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[Voordendag, Annelies](#) ; [Goger, Brigitta](#) ; Prinz, Rainer; Sauter, Tobias; Mölg, Thomas; Saigger, Manuel; Kaser, Georg

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# Investigating wind-driven Snow Redistribution Processes over an Alpine Glacier with high-resolution Terrestrial Laser Scans and Large-eddy Simulations

Annelies Voordendag<sup>1,\*</sup>, Brigitta Goger<sup>1,2,\*</sup>, Rainer Prinz<sup>1</sup>, Tobias Sauter<sup>3</sup>, Thomas Mölg<sup>4</sup>, Manuel Saigler<sup>4</sup>, and Georg Kaser<sup>1</sup>

\*These authors contributed equally to this work.

<sup>1</sup>Department of Atmospheric and Cryospheric Sciences, Universität Innsbruck, Innsbruck, Austria

<sup>2</sup>Center for Climate Systems Modeling, ETH Zurich, Zurich, Switzerland

<sup>3</sup>Geographisches Institut, Humboldt-Universität zu Berlin, Berlin, Germany

<sup>4</sup>Climate System Research Group, Institute of Geography, Friedrich-Alexander-Universität Erlangen-Nürnberg, Erlangen, Germany

**Correspondence:** Annelies Voordendag (annelies.voordendag@uibk.ac.at), Brigitta Goger (brigitta.goger@c2sm.ethz.ch)

## Abstract.

Wind-driven snow redistribution affects the glacier mass balance by eroding or depositing mass from or to different parts of the glacier's surface. High-resolution observations are used to test the ability of large eddy simulations as a tool for distributed mass balance modeling. We present a case study of observed and simulated snow redistribution over Hintereisferner glacier (Ötztal Alps, Austria) between 6 and 9 February 2021. Observations consist of three high-resolution Digital Elevation Models ( $\Delta x=1$  m) derived from terrestrial laser scans taken shortly before, directly after, and 15 hours after snowfall. The scans are complemented by data sets from three onsite weather stations. After the snow fall event the snowpack decreased by 0.08 m on average over the glacier and typical snow redistribution patterns were observed. The decrease of the snow depth is to be attributed to both post-snowfall compaction and redistribution of snow. Simulations were performed with the WRF model at  $\Delta x=48$  m with a newly implemented snow drift module. The spatial patterns of the simulated snow redistribution agree well with the observed generalized patterns. Snow redistribution contributed  $-0.026$  m to the surface elevation decrease over the glacier surface on 8 Feb, resulting in a mass loss of  $-3.9$  kg m<sup>-2</sup>, which is in the same order of magnitude as the observations. With the single case study we cannot yet extrapolate to the impact of post-snowfall events on the seasonal glacier mass balance, but the study shows that the snow drift module in WRF is a powerful tool to improve knowledge on snow redistribution over glaciers and that the model setup can be applied to other mountain glaciers.

## 1 Introduction

The European mountain cryosphere is an important contributor to Alpine water availability and experiences - as the worldwide cryosphere - the effects of global climate warming (e.g. Fox-Kemper et al., 2021; Hock et al., 2022). The annual mass balances of the Alps' glaciers are increasingly more negative since the 1980's (Marzeion et al., 2012; Huss and Hock, 2018; Hugonnet



20 et al., 2021), and extreme glacier mass losses are observed in more recent years (Copernicus Climate Change Service (C3S),  
2023; Voordendag et al., 2023b; Cremona et al., 2023). However, the impact of small-scale processes such as cryosphere-  
atmosphere exchange or wind-driven snow transport on snow accumulation over mountain glaciers still represent a knowledge  
gap (e.g. Mott et al., 2018; Beniston et al., 2018). Spatial observations of snow cover changes on mountain glaciers are sparse  
and often only available on the point scale, and numerical weather prediction models on the kilometre range are not able to  
25 resolve the relevant small-scale boundary layer processes (e.g., surface fluxes) over highly mountainous terrain (Vionnet et al.,  
2016; Goger et al., 2018, 2019). On the other hand, distributed mass balance models (Machguth et al., 2006), e.g. COSIPY  
(Sauter et al., 2020), require high-resolution input fields to deliver respective information about a glacier's surface mass bal-  
ance. Among the usual meteorological variables (e.g., temperature, wind speed, relative humidity, total precipitation, etc), snow  
depth can also be used as an initial condition, improving the accuracy of distributed surface mass balance models.

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In general, snow depth distribution over complex terrain cannot be assumed to be homogeneous. The spatial precipitation  
pattern over mountains is heterogeneous due to multi-scale interactions of the atmospheric flow with topography (Frei and  
Schär, 1998; Isotta et al., 2013; Colle et al., 2013). Furthermore, during or after snowfall, the snowpack is affected by four  
processes: melt, compaction, sublimation, and wind-driven redistribution. Compaction of the snowpack can be driven by the  
35 overburden of its own weight, the pressure exerted by the wind and/or snow metamorphosis processes including melting and  
refreezing. Snow redistribution is the relocation of wind-borne snow, or also called snow drift, from one part of the snow-  
covered area to another (Cogley et al., 2011; Mott et al., 2018). Redistributed snow leads to snow depth decrease in areas  
where snow is eroded, and snow cover increases, where snow particles are deposited. The resulting snow patterns strongly  
depend on the local topography, and the wind speed and direction (Gerber et al., 2017; Vionnet et al., 2013, 2021; Sauter et al.,  
40 2013). The complex terrain makes mountain glaciers subject to heterogeneous snow cover distribution caused by both complex  
precipitation patterns and wind driven redistribution during and after snow fall (Dadic et al., 2010).

It is still a challenge to measure the spatial (re)distribution of the snow cover continuously in a complex alpine environment.  
One possible method to record glacier-wide snow distribution of precipitation and the post-snowfall surface elevation changes  
over a glacier is with repeated Digital Elevation Models (DEM) derived from terrestrial or airborne laser scanning (TLS/ALS).  
45 In recent times, surface elevation changes at mountain glaciers were measured with both TLS (Fischer et al., 2016; Prantl et al.,  
2017; Xu et al., 2019; Mendoza et al., 2020) and ALS (Grünwald et al., 2014, , Table 2). However, these DEMs are acquired  
irregularly and at low temporal resolution. Data series that capture snow fall and snow redistribution thereafter are not available  
so far. This gap was addressed with the installation of a permanent TLS station nearby Hintereisferner glacier (HEF, Ötztal  
Alps, Austria) (Voordendag et al., 2021b). This TLS station acquires a daily DEM automatically, but even hourly acquisitions  
50 are possible if manually initiated. A comprehensive uncertainty assessment shows that this TLS station is able to capture small  
changes at the glacier surface, such as snow (re)distribution (Voordendag et al., 2023a). The high temporal and spatial data  
resolution contribute to improving the process-understanding at HEF and can be used to validate surface elevation changes in  
atmospheric model simulations.



55 On the modelling side, high-resolution large-eddy simulations (LES,  $\Delta x < 100$  m) over complex terrain have become a new possibility for process studies in recent years, e.g. foehn-cold air pool interactions (Umek et al., 2021) or snow accumulation patterns over complex alpine terrain (Gerber et al., 2018). Sauter and Galos (2016) explored with semi-idealized simulations the sensible heat flux pattern over an alpine glacier and found that its heterogeneity depends on the wind speed and direction. These results were confirmed by real-case LES by Goger et al. (2022) over HEF, finding that the turbulent surface fluxes on the glacier were strongly governed by the mesoscale flow direction and upstream conditions. However, all of the aforementioned studies focused on processes in the atmospheric boundary layer, but not on wintertime atmosphere-cryosphere exchange like wind-driven snow redistribution.

While most atmospheric numerical weather prediction (NWP) models include a multi-layer snow scheme in their land-surface models, e.g. the three layer NOAA-MP scheme in the Weather Research and Forecasting (WRF) model (Niu et al., 2011) or the multi-layer snow model in the Integrated Forecasting System (IFS, Arduini et al., 2019), the explicit treatment of wind-driven snow redistribution is often not implemented. Some studies couple full stand-alone snowpack models with NWP models. For example, Vionnet et al. (2014) coupled the Crocus snow model with the Meso-NH LES model to explore snow accumulation patterns. They found that the wind-induced snow redistribution is responsible for an increase in spatial variability in snow depth. The most recent development in this direction is CRYOWRF (Sharma et al., 2023), where SNOWPACK (Lehning et al., 1999) was coupled to the WRF model, also including a snow drift module. First results suggest that CRYOWRF can successfully simulate snow accumulation and redistribution both over the Antarctic and the Swiss Alps, but the disadvantage of this and other approaches is the extra coupling of the models with a (previously) stand-alone snowpack model.

In this work, we show a case study with high-resolution TLS observations of wind-driven snow redistribution on HEF. The worldwide unique setup of a permanent TLS station overlooking the glacier provides us with sub-daily observations at a horizontal resolution of 1 m. However, since the mere detection of change in snow depth does not allow for resolving the different processes at work, we estimate the different contributions to the snow depth change from discussing the weather conditions during the studied event as mirrored by on-site weather station data. In turn, we can evaluate simulations with the WRF model at  $\Delta x = 48$  m with a newly implemented snow drift module by Schmid (2021) with this rich observational data set. The simulations have the potential to resolve the main driving processes to snow surface depth change such as compaction and wind-driven snow redistribution. Finally, we assess the potential impact of these post-depositional processes of the snow cover on the glacier's mass balance.

## 2 Methods

This paper presents a case study between 6 and 9 February 2021. First, this section introduces the study area with its available observations and the case study selection. Second, we elaborate on the available TLS data. Last, the model setup and the novel snow drift module for this particular case study are introduced.



## 2.1 Study area and available observations

The Hintereisferner (HEF) is part of the Rofental catchment (Strasser et al., 2018) and is a principal research site to study glaciological processes and glacier mass balances since the early days of glacier research (Blümcke and Hess, 1899). Annual and seasonal glaciological mass balance measurements are acquired since 1952/53, and HEF is classified as one of the ‘reference glaciers’ by the World Glacier Monitoring Service (WGMS, Zemp et al., 2009). HEF has presently (2018) a length of approximately 6.3 km and stretches from its highest point, the Weißkugel (3738 m a.s.l.), down to a terminus altitude of 2460 m a.s.l.

The glacier and its surroundings are well-equipped with several automated weather stations (AWS). The major station for this study is *Im Hinteren Eis* (IHE), which is located on the orographic right side of the glacier on the ridge at the Austrian-Italian border (Fig. 1). This measurement location is equipped with a permanently installed TLS, which is extensively described in Voordendag et al. (2021b, 2023a) and Section 2.3 of this study. Additionally, two webcams, which deliver images every thirty minutes<sup>1</sup>, are installed at the position of the TLS and overlook the glacier surface. About 50 m from the container with the TLS, an eddy-covariance flux tower is installed at the mountain ridge (Table 1) providing, among the usual meteorological variables, observations of turbulence kinetic energy and surface fluxes. After post-processing (Stiperski and Rotach, 2016; Rotach et al., 2017), the averaged variables are available in a 15-minute interval. The second location is *Station Hintereis* (StHE, Fig. 1a), located on the orographic left side of the glacier equipped with an AWS (Table 1) and a mountain hut used for logistical support. Last, a temporary AWS was installed at the glacier in the line of sight between IHE and StHE from 7 Dec, 2020 to 22 Feb, 2021. This station is called AWS28 hereafter, as it was installed at an altitude of approximately 2800 m a.s.l.. It provides common meteorological measurements (Table 1). The meteorological observations from HEF are used to explain the meteorological situation and to validate the simulations (see Sect. 2.4) on a point scale.

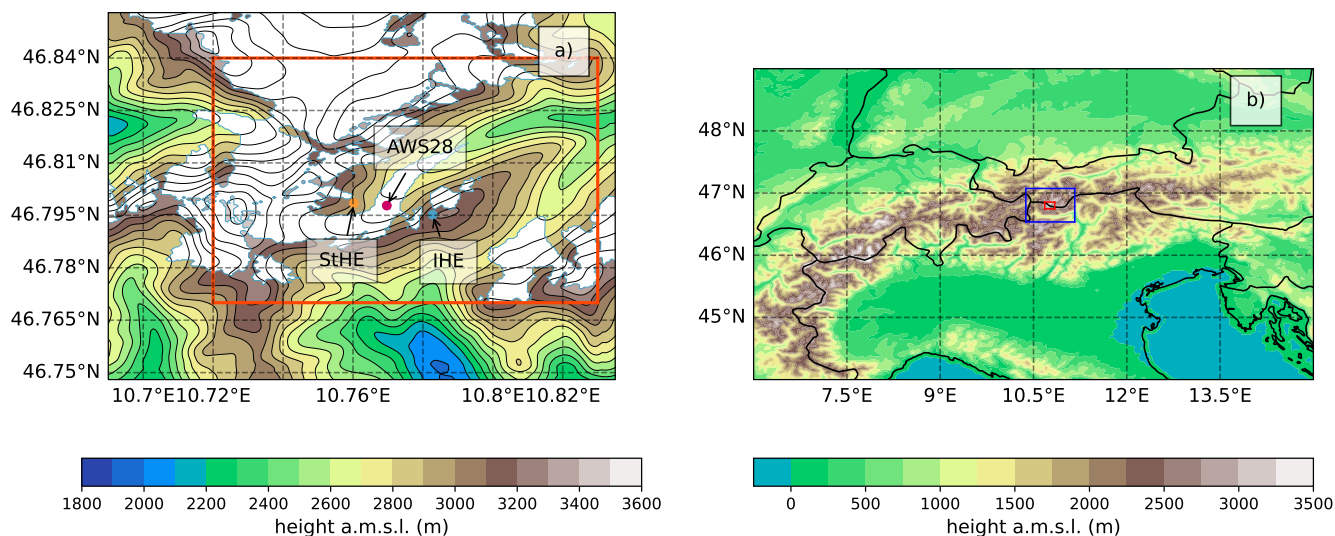
## 2.2 Case Study Selection

The case study has been selected based on the meteorological observations and the images recorded by the webcams by applying the following criteria:

- the period should be in winter and show a pronounced change of the synoptic weather situation with a subsequent significant accumulation of fresh snow;
- wind speeds above  $5 \text{ m s}^{-1}$  should be observed during or directly after the snowfall event to ensure snowdrift;
- no surface elevation change due to melt should be observed;
- frequent TLS scans must be available (Sect. 2.3).

The time window of 6 Feb-9 Feb, 2021 met these criteria.

<sup>1</sup>[www.foto-webcam.eu/webcam/hintereisferner1/](http://www.foto-webcam.eu/webcam/hintereisferner1/)



**Figure 1.** Panel a): Overview of the innermost model domain ( $\Delta x = 48$  m, red rectangle) with the model topography (contour lines) and the glacier areas as represented in the model (white area with light blue outlines). The locations of the three stations StHE, AWS28, and IHE are highlighted in colors. Panel b): The mesoscale domain ( $\Delta x = 1$  km) spanning the Alps with the two LES domains highlighted in blue ( $\Delta x = 240$  m) and red ( $\Delta x = 48$  m).

115 The large-scale synoptic situation from ERA5 reanalysis data revealed that on 6 Feb, a through was moving westward from France towards the Alps. The associated frontal system passed our area of interest in the early hours of 8 Feb. The frontal system was associated with pre-frontal snow fall, which ceased in the early morning hours of 8 Feb. After the through passage, winds at upper levels shift towards Westerlies, while at crest levels, winds shift to Westerlies to South-westerlies. At around 12 UTC, the wind speeds increased to over  $5 \text{ m s}^{-1}$ , providing excellent snow drift conditions. Webcam imagery (Fig. 2) of 120 8 Feb, 15:30 local time (UTC+1) shows blowing snow at the mountain ridges surrounding the glacier, indicating high wind speeds and snow drift.

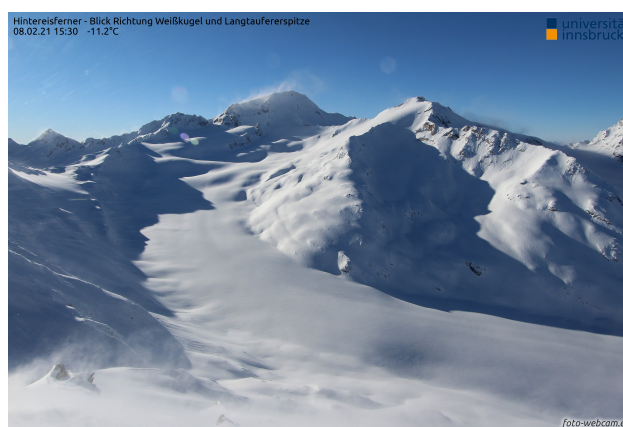
### 2.3 Terrestrial laser scanning acquisitions

The location IHE is equipped with a permanently installed and automated TLS station (Voordendag et al., 2021b). The TLS station is in operational daily use since 2020, and thus delivers a daily point cloud of HEF under clear weather conditions (e.g. 125 no clouds between TLS and target surface). The TLS station is normally set to a daily acquisition at 01:42 UTC, but as the end of a snowfall period was observed on the webcam images on 8 Feb, 10:22 UTC, an additional scan acquisition was initialized. This led to three usable scans for the case study: shortly before the snowfall event on 6 Feb (01:42 UTC), directly after the snowfall 8 Feb (10:22 UTC) and approximately 15 hours after snowfall ended on 9 Feb (01:42 UTC). In the following text, we refer to these scans as scan 1, 2, and 3, respectively. The acquired point clouds are registered to each other with the RiSCAN PRO software (RIEGL, 2019) and gridded to Digital Elevation Models (DEMs) with an one-meter horizontal resolution. Voordendag 130



	Im Hinteren Eis	Station Hintereis	AWS28
Latitude	46.795761	46.798896	46.79779
Longitude	10.783409	10.760373	10.76967
Altitude (m a.s.l.)	3264	3031	2782
Air pressure	Setra278 CS100	Setra278 CS100	Vaisala PTB 110
Air temperature	Rotronic HC2-S3 Ventilated; at 1.50 m and 5.50 m	Campbell Scientific EE181 Naturally ventilated; at 2.10 m	Campbell Scientific CS215
Precipitation	Geonor T200B	Ott Pluvio2L	
Radiation	Kipp&Zonen CNR4	Kipp&Zonen CNR4	Kipp&Zonen CNR4
Relative humidity	Rotronic HC2-S3 Ventilated; at 1.50 m and 5.50 m	Campbell Scientific EE181 Naturally ventilated; at 2.10 m	Campbell Scientific CS215
Snow depth	Campbell Scientific SR50ATH-L	Campbell Scientific SR50A	Campbell Scientific SR50A
Wind speed and direction	Lufft Ventus-UMB At 1.50 m, 3.00 m and 6.00 m	Young 05103-45 Alpine At 3.30 m	Young 05103
3D sonic anemometer	Metek uSonic-3: At 3.18 m		
TLS	RIEGL VZ-6000		
Webcams	Canon EOS1200D (2x)		

**Table 1.** Location, altitude and available observations at Im Hinteren Eis, Station Hintereis and AWS28 during the period of the case study.



**Figure 2.** Webcam image of 8 Feb, 2021, 15:30 (UTC+1) showing signs of snow drift at the mountain ridges. The image is retrieved from <https://www.foto-webcam.eu/webcam/hintereisferner1/2021/02/08/1530>.

et al. (2023a) investigated surface change processes that can be captured by the TLS station: They found that the scans have an uncertainty of  $\pm 0.10$  m in vertical direction after the registration in RiSCAN PRO. Additionally, the high-resolution point clouds are gridded to DEMs with a  $\Delta x = 48$  m allowing a direct evaluation of the numerical simulations.



## 2.4 Numerical model

135 We employ the Weather Research and Forecasting (WRF) model version 4.1 (Skamarock et al., 2019) in a nested set-up for the numerical simulations of the case study period. The numerical setup is the same as described and used in Goger et al. (2022), therefore we only mention the most relevant aspects. As model boundary conditions, we use ERA5 reanalysis data, feeding the outermost WRF domain ( $\Delta x = 6$  km) spanning Europe, and subsequently nesting down across  $\Delta x = 1$  km (mesoscale domain) to the two LES domains at  $\Delta x = 240$  m and  $\Delta x = 48$  m, respectively. We put a special focus on the correct representation of

140 land-use and glacier outlines in the LES domains, therefore we use the CORINE land-use data set (European Environmental Agency, 2017) with an additional correction of the ice surfaces of the glaciers as described in Goger et al. (2022). NOAH-MP (Niu et al., 2011) is used as a land-surface scheme, which includes a three-layer snow model. Furthermore, we implemented a novel snow drift module as described in Section 2.4.1. We use the Thompson microphysics (Thompson et al., 2008), the MM5 revised surface layer scheme (Jiménez et al., 2012), and the RRTMG two-stream radiation scheme (Iacono et al., 2008) with

145 topographic shading for all domains. For the boundary layer turbulence we employ the MYNN parameterization (Nakanishi and Niino, 2009) for the two outermost domains, while we switch it off in the LES domains and employ the turbulence closure after Deardorff (1980). Furthermore, we also use the online averaging module by Umek et al. (2021), therefore, all model output shown in the following are 15-minute averages. The model is initialized on 8 Feb, 00 UTC, and runs for 24 hours. We would like to stress that the snowpack in the model is initialized at the same time. This might introduce a slight bias in

150 snowpack density, since the model initializes the snowpack as "fresh snow", while in reality, an older snowpack is already present at the glacier and its surroundings. However, due to the expensiveness of the LES, a long spin-up period of e.g., weeks, is not feasible with our current set-up. We consider this possible shortcoming in our later analysis of the model data, and keep in mind that the modelled snowpack density profile likely differs from reality.

### 2.4.1 Snow drift module

155 A snow drift scheme based on the work of Sauter et al. (2013) was recently added to WRF by Schmid (2021). While the detailed description of the module and the implementation in WRF is subject to another paper (Sauter et al., 2023, in prep.), we outline the governing equations and most relevant features of the scheme in the following paragraphs.

The scheme builds on the widely-used approach of dividing the process of drifting snow into a saltation layer and snow particles in suspension. The saltation layer is fully parameterized and acts as a lower boundary condition for the flux of

160 snow into suspension. The drifting snow mainly interacts with the mean flow, while neglecting particle interactions. The mass conservation of snow particles is given by the continuity equation

$$\frac{\partial \phi_s}{\partial t} + \frac{\partial(\phi_s u_i)}{\partial x_i} = \frac{\partial}{\partial x_3} \left( \nu_t \frac{\partial \phi_s}{\partial x_3} - V \phi_s \right) + \left( \frac{\partial \phi_s}{\partial t} \right)_{sub}, \quad (1)$$

where  $\phi_s$  is the mass concentration of snow particles in the suspension layer,  $\partial \phi_s / \partial t$  is the local rate of snow concentration change,  $x_i$  are the Cartesian coordinates,  $u_i$  are the Cartesian components of the velocity vector,  $\nu_t$  is the turbulent viscosity,





165 and  $V$  is the terminal fallout velocity. The fallout velocity

$$V(z) = -\frac{A}{r(z)} \sqrt{\left(\frac{A}{r(z)}\right)^2 + B \cdot r(z)}, \quad (2)$$

depends on the snow particle radius at height  $z$

$$r(z) = r_0 \cdot z^{-0.258}, \quad (3)$$

$A$  and  $B$  are constants and are calculated with:

$$170 \quad A = \frac{6.203 \cdot \nu_{air}}{2} \quad (4)$$

and

$$B = \frac{5.516 \cdot \rho_{ice}}{4 \cdot \rho_{ice}} \cdot g. \quad (5)$$

Here  $\nu_{air}$  represents the viscosity of air,  $\rho_{ice}$  the ice density and  $g$  the acceleration due to gravity.

$r_0$  is the particle radius at ground level following (Gordon et al., 2010) with:

$$175 \quad r_0 = 0.5 \left( \frac{7.8 \cdot 10^{-6} u_*}{0.036} + 31 \cdot 10^{-6} \right), \quad (6)$$

and  $u_*$  the friction velocity. Optionally,  $V(z)$  and  $r_0$  can be set to constant values in the model settings.

The last term in Equation 1 accounts for the mass loss of suspended snow due to sublimation, where sublimation-loss rate of suspended snow is approximated by  $\psi_s \phi_s$ , with  $\psi_s$  as the sublimation-loss rate coefficient. This coefficient describes the change of snow particle mass due to heat exchange, and ventilation effects. The scheme considers the effect of sublimation  
180 on the vertical temperature and humidity profiles in the boundary layer. This feedback mechanism self-limits the sublimation process, because its intensity depends on the saturation deficit of the atmospheric environment (Sauter et al., 2013).

In the saltation layer, snow mass concentration is gained by aerodynamic entrainment from the snowpack below. Snow transport occurs when the surface shear stress exceeds the cohesive bond of the particles. The erosional mass flux is therefore proportional to the excess surface shear stress

$$185 \quad q_e = e_{salt} \rho_a \left( \left[ (u_{th} - u_*) \left( \frac{\phi_{salt}}{\phi_{max}} \right) + u_* \right]^2 - u_{th}^2 \right), \quad (7)$$

where  $\rho_a$  is the air density,  $u_*$  the surface shear stress,  $\phi_{salt}$  the concentration in the saltation layer,  $u_{th}$  the friction threshold velocity, and  $\phi_{max}$  the maximum particle concentration in the saltation layer. Since the particle erosion process depends on



the cohesive bonds of the snow particles, the surface shear stress threshold is a function of the snow density  $\rho_s$  (Walter et al., 2004):

$$190 \quad u_{th} = 0.0195 + (0.021\sqrt{\rho_s}). \quad (8)$$

The efficiency of the erosion process is governed by the heuristic parameter  $e_{salt}$  [-]. Particle drag reduces the momentum, which in turn limits the capacity to eject further particles. When  $\phi_{salt}$  reaches  $\phi_{max}$ , the friction velocity reduces to the friction threshold velocity and the release of snow particles is stopped. The upper limit of  $\phi_{max}$  is given by the semi-empirical relationship (Pomeroy and Male, 1992),

$$195 \quad \phi_{max} = \frac{\rho_a}{3.29 \cdot u_*} \left(1 - \frac{u_{th}^2}{u_*^2}\right). \quad (9)$$

When the friction velocity drops below the threshold, particle deposition takes place. The deposition flux  $q_d$  corresponds to the downward flux and the modified shear stress ratio

$$q_d = V\phi_s \cdot \max\left(\frac{u_{th}^2 - u_*^2}{u_{th}^2}, 0\right). \quad (10)$$

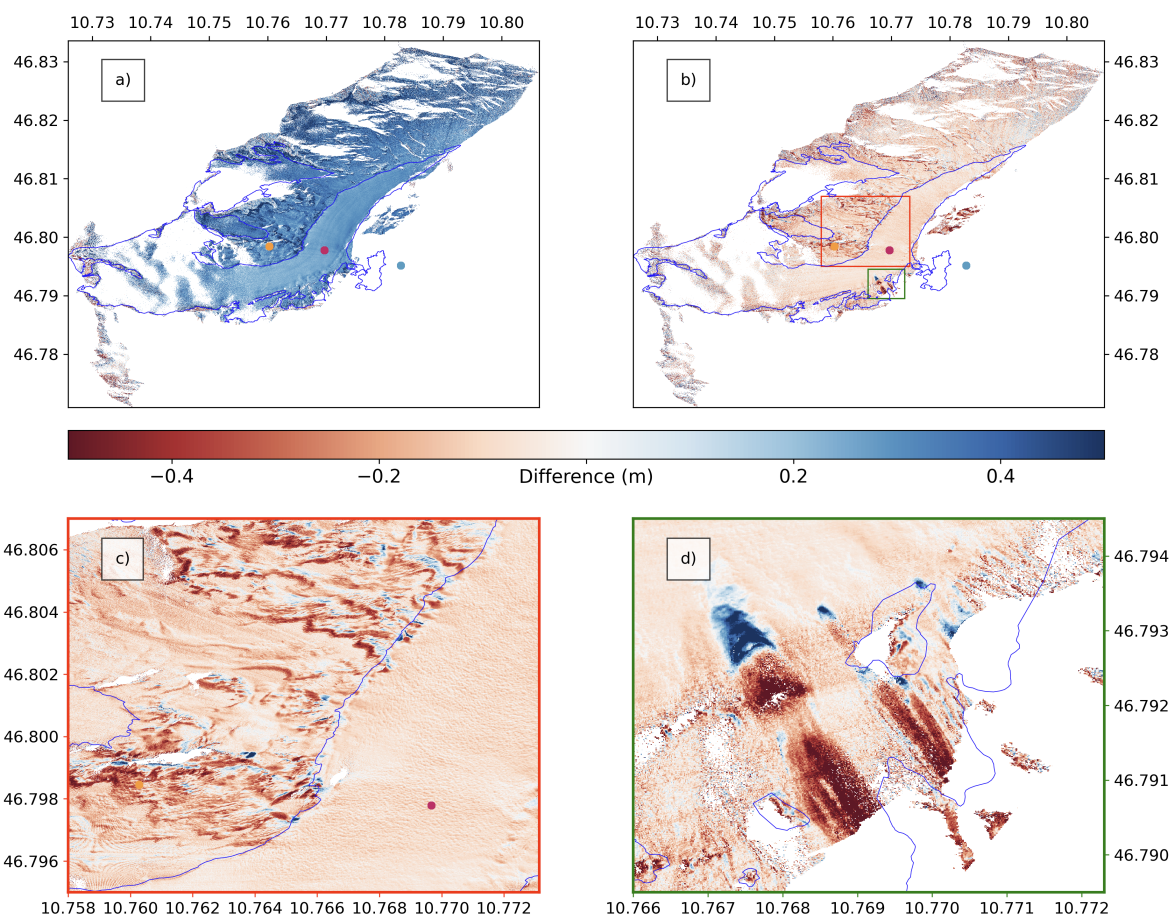
200 The first term on the right-hand side describes the vertical turbulent mixing of the snow and the terminal fall velocity  $V$ , while the second term shows the effect of sublimation in snow mass flux change.

### 3 Results

First, the observed snow depth changes from the TLS acquisitions are introduced and discussed in detail. In the next subsections, we explain the accompanying processes with the aid of further observations at point scale. We use the observations to evaluate the results from the LES, especially in terms of wind patterns and the resulting snow redistribution. Last, we compare 205 the observed and modelled spatial patterns.

#### 3.1 Observed snow depth changes with the TLS

The three TLS scans reveal the changes in snow depth over several days, and also show the heterogeneous snow distribution over the glacier and its surroundings. First, the DEM of Difference (DoD) between scan 1 and 2 shows an increase of surface elevation over almost the entire area of interest (Fig. 3a). The surface elevation increase is snowfall: the precipitation gauges at 210 IHE and StHE registered precipitation, and the snow depth sensor at AWS28 observed a snow depth increase as well (Fig. 4). The snow was evenly distributed over the glacier surface, but the slopes adjacent to HEF showed a more heterogeneous snow distribution between scan 1 and 2 (Fig. 3a), which might indicate snow redistribution during the snow fall event. From the TLS data, 0.28 m of snow were deposited on average over the glacier and the snow depth sensor observed an increase of 0.45 m between scan 1 and 2 at AWS28. In this study, we do not elaborate on the snow depths at IHE and StHE, as the snow depth



**Figure 3.** DEM of Difference of the TLS scans ( $\Delta x=1$  m) between a) scan 1 and scan 2, and b) scan 2 and scan 3. c) is a zoom of the red box in b) showing signs of snow redistribution, and d) is a zoom of the green box in b) showing avalanches. The glacier outlines (blue) are derived from the ALS data acquired by the Federal Government of Tyrol in 2018. IHE (blue), StHE (orange), and AWS28 (pink) are also plotted.

215 sensors are installed at unrepresentative locations of the surroundings, i.e. at wind-exposed ridges that are almost never covered with snow.



The DoD between scan 2 and 3 shows a general decrease of the snow depth over the glacier of 0.079 m on average (Fig. 3b). This is in agreement with the snow depth observations at AWS28, where a decrease of 0.08 m was observed by the snow depth sensor between scan 2 and 3 (Fig. 4). A zoom in on the glacier surface on the orographic left side of the glacier (Fig. 3c) and a  
220 look at the webcam images reveal patterns which are likely the results of snow redistribution, given their spatial structure. On the glacier surface around AWS28 (pink dot, Fig. 3c), a wavy pattern is evident with magnitudes between approximately -0.15 and -0.05 m. This is comparable to snow bedform observations over similar flat surfaces (Filhol and Sturm, 2015; Kochanski et al., 2018). With the resolution of the snow structure at  $\Delta x=1$  m and the webcam images, we cannot distinguish between the snow bedforms (i.e. waves, dunes, barchans or ripples), but the structure is wind-driven. At the slopes adjacent to the glacier  
225 surface, snow erosion is observed at the windward southwest slopes and this snow is deposited directly at the closest northeast leeward slopes. This is particularly evident around the location of StHE and the orographic left side of HEF (Fig. 3c). These structures are mainly induced by the rough surface caused by rocks at the slopes surrounding the glacier and again, indicate wind-driven snow.

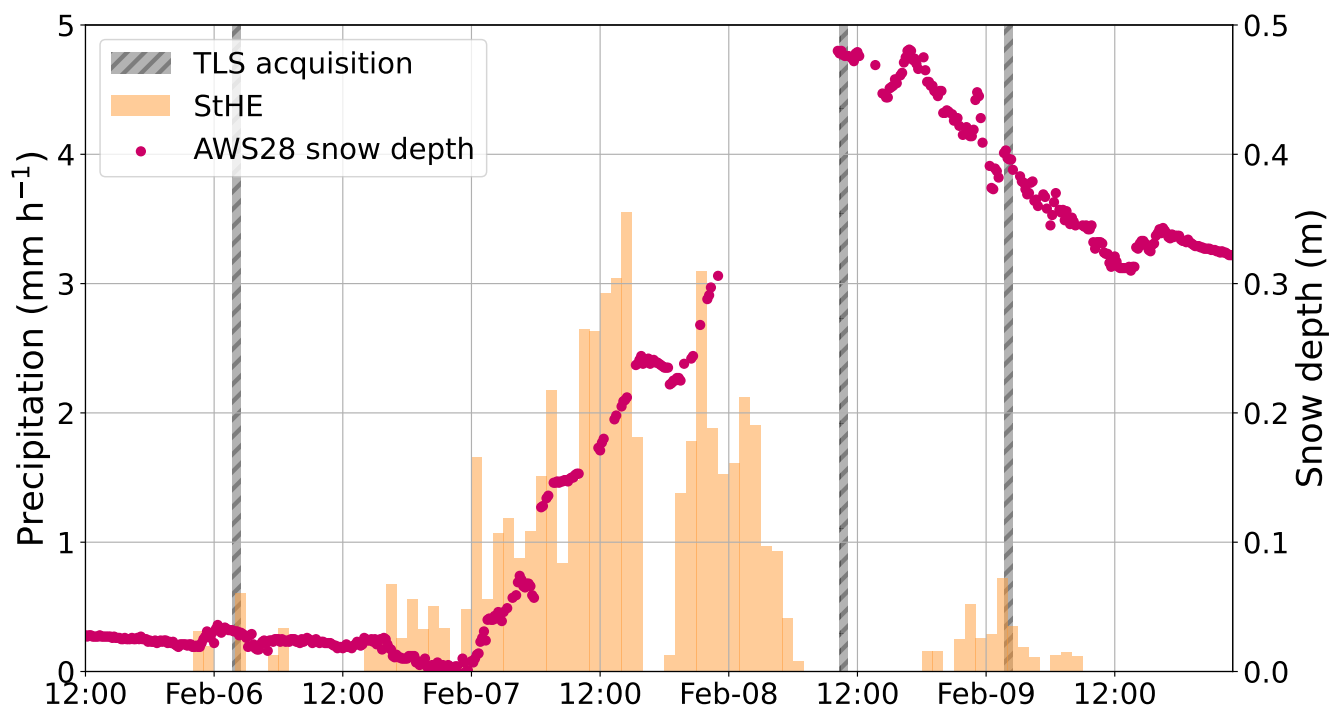
230 Avalanches, induced by fresh snow or wind slabs, are observed over the orographic right side of HEF at an altitude of 2910 m a.s.l. (Fig. 3d). The release zone of the avalanche with magnitudes between -0.25 and -0.68 m is indicated by the dark red color, whereas the dark blue zone shows the deposition area of the avalanche up to +1.26 m high.

We here elaborated on the elevation changes at the glacier and the surroundings from the TLS observations, but we aim  
235 to investigate the nature of these changes. The surface elevation changes are caused by redistributed snow, compaction and sublimation. Melt can be excluded during the cold case study period. To distinguish between these processes, we now analyse the additional meteorological observations and the numerical simulations with the WRF model.

## 3.2 Meteorological situation at the glacier

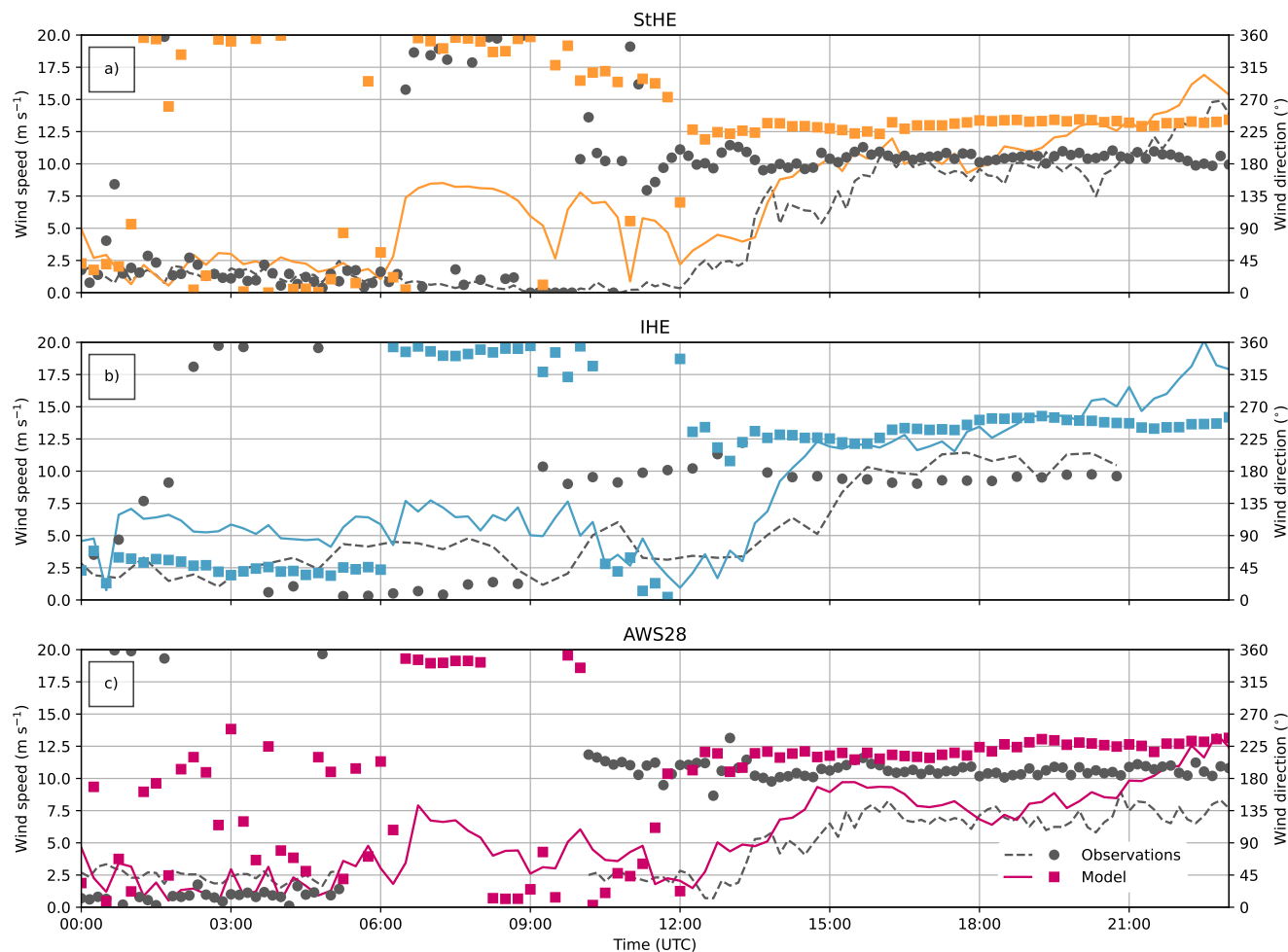
### 3.2.1 Precipitation

240 Mostly small precipitation amounts were registered on 5 and 6 Feb at StHE (Fig. 4). The situation changed when the front approached our area of interest. Webcam images, the precipitation gauge, and the increasing snow depth at AWS28 suggested that fresh snow was accumulated on 7 and 8 Feb. The last precipitation was registered at 06 UTC on 8 Feb. Precipitation was also registered during the acquisition of scan 1 and 3, but this precipitation was not evident in the TLS data and on the webcam images. For scan 1, the precipitation was actually registered after the TLS acquisition, as seen in the 10-min data. For scan 3, we  
245 speculate that this is snow drift that is captured by the precipitation gauge. Snow redistribution patterns can only be simulated correctly if the modelled precipitation and wind patterns agree with the observations. In agreement with the observations (Fig. 4), the model simulates precipitation until 06 UTC (Fig. 6). We omit a direct comparison of the precipitations amounts at the precipitation gauge and the model, as precipitation observations are subject to undercatch, which is a well-known problem for precipitation (Goodison et al., 1998; Rasmussen et al., 2012; Smith, 2007; Colli et al., 2014). The data of the precipitation



**Figure 4.** Precipitation (StHE: orange) and relative snow depth observations (AWS28: pink) during the case study period. Note that the precipitation observation is not corrected for undercatch and the snow depth is arbitrarily chosen to start at 0 m at the minimum observed during the case study period, despite the glacier was already covered in snow and thus, the actual snow depth was more than 0 m. The grey bars indicate the time of the TLS acquisitions.

250 gauge at IHE is not analysed here, as the gauge is placed at a wind-exposed ridge that is hardly ever covered with snow and is thus prone to large amounts of undercatch. The timing and registration of snowfall support the assumption that the snow depth increase between scan 1 and scan 2 (Fig. 3a) was due to solid precipitation.



**Figure 5.** Time series of observations (grey) from the three weather stations and corresponding model output from the closest grid point in the model (IHE: blue; StHE: orange; AWS28: pink) of wind speed (lines) and wind direction (dots/squares) at 8 Feb 2021. Missing values in the observations are at IHE from 21:00 UTC onward, and at AWS28 between 05:45 UTC and 11:00 UTC.

### 3.2.2 Wind speed and direction

Observed time series of the wind speed and wind direction reveal the meteorological situation at the glacier and its surroundings on 8 Feb (Fig. 5). All stations observed low wind speeds (less than  $5 \text{ m s}^{-1}$ ) with mainly northerly flow during the night and the morning hours (8 Feb, 00 UTC–09 UTC). The wind direction changed towards southwesterly at around 09 UTC, while wind speed increased to more than  $5 \text{ m s}^{-1}$ . The wind speed increased even to more than  $10 \text{ m s}^{-1}$  at the south-facing slope (StHE) and the crest (IHE), while the wind speeds on the glacier remained below  $10 \text{ m s}^{-1}$ . After 15 UTC, higher wind speeds (more than  $10 \text{ m s}^{-1}$ ) allowed for wind-driven snow distribution. Similar to the weather station observations, the model simulates low



260 wind speeds during the nighttime. However, the sudden increase in wind speed sets in an hour earlier than in the observations  
(Fig. 5). Furthermore, the wind direction deviated slightly (below  $30^\circ$ ) from the observations throughout the simulations. Both  
the observations and the model suggest dominating southwesterly wind directions with wind speeds over  $5 \text{ m s}^{-1}$  at all three  
stations after 8 Feb, 12 UTC. This agrees with the TLS observations (Fig. 3c), which also indicate snow redistribution due  
to these strong, southwesterly winds. Overall, the model simulates the wind field on the glacier and the surrounding with  
265 confidence and provides good input conditions for the snow redistribution and snow water equivalent (SWE) estimates.

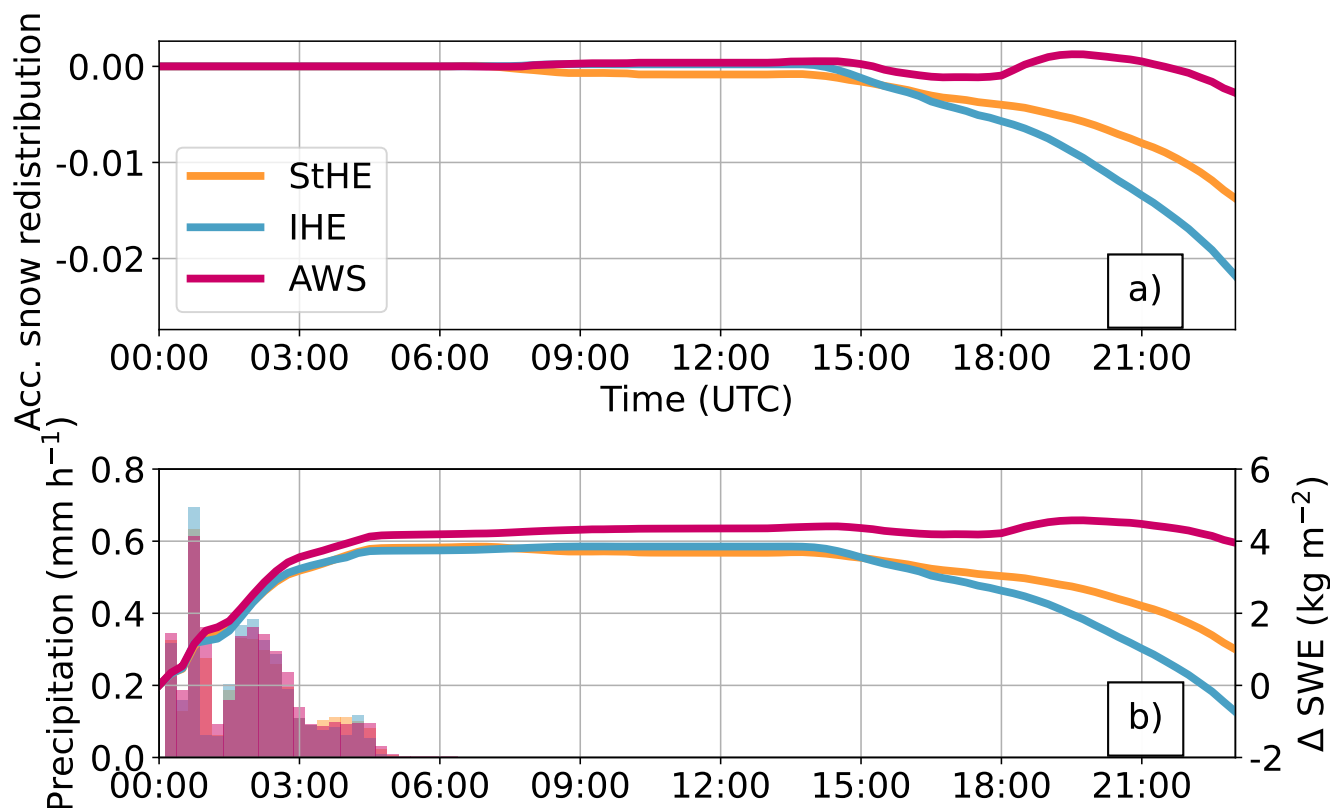
### 3.2.3 Snow Water Equivalent, compaction and snow redistribution at point scale

The TLS observations do not give information on the individual contributions of redistributed snow, sublimation, and com-  
paction to surface elevation changes. Compaction of snow can be detected, if the SWE remains constant after a snow fall event  
together with a simultaneous, continuous decrease in snow depth. SWE observations are not available during the case study  
270 period, but to investigate the possible amounts of compaction at HEF, we had a look at data from an AWS that was installed  
at HEF in the winters of 2021/22 and 2022/23 at an altitude 3030 m a.s.l and provides SWE and snow depth data (Schröder,  
2023). In these two winters, we examined snow depth and SWE data of nine snow fall events with amounts between 0.14 m  
and 0.38 m of fresh snow at the AWS. 16 hours after the snow fall, the snowpack decreased between -6.5% and -25% of the  
respective fresh snow amounts. In the mean time, no significant changes in the SWE were observed. Even though the winters  
275 are not directly comparable (i.e., the winter of 2021/22 had extremely low precipitation amounts (Voordendag et al., 2023b)),  
the order of magnitude of compaction indicated that this process also likely occurred between scan 2 and 3 in February 2021.  
Furthermore, similar amounts of compaction are observed on other glaciers and snowpacks (Gugerli et al., 2019; Koch et al.,  
2019; Voordendag et al., 2021a). When we apply the compaction rates to the 0.28 m of fresh snow in the case study period, we  
find that between 0.018 m and 0.071 m of the surface elevation decrease can likely be attributed to compaction.

280 We now utilize the model output for further process understanding with a *qualitative* analysis of the modelled snowpack.  
This allows us to understand possible processes governing snowpack formation; therefore we start with snow redistribution at  
point scale. In the model, snow redistribution only occurs when the parameterised friction threshold velocity is exceeded by the  
current friction velocity. This value depends on the snow density (Eq. 8). Snow drift and subsequent redistribution is therefore  
only simulated after the increase of the wind speed to more than  $10 \text{ m s}^{-2}$  after 14 UTC. The simulated snow redistribution is  
285 found to be 0.022 m at IHE, 0.014 at StHE, and 0.003 m at AWS28 at the end of the simulation period (Fig. 6a). The differ-  
ences between the weather stations is directly related to the higher wind speeds at IHE and StHE than on the glacier at AWS28  
(Fig. 5). Furthermore, it is interesting to note that the onset of snow drift initially leads to mass loss at all stations, at AWS28,  
however, snow drift briefly accumulates snow again after 18 UTC, while at the end of the simulation, the overall effects of  
snow drift is mass loss. At the other two stations (StHE and IHE), snow drift continuously contributes to snow mass loss.

290

Simulated changes in SWE give more insights on mass changes in the snowpack. During the precipitation event at the  
beginning of the simulation, the simulated SWE also increases, with similar values for IHE and StHE, but with about  $0.4 \text{ kg m}^{-2}$   
higher values of SWE at AWS28 (Fig 6b). After snowfall ended and before snow drift started, the SWE values remain constant



**Figure 6.** Time series at 8 Feb 2021 from the closest grid point in the model to StHE (orange), IHE (blue), and AWS28 (pink) in panel a) of accumulated snow redistribution (relative to the start of the simulation at 00 UTC), and panel b) solid precipitation (bars), and the relative change of snow water equivalent ( $\Delta$  SWE, lines).

for the three stations, indicating that the surface elevation change during this period (Fig. 4) is snow compaction. As soon as snow drift started, SWE reduces as snow gets eroded at the locations of the stations, but with spatial differences between IHE, StHE and AWS28 (Fig 6). The ridge location IHE exhibits the biggest reduction in SWE due to its exposed location over the entire simulation time. AWS28, however, shows a smaller total loss of SWE, mainly because of the sheltered location of the glacier. To summarize, SWE increased until 06 UTC due to solid precipitation, while SWE remained constant until 15 UTC due to compaction, until snow redistribution led to a continuous decrease in SWE after 15 UTC at all station.

Higher wind speeds result in more snow particles in the air and a higher likeliness of sublimation. Along with that, sublimation also depends on the vapor pressure in the ambient air as well as on the snow particle size. However, the values of simulated sublimation remain very small (less than 1 kg at the end of simulation for the entire air column over all glaciated areas in Fig. 1) throughout the rest of the simulation (not shown). Therefore, we will not discuss this process in more detail, also because the sublimation contribution is much smaller than the uncertainty of the TLS.

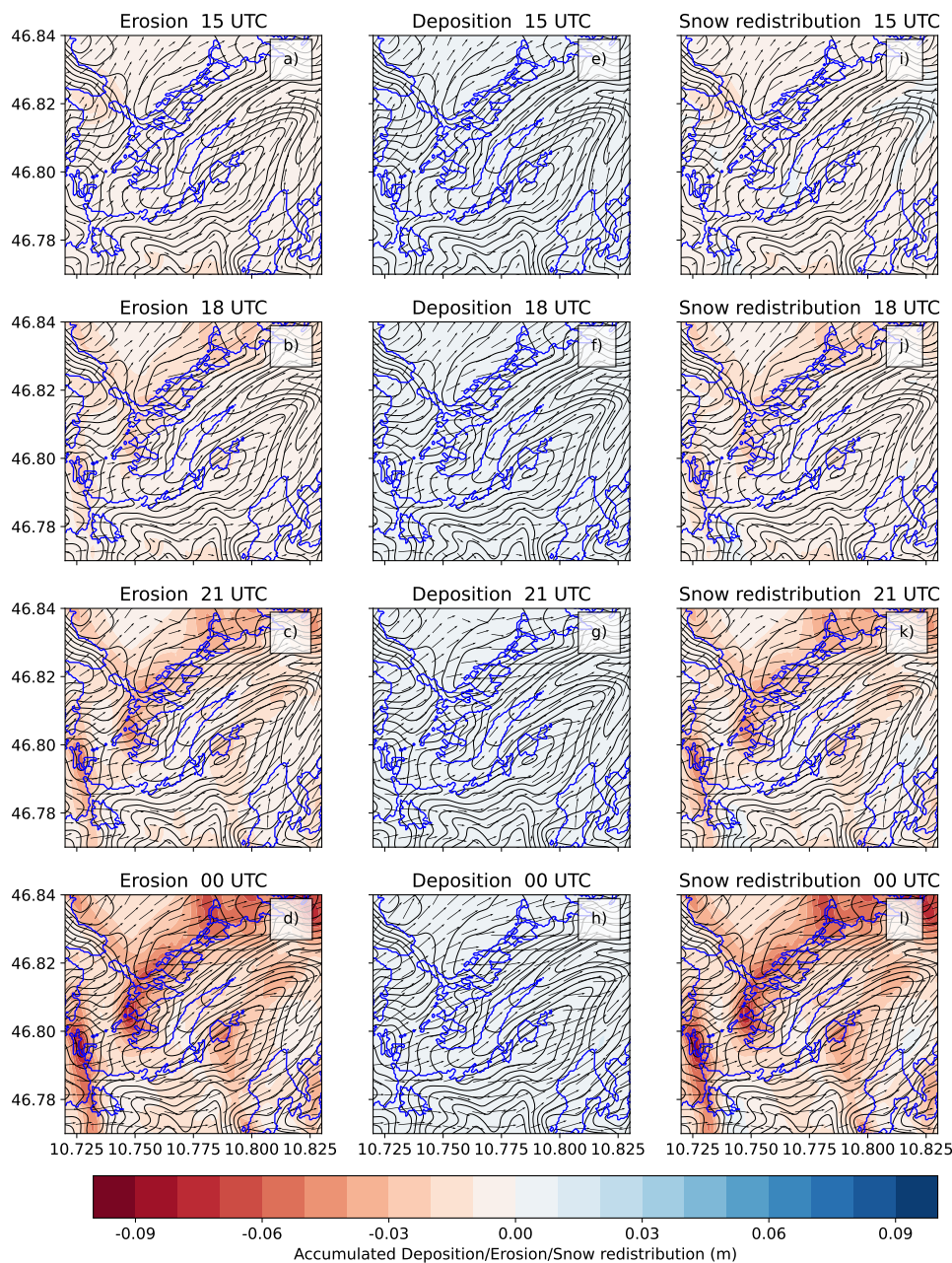




305 To summarize, during the precipitation phase, the SWE increases, while SWE mostly remains constant between 05 UTC and  
14 UTC, and thereafter snow drift starts to erode mass.

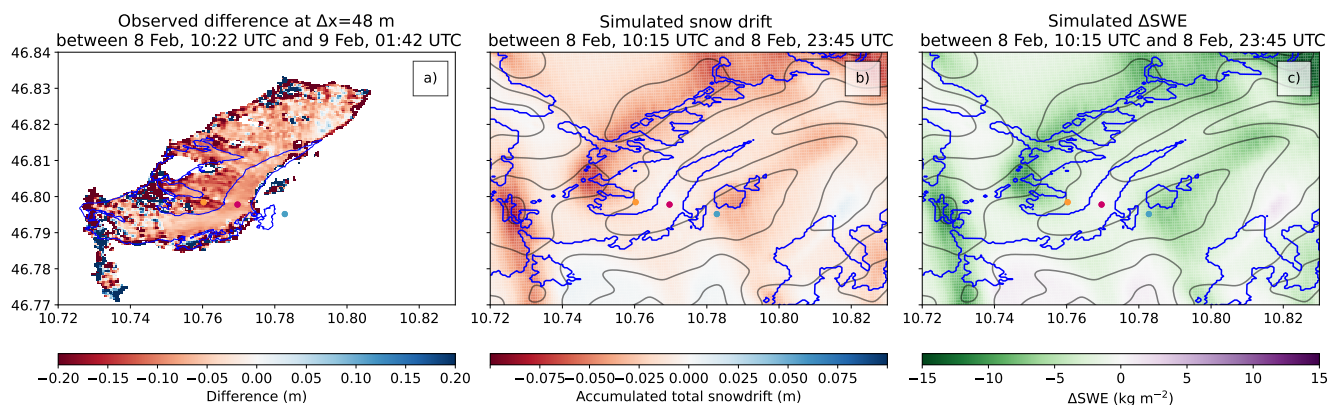
### 3.3 Spatial patterns of simulated snow redistribution processes

To explore the spatial patterns beyond the point scale, we analyse the modeled snow redistribution relative to the start of the  
simulation at 8 Feb, 00 UTC on the glacier and its surroundings (Fig. 7). The wind arrows reveal southwesterly flow at HEF  
310 and its surroundings with wind speeds over  $5 \text{ m s}^{-1}$  when snow redistribution occurred. The simulated flow patterns were,  
in agreement with the observations, almost perfectly down-glacier (Fig. 5). In the model, the governing process for snow  
redistribution is erosion (Fig. 7a-d), which was especially strong at the mountain ridge northwest of HEF. This is in accordance  
with the webcam images, which suggested that snow erosion mainly occurred at the surrounding mountain ridges. Snow  
deposition (Fig. 7e-h), however, was very small compared to erosion and does not exceed 0.01 m after 24 hours of simulation  
315 time. Therefore, the final snow redistribution (Fig. 7i-l) in the model is mainly governed by erosion, but some areas on leeward  
slopes get more deposition than erosion (Fig. 7l, i.e. around coordinates  $46.795^\circ\text{N}$ ,  $10.82^\circ\text{W}$ ). At the end of the simulation,  
the model suggests that around 0.09 m of snow were eroded at the mountain ridges, while at the glacier 0.03 m of snow were  
eroded. Although there was a weak positive signal in snow redistribution in the vicinity of AWS28 at 15 UTC (Fig. 6 and 7i),  
the sum of erosion and deposition resulted in an overall decrease in snow cover on the glacier. A main reason for this were  
320 the rather high wind speeds throughout the domain, as in the observations, wind speeds reach up to  $10 \text{ m s}^{-1}$  after 15 UTC,  
therefore snow can be easily eroded and transported towards the North-East and out of the domain.



– 10 m s<sup>-1</sup>

**Figure 7.** Simulated snow erosion (panels a-d), snow deposition (panels e-h) and the resulting net snow redistribution (panels i-l) in colors from 15 UTC, 18 UTC, 21 UTC and 00 UTC. Note that all values related to snow redistribution are summed up from simulation start (8 Feb, 00 UTC). Horizontal near surface wind speed and direction are indicated by black arrows. The black contours of 100 m equidistance show the model topography, blue contours indicate the glacier outlines.



**Figure 8.** a) Observed snow depth changes over HEF at  $\Delta x=48$  m between 8 Feb, 10:22 UTC and 9 Feb, 01:42 UTC, and b) the simulated snow redistribution and c) the simulated change in SWE between 8 Feb, 10:15 UTC and 9 Feb, 00:00 UTC. Please note the different orders of magnitude in a) and b).

### 3.4 Comparison of Observations with Simulations

A similar snow redistribution pattern as in the model also appears in the snow depth change observations by the TLS calculated to the model grid size of  $\Delta x=48$  m between scan 2 and 3 (Fig. 8a). However, we compare the snow depth changes from the observations (between 8 Feb, 10:22 UTC and 9 Feb, 01:42 UTC) to the simulated snow redistribution (between 8 Feb, 10:15 UTC and 9 Feb, 00:00 UTC, Fig. 8b). We compare these different variables and see that the order of magnitude of the snow depth changes is twice as large as the simulated snow redistribution due to snow drift, because the observed snow depth changes also include compaction. The average simulated decrease caused by snow redistribution is 0.026 m over the glacier (Fig. 8b), which equals a decrease in SWE of  $-3.9 \text{ kg m}^{-2}$  (Fig. 8c) in the simulated period over HEF. Considering that the observed compaction is estimated to be between 0.018 and 0.071 m of the fresh snow (Sec. 3.2.3) and the total decrease in observed snow depth is 0.079 m, we conclude that this mass and the magnitude of the simulated snow redistribution of 0.026 m is realistic and the observed snow redistribution patterns are captured accurately in the simulations.

## 4 Discussion

The present study combines operational TLS observations for a case study to evaluate LES to detect snow redistribution on an Alpine glacier. Since this is a small-scale phenomenon, it pushes both observations and modelling towards their boundaries.

The observations with the permanent TLS station are worldwide unique. Other studies also investigated snow depth changes with TLS (e.g. Mendoza et al., 2020; Gabbud et al., 2015; Fey et al., 2019) or ALS (Helfricht et al., 2014; Grünewald and Lehning, 2011), but these studies mainly covered coarser temporal resolutions or only covered small parts of a glacier. We were able to capture snow fall and redistribution directly thereafter, but we also note that the TLS observations are at the limits



340 of the capabilities of the system, as the uncertainty of the TLS observations was estimated at  $\pm 0.10$  m in vertical direction with manual post-processing (Voordendag et al., 2023a). For this study, the scans were registered with manually selected tie objects, such as snow-free rocks and the walls of StHE. We assume, looking at the registration of the scans at these tie objects, that the registration is even better than the  $\pm 0.10$  m in vertical direction, and thus the snow depth change between scan 2 and 3 was measured reliably, in a agreements with observations from the snow depth sensor at the glacier.

345 Modelling small-scale boundary-layer processes over mountainous topography is still a challenge for a NWP model like WRF, as discussed in the previous "summer study" by Goger et al. (2022). However, compared to the summer study, the model simulated even more realistic wind patterns over the glacier and its surroundings. Therefore, we assume that no model bias emerges due to erratic wind patterns. The simulated snow redistribution is realistic in terms of spatial structure and magnitude. Also, the simulated spatial patterns are very "smooth" compared to the TLS acquisitions. This is due to the smoothed model topography  
350 restricted by numerical stability and therefore no exceeding slope angles over  $35^\circ$  – the model topography clearly deviates from real topography. The simulations show that snow is eroded mostly at the ridges and the glacier is sheltered as observed in nature. We note that simulated deposition is very small compared to the simulated erosion. High wind speeds immediately redistribute freshly deposited snow again, until it is transported out of the domain, therefore, erosion strongly dominates. Also, the very small-scale snow redistribution areas (Fig. 3c) cannot be captured at a  $\Delta x = 48$  m, since Mott and Lehning (2010)  
355 noted that  $\Delta x = 10$  m or less would necessary to calculate the small-scale deposition patterns we observed with the TLS on the glacier. Still, we assume that the general snow redistribution patterns are well-simulated, as the model captures the larger snow redistribution at the mountain ridges and smaller snow redistribution and lower wind speeds at the less exposed parts of the glacier in agreements with weather station and TLS observations.

One of the advantages of the presented snow drift module in WRF is its sole simplicity compared to fully coupled atmospheric  
360 and snow models (Vionnet et al., 2013; Sharma et al., 2023). The snow drift module is embedded within the WRF code and the land-surface scheme NOAH-MP, respectively. Nevertheless, the NOAH-MP provides only three layers in the snowpack, whereas physical multi-layer snow models, such as SNOWPACK (Lehning et al., 1999) or COSIPY (Sauter et al., 2020), are able to simulate more layers and thus have a better representation of the processes in the snowpack. However, with the aim to investigate the contribution of snow redistribution, it is only necessary to calculate the surface shear stress  $u_{th}$  (Eq. 8) depending  
365 on the snow density of the upper layer in our snow drift module. However, the initialisation of the snowpack in our simulation is not comparable to nature, as the inner domain of the model is initialised with fresh snow only. The computationally expensive LES cannot be run with a long spin-up time to initialise the snowpack correctly. Thus, the model lacks of accurate information on the long-term snowpack evolution: In nature, the lower layers of the snow are compressed, but the upper layer with fresh snow is still uncompressed. It is more likely that snow drift takes place on an uncompressed, fresh snowpack rather than on a  
370 dense snowpack. However, we consider this unproblematic for the simulation, as in both nature and simulation only the fresh snow is eroded. In the model, snow compaction is calculated following Anderson (1976). The results of this snow compaction (not shown) are overestimated, because the model assumes the entire snowpack ( $> 2$  m) to compact and not only the 0.48 m of fresh snow. To conclude, we found realistic amounts of snow redistribution with our simulated results, we therefore conclude



that a three-layer model for the snowpack is sufficient to qualitatively assess wind-driven snow redistribution.

375

In general, snow redistribution contributes to the glacier mass balance (Dadic et al., 2010) and for this specific case study, snow redistribution has a negative effect on the glacier mass balance. In the simulation  $-3.9 \text{ kg m}^{-2}$  of snow is blown away from the glacier and out of the domain in the simulated period. We only focused on one case study, as the period was characterized by low wind speeds during snow fall and increased in the period thereafter. Furthermore, during this period, AWS28 was installed at the glacier and the second scan was taken directly after snow fall. It is clear that with this one "golden day" of snow redistribution on the glacier, we cannot attribute for the seasonal contribution of snow redistribution for the glacier mass balance. Further research is needed to investigate this seasonal contribution using our extensive TLS data set, preferably also to investigate snow redistribution patterns under different prevalent wind directions (e.g., Southerly or North-Westerly). Our study shows that a fresh snow fall event and a rapid increase in wind speeds directly thereafter are favorable conditions for snow drift to occur; therefore, snow drift is likely to occur mostly in connection with frontal passages or downslope windstorms. Obleitner (1994) studied wind regimes at HEF and found a dominance of mostly thermally-induced circulations; however, an updated version of the wind climatology at the glacier would allow new insights on under which wind regimes snow redistribution could occur.

Finally, although the installation of a permanent TLS station in remote mountainous terrain is a logistical challenge, the WRF model setup could be applied to any location worldwide. Therefore, our model setup can also be utilized for snow redistribution studies at other glaciated areas.

## 5 Conclusions

In this study, we introduced unique TLS scans to validate LES with the WRF model for quantifying the effect of snow redistribution over Hintereisferner, a major Alpine glacier in the Austrian Alps. For this purpose, we present a case study between 6 and 9 Feb, 2021, where multiple TLS scans and additional observations of wind speeds and snow depth on the glacier are available. Webcam imagery revealed snow drift in the area. With this rich observational data set, we evaluated large-eddy simulations at  $\Delta x=48 \text{ m}$  with the WRF model including a newly implemented snowdrift module. Our major findings are summarized as follows:

- Surface elevation changes due to snow fall and snow redistribution are observed with the TLS between 6 Feb, 01:42 UTC and 9 Feb, 01:42 UTC, 2021. Simulations were performed for 8 Feb, and run for 24 hours. The high-resolution setup in both observations and simulations at HEF is able to capture the glacier-wide snow redistribution patterns.
- Typical snow redistribution patterns are observed in the TLS scans. They show spatial heterogeneity, while on the glacier the patterns are less prominent than on the orographic left slope.
- Observations with the TLS show a glacier-wide spatially averaged decrease of 0.079 m of the snowpack the 15 hours directly after the snow fall. This reduction of the snow depth is a combination between snow compaction and snow



redistribution. With the observations at hand, compaction and drift cannot be separated, but the estimated compaction of the snowpack decrease is between 6.5% and 25% of the fresh snow amount.

- 410 – The WRF model at  $\Delta x=48$  m simulated the wind patterns at the glacier exceptionally well, and a newly implemented snow drift module allows a detailed comparison with the TLS acquisitions. The simulated integrated glacier-wide snow redistribution is on spatial average 0.026 m. The snow redistribution patterns are captured in a realistic manner compared to the observations.
- A qualitative inspection of the modelling results reveals that snow is mostly eroded on the surrounding mountain ridges, while the glacier itself is in a sheltered location and experiences less snow redistribution.
- 415 – Snow redistribution has a negative effect on the glacier mass balance in this case study with a simulated mass decrease of  $-3.9 \text{ kg m}^{-2}$  in 24 h. However, the contribution of these snow amounts to the seasonal glacier mass balance remains illusive as this study only covers one case study with a specific wind pattern, but this is an opportunity for further research.
- The high-resolution observations with the permanent TLS are worldwide unique and therefore not directly transferable to other glaciers.
- 420 – The WRF model setup with the snow drift module can be applied to any other location in the world, when high-resolution static and meteorological input data are available for the location of interest.

This study highlights the impact of snow distribution over a major Alpine glacier. The set-up, at least on the modelling side, is "portable" to any other location in the world. Snow redistribution patterns depend on the wind field and the local topography; therefore, our work shows the potential impact of small-scale boundary layer processes on glaciers' mass balance.

425 *Code availability.* The snow drift module for WRF can be found in the following Github repository of Christina Schmid: <https://github.com/ChristinaSchmid/snowdrift>. The namelists for the LES simulations are available at [https://github.com/gogerbr/wrf\\_les\\_hintereisferner](https://github.com/gogerbr/wrf_les_hintereisferner)

430 *Author contributions.* AV selected the case study period, and conducted and post-processed the TLS observations. BG conducted the WRF simulations and analyzed the model output. AV and BG wrote the original manuscript. RP oversaw the meteorological and mass balance observations at HEF. TS developed the snow drift module and implemented it with CS into the model code, while the code is currently maintained by MS. GK, TM, and TS conceived the project idea and oversaw the entire progress of the project. All authors contributed to the manuscript and improved it where necessary.

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