is considered in model calculations with a coarse grid resolution of 50 m. Furthermore, the magnitude of QH_{snow} is much smaller for coarser than for higher grid resolution simulations because of the smaller temperature gradient that evolved above the snow for simulations run on $(\Delta x, \Delta y) = 50$ m and $\Delta z = 6$ m. The results suggest that a high vertical resolution (Δz) of the first numerical grid level above ground of less than 1 m is required to produce strong

temperature gradients above snow and to adequately model the heat exchange between the snow cover and the adjacent atmosphere. The comparison of simulation results from coarse (50 m) to fine (5 m) grid resolutions shows a difference in QH_{snow} of 40% for high SCF and 70% for low SCF. The lower mean values of QH_{snow} for coarse grid resolutions can be explained by the inadequate representation of thermal internal boundary layers and the mitigation of advective heat fluxes. A resolution smaller than 5 m would, however, be necessary to calculate the very shallow stable layers adjacent to the snow surface, where strong atmospheric stability results in the suppression of turbulence.

4. Conclusions

In this study, we numerically investigated the boundary layer flow development at an integration time of 700 s and the associated heat exchange processes over a gradually decreasing snow cover in spring. The numerical results point to the high sensitivity of the wind system and the magnitude of the associated heat exchange over patchy snow covers to the snow-cover fraction and the wind forcing. The analysis of the sensitivity of the flow field dynamics on the grid resolution revealed that a high horizontal grid resolution of 5 m and vertical grid resolution of 0.4 m is necessary to capture the interaction between the thermal properties of the heterogeneous land cover and the ABL flow over snow patches with mean perimeters ranging between 67 and 150 m. Horizontal grid resolutions of 25 m and lower are shown to be too coarse to resolve the strong temperature gradients close to the snow surface, the evolution of small-scale thermal internal boundary layers, and of thermal wind systems above patchy snow covers at this scale. The near-surface ABL decoupling over snow patches and the local advection of sensible heat is based on an adequate representation of thermal internal boundary layers and can thus be calculated with a horizontal resolution of 5 m and less and a corresponding vertical grid resolution of 0.4 m close to the ground. What cannot yet be captured with that high model resolution is the strong suppression of turbulent sensible heat fluxes within atmospheric layers adjacent to the snow that was discussed to be the result of strong atmospheric stability at the lowest centimeters above snow (Mott et al. 2013).

The analysis of the sensitivity of high-resolution flow fields to the wind forcing demonstrates large differences between the development of the small-scale flow fields over patchy snow covers. For the quiescent case, the coexistence of snow-covered and snow-free ground induced the formation of buoyancy-driven flows (katabatic and

anabatic winds). The spatial analysis showed that the spatial persistence of the thermally driven flow system was mainly governed by the snow-cover fraction, inducing more spatially coherent flow structures early in the ablation season when the snow-cover fraction was high. Energy balance calculations demonstrated that well-developed katabatic winds exerted a major control on the energy balance at the patchy snow cover. The control of katabatic winds on the strength of downward turbulent sensible heat fluxes at the snow cover was strongest early in the ablation season when the snow-covered area was still large and katabatic winds led to strong wind velocities. Our and other simulations (e.g., Segal et al. 1991) thus showed that not only net radiation drives the surface energy exchange in a mountain valley (Ellis and Leather 1998; Rotach and Zardi 2007), but secondary flow features associated with heterogeneous surface characteristics such as patchy snow covers also have a strong impact on the magnitudes of sensible heat fluxes. Thermal wind systems had no effect on the energy balance if simulations were run for coarse horizontal grid resolution of 50 m. Simulations on coarser grid resolutions also demonstrated that the underestimation of the near-surface temperature gradient and the associated underestimation of thermal flows over patchy snow covers lead to a difference in the spatially averaged sensible heat flux over snow of up to 95% if compared to simulation on high horizontal resolution (5 m), in particular for high SCF when the katabatic winds are strongest.

In contrast, for forced cases, the atmospheric flow field that developed after an integration time of 700 s was mainly controlled by the strength of the synoptic wind and the complexity of the underlying terrain. The spatial analysis showed that the snow-cover fraction did not significantly influence the spatial characteristics of the terrain-forced flow field. In the absence of buoyancy-driven flows, the topographically modified flow and the local advection of sensible heat appeared to dominate the spatial variability of turbulent sensible heat fluxes at the snow cover. The high spatial heterogeneity of turbulent heat fluxes over patchy snow covers are attenuated by strong synoptic winds or cloudy conditions.

The numerical results clearly show that local advection of sensible heat by the ambient wind is reduced if the near-surface flow field is mainly thermally driven. In the absence of an ambient wind, deep stable layers are formed, especially if the snow-cover fraction is high, resulting in an isolation of the atmosphere over snow patches from its warmer surrounding atmosphere. Calm wind conditions involved a peak of the downward sensible heat flux early in the ablation season and a decrease in the spatially averaged downward sensible heat flux over snow with a decreasing snow-cover fraction. As a consequence of the

isolation of the atmosphere above snow induced by the katabatic wind system, the turbulent sensible heat flux above snow appeared to be insensitive to the increasing ambient air temperature with decreasing SCF.

In contrast to the strong isolation effects induced by katabatic wind systems, the synoptically induced flows strongly promote the local advection of sensible heat from bare ground toward the snow cover, resulting in the development of shallow stable layers overlain by convective layers. For the forced case, peak downward sensible heat fluxes were modeled for low SCF, when enhanced warming of the atmosphere due to solar heating promoted enhanced advective transport of sensible heat and, thus, an increase of the local temperature gradients above snow. The numerical results coincide with observations presented in Mott et al. (2013). Although, for forced cases, a considerable amount of the sensible heat advected from bare ground toward snow-covered areas was lost in upward direction, the increased atmospheric turbulence for high wind velocities resulted in a significant turbulent sensible heat flux that was transported toward the snow cover. In contrast, simulations on coarser grid resolutions suggest an underestimation of the mean turbulent sensible heat flux over snow by up to 70% for low SCF, if thermal internal boundary layers are not well presented and the local advection of sensible heat is not captured in atmospheric models.

In summary, for the quiescent case, while thermally driven upslope (anabatic) winds developed over snow-free slopes, the magnitude of the turbulent sensible heat flux over snow is mainly driven by katabatic winds that developed over snow-covered slopes. Thus, the efficiency of katabatic winds to contribute to the catchment's energy balance is largest for high SCF. In contrast, significant synoptic winds (lateral boundary forcing) and low SCF increase the importance of local advection of sensible heat for the energy balance at the snow cover and the snowmelt. For synoptic wind conditions, the advective wind and the mechanical turbulence are high enough to add a major part of the advected energy to the SIBL and to contribute to snowmelt. In mountain catchments, katabatic winds typically occur after sunset because of radiative cooling of the near-surface air. However, katabatic winds only play a significant role for the total daily snowmelt of a catchment if the snow-cover fraction is high enough to allow their development in the daytime (likely early in the season). In contrast, strong synoptic winds increase the turbulent sensible heat flux toward the snow surface whenever the air temperature is higher than the temperature of the snow surface. However, the cross-patch wind during strong synoptic forcing can only become an important mechanism that increases snowmelt if the shortwave radiation and the ambient air

temperatures are high and the snow-cover fraction is low enough (likely late in the ablation season).

Following the results from this study and Mott et al. (2013), we conclude that we can observe two kinds of ABL decoupling effects over patchy snow covers. First, the isolation effect (lateral decoupling) is most effective if katabatic winds develop over larger snow patches and advective winds are mitigated. For these wind situations, the snow cover and its melting behavior are only marginally affected by the higher ambient air temperatures that evolve above the increasingly large snow-free areas. Second, vertical boundary layer decoupling is characterized by a suppression of turbulence close to the snow surface. In our numerical studies, that kind of boundary layer decoupling is simulated if the flow is driven by synoptic winds (no katabatic winds) and the snow-cover fraction is high enough to involve a detachment of the synoptic flow to higher atmospheric levels above snow leading to a vertical decoupling of the atmosphere adjacent to the snow cover from the warmer air aloft. Field experiments showed that vertical boundary layer decoupling is typically present during calm wind conditions and in the absence of katabatic winds (Mott et al. 2013). In our numerical study, however, the development of katabatic winds over the steep terrain prohibited the suppression of turbulence close to the ground because of high near-surface wind velocities that eroded the shallow stable layers there.

The grid resolution is a critical factor for the numerical investigation of the effect of thermal internal boundary layers on the energy balance of a patchy snow cover. Although a high horizontal resolution of 5 m and a vertical resolution of 0.4 m were shown to represent the development of shallow SIBLs above snow patches well, an even higher resolution might still lead to an improved representation of atmospheric processes and the atmospheric stability within the first meter above the surface. The high sensitivity of the ABL development to grid resolution and snow-cover fraction indicate that careful attention should be paid to snow-cover distribution and the appropriate sensible heat flux parameterization applied when running mesoscale models in complex terrain with heterogeneous or patchy land cover. The proper application of grid resolution mainly depends on the mean perimeter of snow patches found within the investigation area. Our numerical results suggest that if snow patches are on the scale of the applied horizontal grid resolution, then spurious numerical results will be generated by the model. We therefore recommend to only use mesoscale atmospheric models over patchy land surfaces, if the horizontal resolution is an order of magnitude smaller than the mean perimeter of surface patches and if the vertical grid resolution is

high enough to capture the locally variable profiles below the blending height. If these constraints cannot be met, we suggest applying energy balance models including parameterization for the local advection of sensible heat over patchy snow covers that are, for example, based on concepts of advection efficiency (Marsh and Pomeroy 1996) or boundary layer integration methods (Granger et al. 2002; Essery et al. 2006).

While this is the first systematic and quantitative sensitivity study showing how a partial snow cover influences flow and temperature in the ABL and how this ABL then influences melt, the complex and rich interaction of flow features needs to be further investigated. The very interesting and surprising result of maximum melt in the case of high SCF and rather low ABL temperatures due to katabatic wind development and less melt because of snow patch decoupling and isolation for lower SCFs does certainly depend on the specific terrain–snow patch setting investigated. Also, the discovery of low-level jets in the cases with high synoptic winds due to decoupling over snow patches is a very interesting feature and can be a research theme in future work.

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