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Key Points:

- A direct comparison between lidar flow fields, ARPS simulations, and TLS snow distribution shows complex flow fields and snow distribution
- Wind lidar measurements and ARPS simulations show a leeside separation eddy for a steep mountain ridge
- Alpine3D-simulated snow distribution based on ARPS wind fields does not fully represent measured snow distribution

Supporting Information:

- Supporting Information S1

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A close-ridge small-scale atmospheric flow field and its influence on snow accumulation

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Abstract The rough, steep, and complex terrain in the alpine environment causes a variety of flow patterns such as blocking, speed-up, or flow separation, which influence precipitation, snow deposition, and ultimately snow distribution on the ground. Cloud-terrain interactions, flow-particle interactions, and snow transport affect snow accumulation patterns, but the relative importance of these processes is not fully understood, in particular, in complex mountainous terrain. A unique combination of measurements and model simulations is used in a local case study during a 2 day snowfall event to demonstrate the current understanding of snow accumulation in very steep alpine terrain. Doppler wind lidar measurements show an eddy-like structure on the leeward side of the Sattelhorn ridge (in the Dischma valley near Davos, Switzerland), which could partly be replicated by Advanced Regional Prediction System (ARPS) flow simulations. Snow deposition measurements with a terrestrial laser scanner show a complex deposition pattern, which is only partially captured by Alpine3D deposition simulations driven by the ARPS flow fields. This shows that additional processes such as avalanches may play a role or that a more refined simulation of flow or flow-particle interactions is required to fully understand snow distribution in very steep mountainous terrain.

1. Introduction

Over complex terrain, snow cover exhibits a high spatial heterogeneity and a strong temporal (seasonal and interannual) variability, both on large mountain range scales [e.g., Beniston, 1997; Blanchet *et al.*, 2009; Marty and Blanchet, 2012] and on river catchment scales [e.g., Jost *et al.*, 2007; Grünwald and Lehning, 2011]. Catchment-scale variability plays an important role, especially in regions where snow is the major source of water. Additionally, catchment-scale and local-scale variability of the snow cover influence local ecology [Wipf *et al.*, 2009] and can affect the snow cover stability with respect to avalanches [Schirmer *et al.*, 2010]. To assess and predict seasonal water storage and resources, a better understanding of processes responsible for catchment- and local-scale snow distribution variability is fundamental.

On a large scale, variability in snow accumulation is mainly determined by interaction of mountain ranges with the large-scale atmospheric circulation causing orographic precipitation [Stoelinga *et al.*, 2013]. Increased snow deposition is expected on the windward side of mountain ranges by the process of blocking and lifting [Houze, 2012]. Spatial snow accumulation structures on the catchment scale are strongly affected by precipitation altitude effects [e.g., Wastl and Zängl, 2008] and by the interaction of local topography with the flow field. Wind-induced processes influence cloud dynamics [Choulaton and Perry, 1986; Dore *et al.*, 1992; Zängl, 2008; Zängl *et al.*, 2008; Mott *et al.*, 2014], and close-ground pure particle-flow interactions have a strong influence on snow distribution in complex terrain [Choulaton and Perry, 1986; Colle, 2004; Zängl, 2008; Lehning *et al.*, 2008; Dadić *et al.*, 2010a; Mott *et al.*, 2010; Winstral *et al.*, 2013] such as preferential deposition [Lehning *et al.*, 2008]. Preferential deposition may lead to enhanced snow accumulation on the leeward side of the mountain. Strong upslope flow on the windward side of a mountain ridge reduces the local snow deposition due to reduced fall velocities, while enhanced snow deposition may occur on the foot of the windward side, where wind speeds are weak and on the leeward side of the ridge due to flow separation, weak winds, and higher concentration of snow. On the smallest scale, snow accumulation patterns are formed by drifting and blowing snow due to strong local wind gusts and depending on the small-scale terrain [Mott *et al.*, 2010; Lehning and Fierz, 2008].

Furthermore, snow accumulation in complex terrain strongly depends on elevation and also on the slope angle [Farinotti *et al.*, 2010; Kerr *et al.*, 2013]. Farinotti *et al.* [2010] found slopes steeper than 55° to be barely snow covered. Despite the importance of avalanches, snow slides, and wind for snow depletion in steep terrain, Wirz *et al.* [2011] and Sommer *et al.* [2015] can still show significant snow accumulation in very steep rock walls. The steepness and size of a mountain ridge with respect to atmospheric stability, as typically expressed through the Froude number, further strongly influence the flow patterns that impact preferential snow deposition. Following Baines [1995], for idealized conditions, three flow regimes based on slope steepness, wind speed, and atmospheric stability may be distinguished: (i) "leeside bluff body boundary-layer separation," (ii) "complete attachment," and (iii) "postwave separation." "Leeside bluff body boundary-layer separation" occurs over steep slopes, especially for weakly stably stratified atmospheres. For less steep obstacles under the same stability regime, flow separation may be completely suppressed, while for stronger stability a separation may occur farther away from the ridge, the so-called "postwave separation."

Previous studies found strong indications of preferential deposition in radar data and snow deposition patterns [Mott *et al.*, 2010, 2014]. Moreover, model simulations show reliable results supporting the mechanism of preferential deposition [e.g., Dadić *et al.*, 2010b; Mott *et al.*, 2014]. Recent results from Wang and Huang [2017] show that enhanced deposition may be on windward or lee slopes depending on surrounding topography, wind speed and atmospheric stability. However, a systematic investigation of these effects is missing. Furthermore, these advances in the process understanding are still mainly based on model simulations and the modeling of snow accumulation remains challenging.

To further investigate the processes driving snow deposition across a mountain ridge, we present a case study of a precipitation event at the Sattelhorn ridge in the Dischma valley near Davos, Switzerland. Highly resolved radial wind velocity measurements with a pulsed Doppler wind light detection and ranging (lidar) system are combined with Advanced Regional Prediction System (ARPS) flow simulations to better understand the complex flow field over the Sattelhorn ridge. Snow accumulation measurements by terrestrial laser scans (TLS) and simulated snow accumulation distribution driven by ARPS wind fields complete this unique collection of measured and simulated data. This set of data is in agreement with differences found between modeled and measured snow accumulation on steeper slopes [Mott *et al.*, 2010, 2014] but overall enables us to reveal a more complex picture of preferential deposition in very steep terrain than previously found for more gentle terrain.

2. Data and Methods

Since 2013, the Dischma Experiment (DISCHMEX), has mainly been concerned with processes controlling snow accumulation and ablation in the Dischma valley near Davos, Switzerland, to improve the prediction of seasonal snow water resources in alpine valleys. Snow accumulation and ablation patterns are monitored using terrestrial laser scans (TLSs, section 2.3), and two meteorological stations (section 2.1) have been installed to record the near-surface meteorological conditions. The experiment was augmented with Doppler wind lidar observations (section 2.2) between 20 and 30 October 2015 to resolve the fine-scale flow features in the wind field, especially in the vicinity of one main ridgeline during a precipitation event. This setup is used to highlight the complex interaction of wind and snow accumulation on a very steep rock wall.

During the lidar campaign, high pressure dominated the region from 20 to 28 October 2015 [MeteoSchweiz, 2015]. On 28 October 2015, Foehn conditions set in, followed by a cold front that reached the study area on the night of 28–29 October 2015 from the west and produced measurable snowfall (referred to as snowfall event, hereafter). TLSs were performed on 28 and 30 October 2015, i.e., prior and after the snowfall event on 29 October 2015 between 01:50 and 10:00 UTC+1 (section 2.3). Numerical simulations with ARPS were conducted for different atmospheric conditions (section 2.4) to analyze the detailed flow structure in the vicinity of the Sattelhorn ridge during the event and to force simulations with the Alpine surface processes model Alpine3D [Lehning *et al.*, 2008].

2.1. Meteorological Stations

Measurements from two meteorological stations in the upper Dischma valley are used in this study: station Dischma Moraine, located on a moraine of the Scaletta glacier at 2532 m above sea level (asl) and station Dischma Ridge at 3034 m asl on the ridge between Piz Grialetsch and Scalettahorn. Station Dischma Moraine

Figure 1. (a) Field site in the upper Dischma valley, Davos, Switzerland. Diamonds: Meteorological stations Dischma Moraine (light green) and Dischma Ridge (dark green). The Doppler wind lidar (blue dot) was deployed close to Dürnboden. The maximum extent sampled by the lidar scans is shaded in blue. (b) Plane position indicator (PPI) angles of 25° and 28° intersect with the north slope and clear the main ridgeline of the Sattelhorn, respectively (inset). The blue lines indicate the along-valley (330°, light blue) and cross-valley (227°, dark blue) azimuthal directions of the range height indicator (RHI) scans. Red triangle: Approximate position of the terrestrial laser scanner (TLS), from which the northern slopes of the Sattelhorn are scanned (solid red). The dashed red line marks the southern slopes of the Sattelhorn. The north-south profile (black line) is used to visualize the flow field over the ridge in the Advanced Regional Prediction System (ARPS) simulations (see Figure 6). Gray dots: Grid points of ARPS simulations used to calculate maximum ridge wind speed. Orange area: Area covered by ARPS and Alpine3D simulations. Yellow area: Domain used to estimate the atmospheric stability in the simulations. Basemap: pixmaps © 2017 swisstopo (5704 000 000).

is used as reference for snow depth measurements, while station Dischma Ridge provides information about wind speed and direction of the air mass entering the Dischma valley from the south (Figure 1a and supporting information S1).

To evaluate the near-surface atmospheric stability, surface temperature data from the Intercontinental Measurement and Information System (IMIS) station IMIS DAV2, located at 2561 m asl about 8 km west of the Dischma valley, and from the Weissfluhjoch research site WFJ2 [WSL Institute for Snow and Avalanche Research SLF, 2015] at 2540 m asl located 13 km to the northwest of the Dischma valley, were used. WFJ2 further provides precipitation measurements. Bulk atmospheric stability estimates during the snowfall event

were derived from potential temperature differences between station Dischma Ridge and stations IMIS DAV1 and IMIS FLU2 which are located at the Chrachenhorn at 2871 m asl and at the Flüelapass at 2404 m asl, respectively. Details for all the stations are given in supporting information S1.

2.2. Doppler Wind Lidar

A Halo Photonics Streamline scanning Doppler wind lidar was deployed on a small terrain knoll near Dürrboden in the upper Dischma catchment (Figure 1). A scan pattern consisting of six different scans was repeated every 10 min. The scans included a vertical stare, three plane position indicator (PPI) scans at elevation angles of 5°, 25° (later 28°), and 75°, and two range height indicator (RHI) scans with azimuth angles of 330° and 227° representing the main valley axis and a plane intersecting the north slope of the Sattelhorn (Figure 1). The range gate length was 24 m and radial velocities were retrieved from up to 2.2 km distance under ideal conditions. The elevation angle of the intermediate PPI scan was changed from 25° to 28° on 28 October 2015 around 16:00 UTC+1. At 25° elevation, the scan intersected with the north slope of the Sattelhorn, while at 28° the scan cleared the main ridge line (Figure 1b). Ground return degrades lidar retrievals that intersect with topography, and data from range gates in close vicinity of the surface (0–60 m) are discarded. Hourly averages of the radial velocities from the individual PPI and RHI scans were calculated and interpolated to regular grids with 25 m grid size for further analysis.

Very clean and aerosol-scarce air during the high-pressure conditions of 20–28 October 2015 did not allow for wind retrievals in the vicinity of Sattelhorn. However, successful lidar retrievals were recorded during 16 h of the event from 28 October 2015 07:00 UTC+1 to 23:00 UTC+1 (Figure 2). Thereafter, intermittent scanner problems and subsequent weak returns from the vicinity of the Sattelhorn limited the wind retrievals. Hydrometeors and aerosols were either absent along the Sattelhorn ridgeline or the lidar signal was attenuated in the near field by thick clouds and fog (see supporting information S2).

2.3. Terrestrial Laser Scanning

Snow depths on the Sattelhorn north slope, which spans from the Sattelhorn in the west to Chlein Sattelhorn in the east (Figure 1), were measured with a terrestrial laser scanner (TLS, Riegl VZ-6000) on 28 and 30 October 2015, both before and after the snowfall event (Figure 2). All data were postprocessed following the procedure by *Sommer et al.* [2015]. However, we calculate snow depth (vertically measured) instead of snow thickness (perpendicular to the terrain) to ensure comparability with modeled snow depth (see section 2.4). Net accumulation during the snowfall event is calculated as the difference between the snow depths from the two TLS scans. Similar procedures have been used in previous studies [e.g., *Grünwald et al.*, 2010; *Wirz et al.*, 2011; *Sommer et al.*, 2015], but the present study used the Riegl VZ-6000 TLS (beam divergence of 0.12 mrad), which operates at a near infrared wavelength (1064 nm), making it exceptionally well suited to measure snow-covered terrain [*Riegl*, 2015].

For visualization, the processed data are interpolated on a 2 m grid. Measurement uncertainties are both due to acquisition and postprocessing. Errors due to acquisition are larger for large ranges, large incidence angles, and for rough surfaces [*Deems et al.*, 2013] and may be on the scale of decimeters [*Sommer et al.*, 2015]. They are partially reduced by multistation adjustment [*Sommer et al.*, 2015], after which the remaining standard error of the constant surfaces is about 2 cm. More details on the TLS measurements and postprocessing are given in the supporting information S3.

2.4. Advanced Regional Prediction System and Alpine3D

Measurements with the single Doppler wind lidar provide temporally and spatially highly resolved wind observations but resolve only the radial or along-beam wind components, and not the full 3-D wind field. To gain a better understanding of the wind components resolved by the lidar (section 2.2), to address the complex nature of the flow field, and as boundary conditions for snow distribution simulations, model simulations for the upper Dischma valley were conducted with ARPS [*Xue et al.*, 2001]. A set of simulations (Table 1 and supporting information S4) for different wind speeds, wind directions, and atmospheric stability regimes is conducted to cover the flow conditions at station Dischma Ridge during the snowfall event between 28 and 30 October 2015 (Figure 2). The naming convention of the simulations is “WD_RWS,” where WD indicates the wind direction and RWS the maximum ridge wind speed.

The ARPS model setup closely follows the setup described by *Raderschall et al.* [2008], *Mott and Lehning* [2010], and *Mott et al.* [2014]. All simulations (Table 1) are run with a horizontal resolution of 25 m on a domain

Figure 2. Overview of meteorological conditions and snow depth measurements for the study period (28 October 2015 07:00 UTC+1 to 30 October 2015 09:00 UTC+1). (a) Temperature is given for the two stations Dischma Moraine (gray) and Dischma Ridge (black). Snow depth (red) is given for station Dischma Moraine. (b) Wind direction (black dots) and wind speed (blue) are given for the meteorological station Dischma Ridge. Snow accumulation periods are marked by gray shading. The main snowfall event on 29 October 2015 started around 01:50 UTC+1 and lasted until about 10:00 UTC+1 (orange lines). Cyan lines: Terrestrial laser scans before and after the snowfall event. Yellow shading: Period during which a good lidar signal was obtained from the vicinity of the Sattelhorn north slope. The time line shows the Advanced Regional Prediction System (ARPS) wind fields (Table 1) chosen for the Alpine3D simulations (black). Precipitation (calculated from snow depth changes) is given as snow water equivalent (red).

covering the upper Dischma valley (Figure 1), which has an extent of $6.25 \text{ km} \times 4.55 \text{ km}$. The topography (digital elevation model: dh25 © 2017 swisstopo (5704 000 000)) is slightly smoothed using a low-pass filter.

The simulations have 35 vertical terrain-following levels with the first level at an average height of 2.35 m. The first 100 m of the atmosphere contain 8 to 12 levels. Simulations are integrated up to between 160 and 250 s (see supporting information S4) with an integration time step of 0.01 s and an acoustic wave mode time step of 0.001 s. An intermediate integration time has been chosen to make sure turbulent flow characteristics evolve, even though they may not be fully developed [Raderschall *et al.*, 2008]. Longer integration times, however, lead to numerical instabilities caused by the highly complex terrain in the model domain. For all simulations the solar zenith angle corresponds to noon. Simulations over bare ground are run with a roughness length of 0.01 m, while for simulations with a snow cover the roughness length is 0.005 m, as discussed in Mott *et al.* [2015].

All except for three simulations are initialized with a neutrally stratified atmosphere and a near-surface potential temperature of 295 K over bare ground, as no high-resolution information on snow cover, the atmospheric stability, and surface temperatures is available for the whole simulation domain (section 2.1). Bare-ground conditions were chosen as many slopes were snow-free prior to the snowfall event. The combination of data from three meteorological stations at varying altitudes (stations Dischma Ridge, FLU2, and DAV1) suggest a slightly stable stratification (Brunt-Väisälä frequency $N_{\text{stat}} \approx 0.01 \text{ s}^{-1}$). At station WFJ2, temperature differences between the surface and the air at 4.5 m above ground of $1.3\text{--}5.3^\circ\text{C}$ suggest the temporary occurrence

Table 1. ARPS Simulations^a

Simulation	Wind Direction (m s ⁻¹)	Mean Max. Ridge Wind (m s ⁻¹)	Eddy Height (m)	Eddy Upslope Flow (m s ⁻¹)
SE_0.8	145°	0.76	134	0.5
SE_1.5	145°	1.45	50	0.4
SE_2.6	145°	2.64	8	0.7
SE_1.5stable ^b	145°	1.47	–	–
SE_3.9stable ^b	145°	3.90	–	–
S_3.2	180°	3.24	–	–
S_1.5stable ^b	180°	1.46	13–26	0.2
NE_2.3	15°	2.31		
NE_2.8	15°	2.81		
NE_3.4	15°	3.38		

^aAll simulations (except for S_1.5stable, SE_1.5, and SE_3.9stable) are initialized over bare ground for neutral conditions with a near-surface potential temperature of 295 K. Eddy height gives the vertical extent of the upslope flow in the eddy. Eddy upslope flow gives the mean wind speed in the eddy at the lowest level above ground (about 2 m). For S_1.5stable it is the wind speed at the lowest level of the eddy. Simulations initiated with northeasterly wind are only used as input for the Alpine3D simulations.

^bSimulation with a more stable near-surface atmosphere.

of a strongly stable, near-surface atmosphere. Therefore, three simulations (SE_1.5stable, SE_3.9stable, and S_1.5stable) with a more stable near-surface atmosphere were run to evaluate the impact of atmospheric stability on flow structures. SE_1.5stable and SE_3.9stable are initialized with a stability profile as determined from the meteorological stations and a snow cover above 2300 m asl, while S_1.5stable is initialized with a near-surface potential temperature of 285 K instead of 295 K. The simulations initialized with a neutral atmosphere develop slightly stable conditions near the surface (Brunt Väisälä frequency $N \approx 0.01 \text{ s}^{-1}$) while simulations SE_1.5stable, SE_3.9stable, and S_1.5stable develop a strongly stable stratification close to the ground. Considering all model levels up to the twentieth level ($\approx 580 \text{ m}$ above ground), the mean Brunt-Väisälä frequency in the area of the Sattelhorn (Figure 1) in these simulations is $N = 0.02 \text{ s}^{-1}$, $N = 0.03 \text{ s}^{-1}$, and $N = 0.03 \text{ s}^{-1}$ for S_1.5stable, SE_1.5stable, and SE_3.9stable, respectively. Regarding only model level 7 ($\approx 20 \text{ m}$ above ground) to model level 20 over the same area, all simulations have a mean Brunt-Väisälä frequency of $N \approx 0.01 \text{ s}^{-1}$, which corresponds to the Brunt-Väisälä frequency computed from station data ($N_{\text{stat}} \approx 0.01 \text{ s}^{-1}$). Given the very short integration time, the influence of radiation on the atmospheric stability and corresponding flow fields is very weak. Thus, slightly stable atmospheric conditions are likely to occur as measured surface temperatures at the reference station WFJ2 stayed below atmospheric temperatures during the period of the lidar measurements, which is likely for cloudy conditions in late October.

To classify the simulations, maximum ridge wind speeds at the lowest model level are calculated (Figure 1). Furthermore, we analyze flow conditions of the different ARPS simulations and the lidar scans using the framework of flow regimes by Baines [1995] to evaluate flow conditions. These regimes are characterized by the Froude Number (Nh_m/U) and the slope characteristic (l_d/h_m), where N is the Brunt-Väisälä frequency, h_m is the maximum height of the obstacle, l_d the downstream half length of the obstacle, and U the wind speed. For the ARPS simulations, the average Brunt-Väisälä frequency in the vicinity of the Sattelhorn (Figure 1) is calculated, and the mean maximum ridge wind speed at the lowest model level ($\approx 1.68 \text{ m}$ above ground) is used as wind speed U . For the conditions during the lidar measurements, we calculate the Brunt-Väisälä frequency from air temperatures at meteorological stations located at different elevations in the close surrounding of the Sattelhorn area. Wind speed U is taken from station Dischma Ridge. The cross section is chosen as marked in Figure 1, while the resolution for the ARPS simulations is 25 m and the resolution for the evaluation of conditions during the lidar measurements is 2 m. This framework allows for the comparison between the flow regime over the Sattelhorn in the ARPS simulations with the flow regime determined for the period of lidar measurements. Additionally, the wind components corresponding to the radial velocities measured by the lidar in the PPI scans and the RHI scans across the valley (227° azimuth) are extracted from the different simulations and are used to validate the representativeness of the ARPS simulations.

To evaluate the influence of the local flow field on snow deposition and to address the process of preferential deposition, the highly resolved ARPS mean flow fields are used as input for the alpine surface processes model Alpine3D [Lehning *et al.*, 2008] (see supporting information S4), run with a horizontal resolution of 25 m on the same domain as the ARPS simulations (Figure 1). Because of the very complex topography, numerical instabilities in the ARPS simulations already occur after intermediate integration times. Thus, ARPS simulations, and consequently Alpine3D simulations, are restricted to a resolution of 25 m. As mean flow fields from ARPS are not available for every time step of Alpine3D (every hour), the use of flow fields representing several time steps saved computational time, as described in, e.g., Mott and Lehning [2010] and Groot Zwaafink *et al.* [2011]. Overall, the Alpine3D simulations were forced by lower wind speeds at the Sattelhorn ridge than observed at station Dischma Ridge (Figure 2 and supporting information S5), because the Sattelhorn ridge is not as exposed as the station Dischma Ridge and ARPS simulations with higher wind speeds show numerical instabilities. Precipitation is based on snow depth measurements of the station Dischma Moraine (assuming a snow density of 100 kg m^{-3} , approximately corresponding to the density calculated in the Alpine3D simulations) during the main snowfall period. The same snowfall is applied on the whole domain, which is relatively small, and therefore, this assumption should be reasonable. The two periods with increasing snow depth (Figure 2) later on 29 October 2015 are likely due to snow redistribution, as no snowfall is registered at the snow gauge on Weissfluhjoch (WFJ2). During the second period, highly variable winds with wind gusts $>4 \text{ m s}^{-1}$ are registered at both stations Dischma Ridge (Figure 2) and Dischma Moraine (not shown). Wind speeds during the first period are lower at both stations. Missing data in the wind measurements at station Dischma Moraine could, however, indicate rimed instruments, which might result in an underestimation of wind speeds. Additionally, some fluctuation in the snow measurements are likely due to measurement errors.

To compare snow deposition simulations to measured snow depth, a simulation is run with the full snow drift module of Alpine3D (using the parameterization for saltation developed by Doorschot and Lehning [2002]). An additional Alpine3D simulation without snow drift (as in Mott *et al.* [2014]) is performed to distinguish pure snow deposition due to precipitation (i.e., preferential deposition) from snow redistribution. In the snow drift module of Alpine3D, snow transport by suspension and preferential deposition is modeled based on the 3-D wind field calculated by ARPS. Saltation serves as the lower boundary condition of the suspension layer, i.e., snow needs to be entrained from the ground by saltation to potentially enter the suspension layer.

3. Results and Discussion

First, we give a general overview of the meteorological conditions in the upper Dischma valley during the snowfall event on 28–30 October 2015 (section 3.1.1), followed by an analysis of the radial velocity patterns in the lidar planes (section 3.1.2) and a discussion in the light of 3-D flow fields from the ARPS simulations (section 3.1.3). Furthermore, we discuss the flow over Sattelhorn as a stably stratified flow over an obstacle (section 3.1.4). Corresponding snow accumulation maps are introduced (sections 3.2.1 and 3.2.2) and the relationship of snow accumulation patterns and the flow field is discussed (section 3.2.3).

3.1. The Flow Field in the Vicinity of the Sattelhorn

3.1.1. Flow Conditions and Precipitation Measured by Meteorological Stations

The mean wind direction at the ridge (station Dischma Ridge) in the upper Dischma valley was south to south-east for the first 21 h of the event (Figure 2). Ridgetop wind speeds began to weaken around midnight (00:00 UTC+1) of 28 to 29 October 2015, turned to a northerly direction between 04:00 and 06:00 UTC+1, and back to southeasterly flow on 29 October 2015 around 18:00 UTC+1. The air temperature at 2532 m asl dropped below 0°C shortly after midnight on 29 October 2015 (station Dischma Moraine), while ridgetop temperatures at 3034 m asl remained below 0°C (station Dischma Ridge). Precipitation measurements at the meteorological station WFJ2 at the Weissfluhjoch show the onset of snowfall at 23:00 UTC+1 on 28 October 2015 which lasted until 08:00 UTC+1 on 29 October 2015 (not shown). Snow depth measurements at station Dischma Moraine in the upper Dischma basin indicate that snow accumulation started around 01:50 UTC+1 on 29 October 2015 and lasted until about 10:00 UTC+1 on 29 October 2015 (Figure 2). The difference in precipitation onset times for Weissfluhjoch and the upper Dischma valley may be because the stations are located in complex terrain separated by a distance of about 13 km or due to above freezing temperatures at the station Dischma Moraine until around midnight on 29 October 2015.

3.1.2. Radial Velocity Patterns in the Lidar Scans and the ARPS Simulations

Radial velocity fields from the lidar PPI scans (Figure 3) and velocity azimuth display (VAD) retrievals of the horizontal winds (see supporting information S2) on 28 October 2015 between 07:00 and 23:00 UTC+1 confirm

Figure 3. Hourly mean (a) 14:00–15:00 UTC+1 and (b) 17:00–18:00 UTC+1 radial velocity field in the upper Dischma valley from plane position indicator (PPI) Doppler wind lidar scans on 28 October 2015. The PPI angles are 25° (Figure 3a) and 28° (Figure 3b). PPI radial velocity extracted from the Advanced Regional Prediction System (ARPS) simulation SE_1.5 for the PPI angles (c) 25° and (d) 28°. Positive radial velocity (red) shows flow away from the lidar and negative radial velocity (blue) shows flow toward the lidar. Flow field (wind speed shaded) on the lowest model level over the Sattelhorn simulated by ARPS for simulation (e) SE_1.5 and (f) SE_2.6. The black dot marks the position of the lidar. Sattelhorn and Chlein Sattelhorn are marked by a triangle and diamond, respectively. The scan direction of the range height indicator (RHI) scan is marked by the black line. Contour lines in Figures 3a and 3b: dh25 © 2017 swisstopo (5740 000 000).

the prevailing southerly wind direction. PPI scans at an elevation angle of 25° show a radial wind component that is directed away from the lidar on the leeward (northern) side of the Sattelhorn ridge line (Figure 3a), while at a higher elevation angle of 28° the cross-barrier flow component over the Sattelhorn is visible (Figure 3b). Radial velocities from the RHI scans confirm this pattern (Figure 4). Hourly mean radial lidar velocity fields reveal a very persistent flow structure, with a near-surface flow away from the lidar along the upper slopes of the Sattelhorn, while the flow above the ridgetop is directed toward the lidar.

This eddy-like structure develops for southerly to southeasterly wind directions and ridgetop wind speeds between 4 and 10 m s⁻¹ and is observed over a period of 16 h. Maximum wind speeds in the upslope flow toward the ridge are between 4 and 10 m s⁻¹ and thus of the same magnitude as the ridge level flow speeds. The horizontal extent of the eddy-like structure typically varies between 400 and 1000 m (Figure 5),

Figure 4. Hourly mean radial velocity fields over the Sattelhorn in the range height indicator (RHI) Doppler wind lidar scan on 28 October 2015. Positive radial velocity shows flow away from the lidar and negative radial velocity shows flow toward the lidar. Times are in UTC+1. Topography: dhm25 © 2017 swisstopo (5740 000 000).

Figure 5. Blue bars show estimated horizontal extent of the eddy-like structure (a) for the hourly mean Doppler wind lidar range height indicator (RHI) radial velocity across the Sattelhorn and (b) radial velocity from Advanced Regional Prediction System (ARPS) simulations. Several blue bars indicate a splitting of the eddy-like structure in several cells. Eddy-like structures are shown for the time period 07:00 UTC+1 till 23:00 UTC+1 on 28 October 2015 (Figure 5a) and for the ARPS simulations (Table 1) with different ridge wind speeds and wind directions (Figure 5b). Wind speeds at the ridge are given as a black curve. In Figure 5a, wind speeds measured at the meteorological station Dischma Ridge (Figure 1) are plotted. Periods with constant southerly (southeasterly) wind direction are marked with a dark (light) green bar. Red bars in Figure 5b show the eddy size defined from a north to south cross section through the Sattelhorn (dashed black profile, Figure 1). Topography: dhm25 © 2017 swisstopo (5740 000 000).

Figure 6. Flow field over the Sattelhorn simulated by the Advanced Regional Prediction System (ARPS) for the two simulations (a, b) SE_1.5 and (c, d) SE_2.6. Figures 6a and 6c show the simulated radial velocities that would be recorded with a RHI scan (azimuth angle of 227°) of the Doppler wind lidar. The red triangle shows the position of the lidar. Figures 6b and 6d show a cross section of the north to south flow field for the top 500 m over the Sattelhorn ridge, where a leeside eddy forms in some of the simulations. The location of the cross section is marked in Figure 1.

but periods exist where the structure seems to be split into two or more eddy-like cells. The vertical extent of the eddy-like structure reaches up to about 200 m above ground.

Due to the alignment of the RHI scans relative to the Sattelhorn ridge (Figure 1), the observed structure in the radial flow field may either represent a leeside eddy or result from a strong easterly along-slope flow at elevations below the Sattelhorn ridgeline undercutting the cross-barrier flow. With observations from a single lidar, the two patterns cannot fully be distinguished and will hereafter be referred to as eddy-like structure.

Since our simulation setup is based on many assumptions (section 2.4) and is not covering the full range of measured ridge wind speeds due to numerical instabilities arising for high wind velocity cases, we concentrate the comparison between simulated and observed flow fields on the mean flow field characteristics of the flow fields showing most agreement in the radial velocity fields compared to the lidar retrievals.

The characteristics of the PPI lidar flow field (Figures 3a and 3b) is well represented by the radial flow field in the PPI planes extracted from the ARPS simulations (Figures 3c and 3d). The flow away from the lidar on the leeward side of Sattelhorn at a PPI angle of 25° is most pronounced in SE_1.5, while it is missing, very small, or connected to the flow away from the lidar to the northwest of the lidar in the other simulations. The ARPS-based radial wind fields corresponding to the RHI scans of the lidar (Figures 6a–6c) show similar eddy-like structures as observed in the RHI retrievals (Figure 4). For the analysis of these synthetic radial velocity cross sections, eddy-like structures are defined by the occurrence of a change in sign of the radial velocity (black line in Figure 6) in the vicinity of the surface. Most ARPS simulations show smaller (length and especially height) eddy-like structures than those identified in the lidar data (Figure 5). Furthermore, in the ARPS simulations initiated with southeasterly winds, the eddy-like structures tend to become smaller as the ridge wind speeds increase, except for simulations with a strongly stable near-surface atmosphere. The lidar observations did not show such a relationship between wind speed and the size of the eddy-like structures. The length and vertical extent of the eddy-like structure is best represented by SE_1.5, while for stronger wind speeds the vertical extent of the eddy-like structure is much better reproduced by SE_3.9stable compared to SE_2.6. The simulations SE_1.5stable, SE_3.9stable, and S_1.5stable show strong downslope flow

in the lowest model levels, which is likely thermally induced due to a layer of very cold air in the lowest 20 m. Due to limitations in the lidar retrievals (topographic shielding, ground returns), near-surface features (up to 0–60 m above ground), such as these drainage flows, cannot be resolved.

3.1.3. Radial Velocity Patterns in Context of the 3-D Flow Field

To put the radial velocity patterns in the context of the 3-D flow field, we concentrate on the simulations with the best representation of the observed radial flow patterns. A more complete description of the flow fields of all simulations is given in supporting information S6.

Higher-level flow in the ARPS simulations is from a southerly to southeasterly direction, which corresponds to the prevailing wind direction measured at station Dischma Ridge during the lidar retrievals. The flow at the lowest atmospheric level in the ARPS simulations initiated with southerly or southeasterly flow (Figures 3e and 3f) confirms a flow across the Sattelhorn ridgeline. Additionally, simulations initiated with southeasterly winds show flow blocking upwind of the Sattelhorn ridge, forcing a bending of the flow around the Chlein Sattelhorn, which initiates a near-surface along-ridge flow on the leeward side of the Sattelhorn ridgeline. This near-surface flow, while very weak, corresponds to the radial velocities directed away from the lidar on the leeward slopes of Sattelhorn. On the other hand, some simulations are able to reproduce the formation of a leeside eddy behind the east-west oriented Sattelhorn ridge (Figure 6b). In all simulations initiated with southeasterly winds and a neutral atmosphere, the eddy-like structure found in the radial velocities, consists of an along-slope flow component and a leeside eddy, which is smaller for stronger wind speeds and has a small vertical extent (Table 1 and Figure 4). Simulations with a strongly stable near-surface atmosphere show an eddy-like structure in the radial velocity field above the close-ground downslope flows. For S_1.5stable the eddy-like structure is very small and mainly due to a leeside eddy, as no along-slope flow develops. For SE_1.5stable and SE_3.9stable, a larger eddy-like structure develops with a vertical extent of about 80 and 100 m, respectively (above about 27 m above ground), which only consists of an along-slope flow component and has a vertical extent that is comparable to the extent of the eddy-like structure in simulation SE_2.6 but slightly shifted downslope (Figure 5). The larger vertical extent of the eddy-like structure compared to simulation SE_2.6 (in which the vertical extent is about 30 m) occurs because the bending of the flow around Chlein Sattelhorn reaches a higher vertical extent for the simulation with a more stable atmosphere, which is typical for a strongly stably stratified atmosphere, as more stably stratified air is more resistant to vertical motions and thus favors the flow around obstacles [Whiteman, 2000].

Besides showing lower wind speeds compared to the lidar flow fields, the ARPS simulations are all (except for SE_1.5stable and SE_3.9stable) initiated with bare ground for the whole domain, even though meteorological stations and photographs indicate a snow cover in some areas of the simulation domain. Furthermore, given the short integration time of the simulations (section 2.4), eddy size and location are likely to change for longer integration times. Additionally, the real terrain is more complex compared to the ARPS topography. All these factors may contribute to the fact that the model is not exactly able to represent the flow field in the lidar scans.

Overall, our simulations confirm the occurrence of flow separation on the leeward side of Sattelhorn. The flow separation and development of a leeside eddy appears to be supported by along-slope flows, which were previously found to occur in complex alpine terrain [Hug *et al.*, 2005; Wirz *et al.*, 2011]. The interaction with the cross-ridge flow and the along-slope flow may favor or strengthen a leeside eddy. The along-slope flow may further be forced to lift at the western edge of the Sattelhorn ridge enhancing this interaction. This possible interaction agrees well with highly resolved flow simulations over an idealized pyramid-shaped mountain by Voigt and Wirth [2013], who find a leeside bow vortex, which is built and maintained by flow over and around the pyramid. Based on the above analysis, the flow structure observed in the lidar scans is best represented by simulation SE_1.5, while the vertical extent of the eddy-like structure is better represented by simulations SE_1.5stable and SE_3.9stable. Additionally, a leeside eddy forms in the lee of Sattelhorn in simulation S_1.5stable but not in SE_1.5stable. This shows that the small-scale flow field in the lee of the Sattelhorn ridge is strongly sensitive to both wind direction and stability, as already found for the less steep and more ideal area of Gaudergrat [Hug *et al.*, 2005].

3.1.4. Leeside Flow Regimes

While the flow regime concept by Baines [1995] is developed for a 2-D flow over an obstacle with ideal shape, the framework is used here to attempt a classification of the observed flow regimes in the ARPS simulations and the lidar retrievals (section 2.4). The concept may thus not exactly be applicable to our situations of flow across complex 3-D topography. Baines [1995] states that the boundaries between the regimes are

Figure 7. Cross sections from the Sattelhorn ridge to Dischma valley (Figure 1) for (a) ARPS topography and (b) real topography. Two different slopes are shown: *upper slope* (light blue/orange) and the *entire slope* (dark blue/dark red). (c) ARPS simulations and lidar measurements in the classification scheme of Baines [1995] developed for 2-D stratified flow over an obstacle. Southeasterly (southerly) flows are marked with blue dots (triangles) and are scaled by wind speed. Filled (open) markers are for simulations with a slightly stable (strongly stable near-surface) atmosphere. The red bars show the position in the diagram for the lidar measurements spanning the wind speeds observed. The dashed lines mark approximately the boundaries as defined by Baines [1995]. Real topography: dh25 © 2017 swisstopo (5740 000 000).

only approximate and their exact position strongly depends on the shape of the obstacle. Thus, we perform a classification based on two different definitions of the downstream half width of the terrain obstacle (i.e., the slope). The first classification takes into account the *entire slope* (i.e., from the ridge top down to the valley floor), and the second takes into account the *upper slope* (i.e., from the ridge top to the first terrain step, Figures 7a and 7b).

Depending on the definition of the slope, the flow fields fall into different regimes (Figure 7c). For the *entire slope*, all ARPS simulations are in the regime of “postwave separation.” For the *upper slope*, the hill is steeper and thus the flow in the simulations is close to the transition between the three regimes. Based on the observations, the flow regime, during which the eddy-like structure is observed in the lidar retrievals, stretches over the regimes “complete attachment” and “postwave separation” when taking into account the *entire slope* and falls under the regime of “leeside bluff body boundary-layer separation” when regarding the *upper slope* (Figure 7).

As the eddy-like structure observed in the lidar retrievals is mainly located in the area of the *upper slope* and develops directly at the ridge (Figure 4), it is likely a “leeside bluff body boundary-layer separation.” This agrees well with the strong wind speeds observed during the lidar measurements and the classification regarding the *upper slope*. On the other hand, the flow structure observed in the ARPS simulations is mainly identified as “postwave separation,” which is in agreement with the separation starting about 100 m downstream of the ridge (Figure 6b). Additionally, a change in wind direction on the leeward side of the Sattelhorn ridge, could be a sign of a developing wave. The main reason for this, and for the compression of the eddy for larger

wind speeds, are the lower ridge wind speeds compared to the wind speeds in the lidar retrievals. However, simulations with higher wind speeds are not available as they become numerically unstable (section 2.4). Steeper slopes of the real terrain compared to ARPS topography slightly enhance this effect.

Given the large vertical extent of the eddy-like structure and the classification in the *Baines* [1995] diagram, the eddy-like structure observed in the lidar scans is likely a combination of an along-slope flow and a “leeside bluff body boundary-layer separation” eddy. Thus, the eddy-like structure likely develops in an atmosphere stable enough to favor along-slope flow up to about 100–200 m above ground but weak enough for a leeside eddy to develop. A “leeside bluff body boundary-layer separation” eddy may further be favored by the stronger wind speeds in reality compared to our ARPS simulations (section 3.1.4). Additionally, the wind direction, which is fluctuating between about 130° and 200° may play a crucial role. The eddy in some of the ARPS simulations is, however, more likely to be a “postwave separation” eddy.

3.2. Snow Accumulation on Lee Slopes

3.2.1. Snow Depth

At the meteorological station Dischma Moraine, an increase in snow depth of 6.0 cm is measured between 28 October 2015 07:50 UTC+1 and 30 October 2015 09:10 UTC+1 (i.e., in the time period between the two TLS measurements, Figure 2). During the snowfall event (29 October 2015 01:50–10:00 UTC+1), an increase in snow depth of 13.67 cm is measured at station Dischma Moraine, while cumulative snow accumulation is 20.04 cm. Decreases in snow depth during the snowfall period (Figure 2) may be due to measurement errors, settling or snow drift. Station Dischma Moraine lies just outside our Alpine3D simulation area. However, modeled snow depth changes in the vicinity of station Dischma Moraine are between 5 cm and 10 cm in the same period and, therefore, the observation is within the range of the simulations. The average new snow depth over the TLS area is about 3.1 cm with a standard deviation (σ) of 27 cm, which is on the same order of magnitude as the standard deviation of TLS snow depth measurements in *Wirz et al.* [2011]. This is, however, less than the modeled mean snow accumulation in the TLS area on the north slope of the Sattelhorn (preferential deposition only: 16.8 cm, $\sigma = 2.6$ cm; including wind drift: 16.9 cm, $\sigma = 2.7$ cm).

The uncertainty of the TLS snow depth measurements is as high as the average snow depth change (section 2.3). Additionally, the input of snow precipitation in the Alpine3D simulations is a rough estimation based on snow depth measurements. However, snowfall is likely to be overestimated, as settling, snow drift, and measurement errors cannot be distinguished in the observations. Thus, we stick to the maximum possible precipitation used as input for Alpine3D. Compared to the snow precipitation at the reference station WFJ2 (9.4 mm snow water equivalent, SWE), the estimated snowfall is large. However, a difference in snow accumulation is reasonable, as the WFJ2 site is located about 13 km farther north. Thus, we concentrate our discussion on the relative distribution. There may be a loss of mass out of the accumulation area by avalanches, which is not modeled in Alpine3D (section 2.4). Additionally, enhanced snow accumulation on the north slope of the Sattelhorn in the Alpine3D simulations might be due to missing inertia in the model, meaning that particles travel too strictly with the flow in the model. This may lead to deposition zones either on the windward slope (i.e., on the Sattelhorn south slope (Figure 1a) for southerly winds) or farther downstream on the leeward slopes, depending on topography, advection, and atmospheric stability [*Wang and Huang*, 2017].

3.2.2. Snow Accumulation Patterns

TLS new snow accumulation of the snowfall event on 29 October 2015 shows a ridge-to-valley decrease in new snow depth and a pattern of larger accumulation at the eastern and western edge of the Sattelhorn north slope (Figure 8a). On top of this, a pattern of ridge-to-valley stripes is observed. The pattern of stripes with reduced snow accumulation (gray) may have two causes. On one hand, it may be a pattern of small avalanches and snow slides. On the other hand, it may be a sign of cross-slope loading, i.e., filled gullies and chutes on the leeward side of terrain features and eroded areas due to along-slope winds (Figure 8c). A sign for avalanches and snow slides is enhanced snow accumulation at the bottom of the stripes with reduced snow accumulation. This is visible for many of the stripes with reduced accumulation. On the other hand, for cross-slope loading no distinct accumulation is expected at the slope toe. Some of the ablation bands show parallel accumulation on the leeward slopes of secondary features, which might be a sign of cross-slope loading. Cross-slope loading may further enhance the possibility of avalanches, where snow is accumulated excessively. No such stripes are seen on Alpine3D snow accumulation maps (Figure 8b). On one hand, the model does not simulate avalanches. Simulating avalanches by gravitational models [*Bernhardt et al.*, 2012; *Warscher et al.*, 2013] might thus be a valuable extension in future studies. On the other hand, cross-slope loading structures are mainly driven by snow drift processes that are not captured well by the model with

Figure 8. (a) Snow depth changes (cm) between 28 and 30 October 2015 measured by TLS (resolution 2 m). (b) Snow accumulation (cm) in the area of the Sattelhorn from the Alpine3D simulation without snow drift (resolution 25 m) for the same period as in Figure 8a. (c) Detailed view of the area marked by an orange square in Figure 8a illustrating signs of an avalanche and of cross-slope loading. Positive snow depth changes are given by green to blue colors. Negative snow depth changes are given in gray. The red contour shows the area of the Sattelhorn north slope, which is covered by the terrestrial laser scan (TLS) measurements. The dashed red contour marks the Sattelhorn south slope. The area of the Sattelhorn north and south slopes (red) are marked in Figure 1. Contour lines: dh25 © 2017 swisstopo (5740 000 000).

a 25 m resolution [Mott and Lehning, 2010]. The ridge-to-valley gradient in snow accumulation on the TLS snow map is, however, visible in both Alpine3D simulations for 30 October 2015 at 09:00 UTC+1. A pattern of enhanced snow accumulation at the eastern and western edges of the Sattelhorn north slope is visible, too. On top of this, there is a distinct pattern of enhanced snow accumulation on the north slope of Sattelhorn, which is not visible in the TLS snow depth change map. This pattern is slightly more distinctive in the simulation with snow drift (Text S7, as differences are almost invisible). The ridge-to-valley gradient most likely is an elevation effect, as temperatures are only slightly below zero (Figure 2). The west to east pattern is most likely defined by the terrain shape and may be a sign of strong along-slope winds.

3.2.3. The Local Flow Field and Snow Accumulation

Based on snow accumulation periods at the station Dischma Moraine, 31–81% of the snowfall on 29 October between 01:50 and 10:00 UTC+1 occurred during southerly winds (measured at station Dischma Ridge, Figure 2). The large uncertainty comes from missing wind measurements on 29 October 2015 between 04:00 and 06:00 UTC+1, during which period 50% of the precipitation has fallen, when regarding the total snow depth changes. The last period of snow accumulation at station Dischma Moraine was during relatively strong southerly winds (mean wind speed 6.3 m s^{-1}). However, this is likely a sign of snow drift due to strong southerly winds without snowfall. Overall, the northern slope of the Sattelhorn cannot be judged as pure lee or windward slope during this snowfall event. In the Alpine3D simulations snowfall during the period of missing wind measurements is treated as northeasterly wind, because lidar measurements indicate that the wind has already turned. Thus, in the Alpine3D simulations (where SWE is calculated from incremental snow accumulation), only 26% of the snowfall is for southerly flow.

The analysis of the flow field for southerly winds gives evidence for flow separation on the leeward side (north slope) of Sattelhorn (section 3.1.2), where based on the theory of preferential deposition [Lehning et al., 2008], an enhanced snow accumulation is expected. Even though in the Alpine3D simulations only 26% of the snowfall was during southerly winds, a distinct snow accumulation on the north slope of Sattelhorn is observed

(Figure 8b), which is likely a sign of preferential deposition. The signs of preferential deposition on the Sattelhorn south slope (Figure 2) are weaker in the Alpine3D simulations (Figure 8b), even though 74% of the snowfall is simulated during northerly winds. The patterns are almost erased by 30 October 2015 09:00 UTC+1 (Figure 8b), mainly because strong incoming solar radiation leads to strong settling and melt on the southern slope (see supporting information S7). Furthermore, snow accumulation due to preferential deposition on the windward side would be expected in the area of the slope toe due to blocking (i.e., outside of the area measured by TLS). Thus, we mainly focus on the interaction of the flow field with snow accumulation during southerly winds, which are furthermore stronger than the northerly winds.

The patterns found in the Alpine3D simulations and the lack of small-scale structures like a cornice and other patterns due to snow drift are generally in agreement with results in *Mott and Lehning* [2010], who show that the influence of snow drift on snow deposition for a grid resolution of 25 m is rather weak and a resolution of 10 m or less is needed for small-scale structures to develop. Unfortunately, due to the complexity of the terrain and size of the domain, our ARPS and therefore Alpine3D simulations are restricted to 25 m resolution.

Slope-scale snow accumulation patterns on the north slope of Sattelhorn measured by TLS show some agreement compared to Alpine3D simulations except for the band of enhanced snow accumulation, which is missing in TLS snow accumulation (section 3.2.2), i.e., a ridge to valley decrease in snow depth and a pattern of larger snow accumulation at the eastern and western edge of the Sattelhorn ridge. One reason for the missing enhanced snow accumulation by preferential deposition in the TLS snow accumulation map might be the steepness of the Sattelhorn north slope (average slope angle: 42.7° , maximum slope angle: 79.7° , with steepest slopes in the area of preferential deposition in the Alpine3D simulations). As very steep slopes cannot keep as much snow as less steep slopes, some signal of preferential deposition might be removed by avalanches, which are likely the main process of snow redistribution [*Sommer et al.*, 2015]. Furthermore, the band of enhanced snow accumulation on the north slope of Sattelhorn occurs where a “postwave separation” eddy occurs. Strong winds at the ridge down to about 100 m away from the ridge may prevent snow from depositing or may even erode snow and transport it to the area of the eddy where wind speeds are strongly reduced. In contrast, for the situation observed by lidar, it is more likely to have a “leeside bluff body boundary-layer separation” eddy. Thus, snow may be transported over the ridge to the lower end of the eddy-like structure (i.e., outside of the area covered by TLS), as winds are separating directly at the ridge. In the area of the eddy-like structure, erosion of snow by strong back flow in the eddy may add to a reduction of snow accumulation. Additionally, processes leading to enhanced snow accumulation on the windward side of the slope [*Wang and Huang*, 2017] or a transport of particles farther downwind, which are not captured in the model, such as particle inertia, could further add to a missing sign of preferential deposition.

After the main snowfall event some snow accumulation occurs at the station Dischma Moraine again later between 29 October 2015 20:20 and 30 October 2015 00:30 during relatively strong southerly winds. This is most likely a sign of snow redistribution. As wind conditions during this period are very similar to wind conditions when the eddy-like structure is observed, it is likely that during this period an eddy-like structure developed in the vicinity of the Sattelhorn north slope, which was likely stronger than the eddy-like structure in the ARPS simulations and thus used to force Alpine3D. This indicates that either a stronger leeside eddy builds during the event compared to ARPS or along-slope winds are stronger than in the ARPS simulations (section 3.1.2). Strong winds in an eddy-like structure may thus cause redistribution of snow, either strengthening cross loading of the slope, or transport snow to places where winds are weaker. This might be an additional reason why the signal of preferential deposition is further diluted in reality compared to the simulations.

4. Conclusion and Outlook

A unique new data set from the Dischma Experiment (DISCHMEX) combines high-resolution Doppler wind lidar measurements during a snowfall event with TLS-based snow accumulation retrievals. The complex interaction between cross-barrier flow and snow deposition on the northern slope of the Sattelhorn, an alpine mountain ridge close to Davos, Switzerland, is evaluated during the 28–29 October 2015 snowfall event. Lidar and TLS observations are combined with numerical simulations from ARPS and Alpine3D to investigate the detailed 3-D wind field and the complex snow accumulation patterns.

Doppler wind lidar measurements resolve the complex flow interactions with the Sattelhorn ridge and show strong evidence for the formation of an eddy-like structure on the leeward side of Sattelhorn. ARPS simulations reproduce this structure, but with a smaller spatial extent and weaker mean flow. Based on these simulations the observed flow field is likely a combination of a cross-ridge and an along-slope flow, which may support the development of a leeside eddy due to topographic forcing. Similar flow patterns were found for simulations over an idealized pyramid [Voigt and Wirth, 2013], giving evidence for the flow pattern to probably occur more generally for steep alpine peaks and ridges.

Our ARPS simulations show that atmospheric stability, wind direction, and wind speed all have a strong influence on the development of the local near-surface flow field over the northern slope of Sattelhorn and thus the development of a leeside eddy. The fluctuation between southerly and southeasterly wind directions and the temporary occurrence of a strong near-surface stability may be crucial for the development of a strong eddy-like structure on the leeward side of the Sattelhorn. A dependence on local inflow direction was presented by Hug *et al.* [2005] for flow over less complex terrain such as the Gaudergrat. A strong dependence on stability was further presented by Wang and Huang [2017].

Due to the lack of detailed atmospheric stability measurements, the simulations are initialized based on stability estimates from weather stations at different altitudes. A more stable near-surface atmosphere may contribute to the large vertical extent of the eddy-like structure in the lidar retrievals as it may maintain along slope flows up to higher elevations. Additionally, stronger wind speeds and the higher complexity of the real terrain make it likely that the flow is in the regime of “leeside bluff body boundary-layer separation” as shown in our analysis of flow regimes, while the flow in the ARPS simulations is likely in the regime of “postwave separation.”

Snow accumulation patterns on the northern slope of Sattelhorn measured by terrestrial laser scanning (TLS) show some agreement to modeled snow accumulation, but they do not show a clear sign of preferential deposition. This might be because excess snow accumulation in very steep terrain is unloaded by avalanches [Sommer *et al.*, 2015] or by redistribution of snow due to stronger winds [Wirz *et al.*, 2011] over the northern slope of Sattelhorn compared to our ARPS simulations. Additionally, different flow regimes may play a crucial role on the exact location of snow deposition on the leeward side of the ridge. The stronger sign of preferential deposition on the northern compared to the southern slope of Sattelhorn in the Alpine3D simulations is likely due to stronger southerly winds and because of weaker solar irradiation, even though only 26% of the snowfall was during southerly winds.

Unfortunately, an initial problem with the lidar scanner and the subsequent lack of successful lidar retrievals due to a lack or overabundance of scattering particles limited the observations to before 23:00 UTC+1 and did not cover the entire snowfall event. A more complete picture of the influence of the small-scale wind field on snow accumulation would require a more continuous lidar data set and corresponding snow accumulation measurements. Actual measurements of the 3-D wind field would require the use of multiple lidars and would overcome the limitations and difficulties arising from the interpretation of a single lidar-based radial velocity field. However, such measurements are hard to achieve due to the complex logistics of placing, powering, and maintaining several lidars in subfreezing, avalanche-prone, and steep environments.

The unique combination of observations of snow distribution, meteorological variables, and of the fine-scale wind field, further augmented with high-resolution ARPS model simulations, shows the state-of-the-art process understanding in complex mountainous terrain. One of the objectives was to investigate preferential deposition in the vicinity of the steep and complex Sattelhorn ridge, but our results could only partially confirm the findings previously reported from less steep terrain such as the Gaudergrat ridge [Mott and Lehning, 2010] or the Wannengrat area [Mott *et al.*, 2010]. We note that TLS–Alpine3D agreement of snow distribution in these previous publications was remarkably good in some areas but already pointed out that quantitative agreement is poor in very steep parts most notably on top of steep ridges. Preferential deposition in less steep terrain was further found to be responsible for increased deposition at the lower half of windward slopes due to blocking or at leeward slopes due to flow separation [Lehning and Mott, 2016]. Snow transport was further identified to be the main driver for smaller-scale drifts such as cornices. Overall, we show that snow accumulation structures in very steep terrain are more complex than previously found for less steep and less complex terrain. This study as well as recent investigations by Wang and Huang [2017] suggest that over steep terrain the process of preferential deposition becomes more complex. The deposition patterns with reduced accumulation at the windward slope and enhanced accumulation at the upper leeward slope, may thus be less

important for steep slopes compared to less steep slopes. This could be either due to different processes dominating during snowfall or due to modulation of the accumulation patterns by snow drift and redistribution of snow by gravitational processes [Sommer *et al.*, 2015]. Future work will include measurements over a larger area and for several storm periods to address open questions on the relative importance of the different processes shaping the snow cover. It may also be required to use higher-resolution numerical simulations and/or work with a flow model that is better suited for very steep and rough terrain. A well-suited model for future studies may be the WRF model when it becomes available in an IBM (Immersed Boundary Method) version.

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