Large-eddy simulations of the atmospheric boundary layer over an Alpine glacier: Impact of synoptic flow direction and governing processes

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Abstract

The mass balance of mountain glaciers is of interest for several applications (e.g., local hydrology or climate projections), and turbulent fluxes can be an important contributor to glacier surface mass balance during strong melting events. The underlying complex terrain leads to spatial heterogeneity and non-stationarity of turbulent fluxes. Owing to the contribution of thermally induced flows and gravity waves, exchange mechanisms are fully three-dimensional, instead of only vertical. Additionally, glaciers have their own distinct microclimate, governed by a down-glacier katabatic wind, which protects the glacier ice and interacts with the surrounding flows on multiple scales. In this study, we perform large-eddy simulations with the Weather Research and Forecasting model at a horizontal grid spacing of 48 m to gain insight into the boundary-layer processes over an Alpine valley glacier, the Hintereisferner. We choose two case studies from the Hintereisferner experiment measurement campaign with different synoptic wind directions (southwest and northwest). Model evaluation with an array of eddy-covariance stations on the glacier tongue and surroundings reveals that the Weather Research and Forecasting model is able to simulate the general glacier boundary-layer structure. Under a southwesterly airflow, the down-glacier wind is supported by the synoptic wind parallel to the glacier axis, a stable boundary layer is present over the ice surface, and local processes govern the turbulence kinetic energy production. Under northwesterly airflow, a cross-glacier valley flow and a breaking gravity wave lead to strong turbulent mixing and to the subsequent erosion of the glacier boundary layer. Stationarity analysis of the sensible heat flux suggests non-stationary behaviour for both case study days, whereas non-stationarity is highest on the northwesterly day during the gravity-wave event. These results suggest that the synoptic wind direction...
has, in addition to upstream topography and the atmospheric stability, a strong impact on whether a local glacier boundary layer can form or not, influencing whether a glacier is able to maintain its own microclimate.

KEYWORDS
boundary layer, complex terrain, glacier, gravity waves, land–atmosphere exchange, large-eddy simulation, WRF

1 | INTRODUCTION

Mountain glaciers are an essential part of the global climate system, and monitoring their mass balance is required for a large number of applications, such as sea-level rise estimates (Marzeion et al., 2012), freshwater availability (Kaser et al., 2010), or natural hazard warnings (Kääb, 2011). Glacier mass balance, at any point on a glacier, is a result of energy, mass, and momentum fluxes at the glacier–atmosphere interface (Hock, 2005). Whereas the largest part of melt energy stems from the radiative flux, the turbulent fluxes of sensible and latent heat can be equally large over shorter time intervals (e.g., hours), especially during strong ice melting events. Driven by the wind shear, as well as by thermal stratification and moisture gradients, these turbulent fluxes are important vertical exchange mechanisms coupling the glacier surface and the atmosphere.

In general, the strength of surface fluxes is connected to surface characteristics and the atmospheric boundary layer (ABL) structure, which is spatially heterogeneous in mountainous environments (Rotach and Zardi, 2007; Lehner and Rotach, 2018; Serafin et al., 2018). Mountain boundary layers are influenced by dynamically driven topographic flows, such as gravity waves (e.g., Vosper et al., 2018) and foehn winds (Haid et al., 2020), but also by thermally induced circulations, such as slope flows and valley winds that form due to differential heating/cooling of the surface compared with the overlying atmosphere (Zardi and Whiteman, 2013). Thermally induced flows interact on multiple scales and are not isolated from each other; for example, slope flows can be eroded by the larger-scale up-valley wind (Rotach et al., 2008), and the up-valley wind can be superimposed or eroded by the larger-scale synoptic flow, dependent on its strength, the synoptic flow direction, and the local topography (e.g., Schmidli et al., 2009; Zängl, 2009; Arduini et al., 2020). Over glaciers, sharp daytime thermal contrasts between the glacier ice and overlying atmosphere at the same elevation result in downslope katabatic winds (also called glacier winds) that govern the glacier microclimate and protect the glacier ice from warm-air intrusions from aloft (Van Den Broeke, 1997b; Smeets et al., 2000; Oerlemans, 2010). The specific vertical structure of katabatic winds with a pronounced jet maximum, as well as large horizontal variability due to the interaction with other complex terrain flows, leads to a three-dimensional glacier boundary-layer structure, so that common boundary-layer assumptions—for example, those used in Monin–Obukhov similarity theory (stationarity, homogeneity, flatness, constant fluxes, and no subsidence)—do not hold (e.g., Rotach et al., 2017; Finnigan et al., 2020; Nicholson and Stiperski, 2020).

Our understanding of the glacier ABL and microclimates stems from a few field campaigns. Thetersonde observations during the PASTEX experiment conducted on the Pasterze Glacier in the Eastern Alps (Van Den Broeke, 1997b) showed that the katabatic wind layer can be up to 30 m deep and its strength depends on interactions with other thermally induced flows, such as the up-valley wind, or the ambient synoptic flow (Van Den Broeke, 1997a; Smeets et al., 1998). On cloudy, overcast days, katabatic winds are fairly weak, whereas on fair weather days a well-developed down-glacier wind can be found over large valley glaciers (Smeets et al., 2000) and even over small perennial ice fields (Mott et al., 2019). An observational study over the Vatnajökull glacier in Iceland revealed that the local persistent katabatic winds are only disturbed during strong passing storms (Oerlemans et al., 1999). More recent eddy-covariance (EC) observations over glaciers show that synoptic patterns have a strong impact on the turbulence and surface flux structure over Alpine glaciers (Litt et al., 2017; Nicholson and Stiperski, 2020). Furthermore, katabatic winds can be enforced or eroded by mesoscale circulations (Conway et al., 2021), revealing complex flow structures where the underlying atmospheric processes are still not fully understood. The observations from an array of EC towers in the Hintereisferner (HEF) experiment (HEFEX) over the HEF Glacier (Ötztal Alps, Austria) showed that the glacier tongue is mostly dominated by down-glacier winds stemming from katabatic (thermal) forcing, but these katabatic winds are frequently disturbed by lateral flow from the northwest (Mott et al., 2020). These two situations (katabatic and disturbed) exhibited different surface flux and advection patterns, but the processes behind the frequent disturbances remained unknown, because the observations from the EC
stations only give limited information about atmospheric processes aloft.

Simulations with state-of-the-art, high-resolution numerical weather prediction models offer a tool to overcome this knowledge gap. For example, high-resolution simulations are nowadays often employed to accompany week-long measurement campaigns (Udina et al., 2020; Wagner et al., 2019). However, the spatial scale of boundary-layer processes over mountain glaciers requires very fine horizontal grid spacings to resolve the relevant processes, dependent on the phenomenon of interest: Cuxart (2015) suggests horizontal grid spacings of less than 10 m for the correct simulation of stable boundary layers. Idealized large-eddy simulations (LESs) revealed that state-of-the-art models are able to simulate the physical properties of katabatic winds over glaciers under idealized settings (Axelsen and van Dop, 2009a; 2009b). A semi-idealized modelling approach with real topography by Sauter and Galos (2016) shows that surface flux patterns over a mountain glacier are strongly affected by the ambient wind and the local topography: Dependent on the larger-scale wind direction and the upstream topography, spatial differences of the surface sensible heat flux up to 100 W·m⁻² were present on the glacier. Therefore, point measurements of turbulent fluxes over mountain glaciers are likely not representative for the entire glacier. These findings were confirmed by Bonekamp et al. (2020) using semi-direct numerical simulations to investigate the impact of small-scale surface inhomogeneities on the surface flux structure over a debris-covered Himalayan glacier. All these aforementioned studies still have a semi-idealized set-up, with a realistic surface, but with an idealized atmosphere (e.g., prescribed wind conditions), whereas quickly changing conditions (e.g., shift of the synoptic wind direction, increasing/decreasing cloudiness) might have a relevant impact on the stationarity and heterogeneity of surface fluxes over mountain glaciers in reality (Litt et al., 2017). Real-case fine-scale LES process studies (i.e., Δx < 100 m) over truly complex terrain are still rare and focus on different atmospheric processes, such as precipitation patterns (Gerber et al., 2018), real-time weather prediction for sport events (Liu et al., 2020), or foehn–cold air pool interactions (Umek et al., 2021). To our knowledge, no real-case LESs of boundary-layer processes over a glacier are available to-date.

In this study, we perform LESs with the Weather Research and Forecasting (WRF) model at Δx = 48 m to answer still open questions from the HEFEX over the HEF Glacier in the Ötzatal Alps, Austria (Mott et al., 2020). Most of the time, down-glacier katabatic winds dominate the glacier surface. However, on several days, this flow is disturbed by flow from the northwest, but the physical process behind these disturbances is still not completely known. Therefore, we conduct simulations of two case-study days in August 2018 with two contrary synoptic wind directions: The first day (August 7, 2018) exhibits persistent down-glacier flow under a southwestern (SW) synoptic influence, while on the second day (August 17, 2018), the down-glacier wind is strongly disturbed by cross-glacier flows under a northwestern (NW) synoptic influence. The simulations aim to show a realistic representation of specific weather situations over a mountain glacier without idealized simplifications or assumptions. The LES set-up allows a validation of model performance with the EC station data, examination of the atmospheric processes leading to the erosion of the down-glacier wind, and an analysis of the spatial patterns and stationarity of the sensible heat flux on the glacier surface.

The paper is organized as follows: In Section 2 we introduce the area of interest, the observational data, and the model set-up; in Section 3 we provide a model validation of the general meteorological variables and analyse the wind field; in Section 4 we discuss the temporal and spatial sensible heat flux structure, the horizontal temperature advection patterns, and the heat budget. After the discussion of the results in Section 5, we present the overall conclusions in Section 6.

2 DATA AND METHODS

2.1 Area of interest and observations

The study area is centred on the HEF, a glacier located in the Ötztal Alps, Austria. HEF (Figure 1a) is an approximately 6.3-km-long valley glacier, stretching from its highest point, the Weißkugel mountain (3,738 m asl), to a height of 2,460 m, where the glacier tongue’s end is located (data from 2018). HEF has been a subject of long-term mass balance studies since the 1952–1953 season and is therefore marked as one of the World Glacier Monitoring Service’s reference glaciers (World Glacier Monitoring Service, 2017). Furthermore, HEF is part of the Rofental catchment, a long-term hydrometeorological monitoring area (Strasser et al., 2018).

The mass balance observations are accompanied by operational meteorological stations in the area. The longest dataset stems from the station at station Hintereis (StHE, 3,026 m asl), located on the southwest-facing slope observing the general meteorological variables and radiation. Data from StHE were used together with a now defunct station close to the glacier tongue for a study of the prevailing wind conditions at HEF and surroundings (Obleitner, 1994). A permanent terrestrial laser scanner (Voordendag et al., 2021) and an EC tower
measuring turbulent statistics has been located on the southern ridge, Im Hinteren Eis (IHE; 3,245 m asl) since 2016.

In August 2018, the HEFEX campaign was conducted on HEF with a main focus on studying heat-exchange mechanisms over the glacier (Mott et al., 2020). During the campaign, four additional EC stations were employed on the glacier, forming two transects: a cross-transect downwind of a small tributary glacier on the northwestern ridge, and an along-glacier transect along HEF’s tongue. The EC stations were equipped with three measurement levels of two-dimensional sonic anemometers (1.7, 1.9, and 2.9 m) to capture the katabatic down-glacier wind; furthermore, temperature and humidity sensors were installed at 1.9 m. The observed turbulence data were rotated using double rotation, detrended, and averaged over 15 min—for more details, see Aubinet et al. (2012), Stiperski and Rotach (2016), and Mott et al. (2020). The HEFEX campaign provides a dataset of turbulence data over the glacier for around a month, and an analysis of the time series shows that HEF’s glacier tongue is mostly dominated by down-glacier winds. However, during certain situations, the down-glacier wind is disturbed by flow from the northwest, together with warm-air advection. The reason behind these disturbances is still unknown after the detailed observational analysis by Mott et al. (2020); therefore, we focus our simulations on two specific case-study days: August 7, 2018, where the large-scale synoptic flow issouthwesterly and down-glacier flows dominate (SW day hereafter), and August 17, 2018, when a northwesterly synoptic flow is present and the wind direction over the glacier tongue was shifting (NW day hereafter).

2.2 | Numerical model

We employ a nested set-up of the Weather Research and Forecasting (WRF) model, version 4.1 (Skamarock et al., 2019), for the two case-study days. The set-up (Table 1) consists of four one-way nested domains with \( \Delta x = 6 \text{ km} \) (d01, spanning Europe), \( \Delta x = 1.2 \text{ km} \) (d02, spanning the Alps, Figure 1b), \( \Delta x = 240 \text{ m} \) (d03, spanning the Ötztal Alps, red rectangle in Figure 1b), and \( \Delta x = 48 \text{ m} \) (d04, spanning the Rofental catchment with HEF in the centre, Figure 1a and blue rectangle in Figure 1b), respectively. All domains have 86 vertical levels in terrain-following coordinates, where the lowest model half-level is located at \( z = 7 \text{ m} \) and the lowest 100 m of the atmosphere is represented by seven model levels. The ERA5 reanalyses (Copernicus Climate Change Service, 2017; Hersbach et al., 2020) (at a horizontal grid spacing of 30 km) are used as boundary data every 3 hr for the coarser-resolution domain d01, where d02, which mainly serves as boundary data for the two innermost LES domains, uses boundary conditions from d01.

Numerical simulations over complex terrain still face numerous challenges related to static data; for example, land use, soil representation, and topography (de Meij
and Vinuesa, 2014; Goger et al., 2016; Jiménez-Esteve et al., 2018; Golzio et al., 2021). Therefore, we replaced the model’s default static data with higher-resolution datasets, using the Harmonized World Soil Database (FAO/IIASA/ISSCAS/JRC, 2012) with a horizontal grid spacing of 30” for soil properties and the European Space Agency Climate Change Initiative land-use data (European Space Agency, 2017) with a horizontal grid spacing of 1 km for land-use categories. The Shuttle Radar Topography Mission (SRTM) 1 Arc-Second Global topography dataset (USGS, 2000) is chosen for the model topography, and three cycles of terrain smoothing with the 1–2–1 smoothing filter (Guo and Chen, 1994) are applied. After the smoothing cycles, slopes steeper than 30° remained present in the two innermost domains with sub-hectometre resolution. Therefore, the coarser domain’s topography was interpolated to the grid of the higher-resolution domains, and slopes steeper than 30° were replaced with slopes from the respective coarser-grid topography to avoid numerical instabilities. This allows the model to keep a part of terrain complexity by reducing the number of terrain smoothing cycles.

In all domains, we use the Thompson microphysics scheme (Thompson et al., 2008) and the MM5 revised surface-layer scheme (Jiménez et al., 2012), whereas for both long- and short-wave radiation, the RRTMG scheme (Iacono et al., 2008) is used with topographic shading. The Noah-MP land-surface scheme (Niu et al., 2011) is employed, which includes a “semitile” subgrid scheme to account for land surface heterogeneity and a three-layer snow model. Sensible heat fluxes are calculated according to the energy balance budgets, dependent on the emitted long-wave radiation, which is again dependent on the ground surface temperature. Glaciated areas are represented by the land-use category “ice” with a constant ground surface temperature of 0° C and a constant albedo of \( \alpha = 0.675 \). The ABL physics in d01 and d02 are parametrized by the level 3 Mellor–Yama–Nakashiki–Niino parametrization (Nakaniši and Niino, 2009), because parametrizations of the Mellor–Yama kind show a good performance for the simulation of thermally induced circulations (Wagner et al., 2014; Goger et al., 2018). For horizontal diffusion, a horizontal Smagorinsky first-order scheme is employed (Smagorinsky, 1963).

After the mesoscale run (d01 and d02) is finished, we use the \texttt{ndown} routine (a part of the WRF modelling framework) to create initial and boundary conditions from d02 output for the first LES domain d03, whereas the terrestrial fields (e.g., topography, land use) and masked surface fields are higher resolution. With this one-way nesting approach, a standalone LES for d03 and d04 is possible. Special adjustments in the static data are made for the LES runs: We use the Coordination of Information on the Environment (CORINE) dataset (European Environmental Agency, 2017) for the two innermost domains, reclassifying the land-use categories to United States Geological Survey categories following the procedure of Pineda et al. (2004). However, the CORINE dataset with \( \Delta x = 100 \) m still does not represent HEF and the surrounding glaciers in a satisfactory way. Therefore, we changed the land-use index of the \texttt{wrfinput} files for d03 and d04 with up-to-date glacier outlines from the Randolph Glacier Inventory (Pfeffer et al., 2014), and for HEF itself we use a shapefile acquired with data from an airborne laser scanning campaign in 2018 (Rainer Prinz, personal communication).

In the LES set-up, no boundary-layer parametrization is employed because we expect that a relevant part of the turbulence is already resolved. We use for both LES domains a three-dimensional subgrid-scale turbulence parametrization after Deardorff (1980) including a prognostic equation for the subgrid-scale turbulence kinetic energy (TKE). The length scale for the calculation of the

### Table 1: Overview of differences in model set-up within the four domains

| Domain       | Δx      | No. of gridpoints | Extension | Δt     | Land Use          | Atmospheric boundary-layer physics | Turbulence closure |
|--------------|--------|-------------------|-----------|--------|-------------------|------------------------------------|-------------------|
| 1 (d01)      | 6 km   | 373 × 364         | 2.283 × 2.190 km² | 20 s   | ESA-CCI (Δx = 1 km) | MYNN                               | Smagorinsky       |
| 2 (d02)      | 1.2 km | 661 × 616         | 793 × 739 km²   | 20 s   | ESA-CCI           | MYNN                               | Smagorinsky       |
| 3 (d03)      | 240 m  | 246 × 246         | 60 × 60 km²     | 0.2 s  | CORINE (Δx = 100 m) | + glacier outlines                  | Deardorff 3D      |
| 4 (d04)      | 48 m   | 251 × 251         | 12 × 12 km²     | 0.2 s  | CORINE            | + glacier outlines                  | Deardorff 3D      |

CORINE: Coordination of Information on the Environment; ESA-CCI: European Space Agency Climate Change Initiative; MYNN: Mellor–Yama–Nakashiki–Niino.
eddy viscosity follows the approach by Schmidli (2013). An analysis of the partitioning of resolved and subgrid turbulence in the Appendix shows that the model can generate a realistic turbulence structure. Instantaneous model output for the standard WRF variables is generated every 15 min, whereas 1 min instantaneous output is available for selected variables such as wind, temperature, and the surface sensible heat flux. Additionally, we use the online time-averaging tool WRF LES Diagnostics by Umek (2020) to create 15-min-averaged values of various variables for better comparability with the observations. Unless stated otherwise, all model outputs presented in subsequent sections are 15 min averages from the innermost LES domain (d04, Δx = 48 m). The closest model grid point from the respective stations is determined via Euclidian distance.

All simulations are initiated at 0300 UTC on the respective case-study day and are run for 18 hr. However, we regard the first three simulation hours as model spin-up time for turbulence and exclude them from data interpretation. This choice is a compromise between enough model spin-up time to generate realistic turbulence and the decrease of mesoscale model performance with spin-up times that are too long. This was also shown in the sensitivity run with a model initialization 12 hr earlier to check the impact of spin-up time on the meteorological fields. A comparison with observations suggested equal or slightly worse model performance (in terms of bias and the root-mean-square error) for the sensitivity run, therefore, we decided to analyse the model runs initialized at 0300 UTC.

3 | TEMPERATURE AND WIND PATTERNS

3.1 | Time evolution of meteorological variables

At first, we perform a classical model evaluation of standard meteorological quantities such as 2 m temperature, as well as wind speed and direction from the lowest model half-level. We checked the surrounding eight grid points as in Goger et al. (2018) of each HEFEX station to check on the inner-model spatial variability. The model is able to capture the variability between the stations to some extent, although the fields are more smoothed than in the observations, and station hefex-3 represents the meteorological situation over the glacier tongue well. The observed values in the following time series from hefex-3 are taken from measurement level 2, located at z = 1.9 m.

The observed 2 m temperature over the glacier tongue remains quite constant (around 8° C) on the SW day (Figure 2a), whereas minor changes are visible for the south-facing slope StHE. The observed horizontal wind speed (Figure 2c) shows differences of less than 1 m·s⁻¹ between the glacier tongue (hefex-3) and the slope aloft (StHE). The wind direction at the two locations suggests persistent down-glacier flow during the entire day (Figure 2e). This down-glacier wind direction corresponds to both the larger-scale synoptic flow on this day (SW) and to a possible katabatic glacier wind. However, if the circulation were purely thermally induced, the wind direction of StHE would show an upslope wind direction after sunrise (around 180°) indicating upslope flows. Since this is not the case, we conclude that the synoptic flow erodes the small-scale slope flows but might support the katabatic down-glacier wind. In general, the model is able to simulate the 2 m temperature at StHE with a slight warm bias (around 2° C), whereas the agreement is better at the glacier tongue (StHE). However, the horizontal wind speed is overestimated by the model for both locations by about 2 m·s⁻¹ before noon, and the wind speed increases in the afternoon up to 8 m·s⁻¹ with progressing simulation time. The wind direction is well simulated at both stations, suggesting that the model is able to simulate the general wind structure over the glacier valley.

The second case-study day is dominated by northwest synoptic flow; therefore, we show time series from the ridge EC tower IHE, since its location is representative of the flow across the glacier valley. The temperature difference between IHE and the hefex-3 station on the glacier tongue is high due to the larger height difference (Figure 2b), and the temperature structure over the glacier is disturbed between 0700 and 1200 UTC. The observed wind speed is smaller than on the SW day and does not show any specific diurnal cycle. However, the wind speed at the glacier tongue is quickly changing in accordance with frequent changes in the wind direction (Figure 2f). This hints that a heavily disturbed down-glacier flow is present on the NW day. The model is able to simulate the 2 m temperature fairly well at IHE, but the rapid changes above the glacier surface are not captured by the model. As on the SW day, modelled wind speeds are overestimated at both stations (Figure 2d). At the glacier tongue, the model also suggests rapid changes in wind direction, but with an offset of about 50°. In the model, most of the day a cross-valley flow is present, whereas the observations from hefex-3 suggest a heavily disturbed down-glacier flow with quickly changing wind directions. One reason for this might be the different heights from the observations (around 2 m) and the height from the lowest model half-level (around 7 m), suggesting that the model is not able to simulate small-scale processes on the glacier tongue. However, the model simulates the dominating larger-scale synoptic flow direction (northwest) at the
more exposed ridge station IHE well. None of the wind directions correspond to possible thermally induced circulations, and we conclude that the synoptic flow dominates the wind patterns over the glacier surface. This hints that the model simulates a process related to the cross-glacier flow, but too strongly over the glacier tongue. Since the conditions on this day are dynamically driven, it is possible that channelling effects or gravity waves influence erode the glacier boundary layer in the model. The time series alone cannot, of course, confirm the processes at work, and a more detailed analysis of these physical processes will be provided in the following sections.

To summarize, the model is able to simulate the general meteorological variables on the glacier and its surroundings in a realistic way, although we have to point out that the wind speed is systematically overestimated and the fast, chaotic changes in wind direction on the NW day have a directional shift in the model. We decide to focus for both days on the time period from 0600 to 1200 UTC, since during this time frame the model performs best. This allows a process-based interpretation of model data. If not stated otherwise, all results presented in the following sections are from this time frame of 6 hr.

3.2 Horizontal spatial patterns of potential temperature

The potential temperature provides information about the thermal stratification and, together with the wind
field, the spatial variability of the glacier boundary-layer structure (Figure 3). In the early morning of the SW day, the southeast-facing slopes are already heated, while potentially colder air is present over the ice surfaces (Figure 3a). Plumes of potentially warmer air are transported towards the glacier tongue with the westerly wind, while small plumes of potentially colder air are visible downwind of the smaller tributary glaciers. During the next 4 hr (Figure 3b,c), the synoptic wind direction shifts to southwesterly, supporting the persistent down-glacier wind. However, when the larger-scale wind direction turns towards southerly, the body of potentially colder air over the glacier becomes less pronounced (Figure 3d) and weakens in the late afternoon (not shown). In general, the SW day case study shows a heterogeneous potential temperature structure and large horizontal gradients between the ice surfaces and their surroundings.

The potential temperature structure is entirely different on the NW day. Although the glacier surfaces are still potentially colder than their surroundings, the horizontal temperature gradients are weaker (Figure 3e–h) and the potential temperature field is spatially homogeneous. During most of the time, a strong northwesterly cross-glacier flow is present, perpendicular to the down-glacier wind direction, thus preventing a small-scale glacier wind from forming. The cross-glacier flow often shifts its wind direction slightly and turns towards up-glacier in the afternoon (Figure 3h). The reason for the fast-changing wind direction over the glacier tongue is discussed in Section 3.3.

### 3.3 Vertical structure

#### 3.3.1 Cross-sections

Alpine glaciers are not isolated from their environment, because the surrounding complex terrain has a strong influence on stationarity, heterogeneity, and the general turbulence structure. We explore the related atmospheric processes with vertical cross-sections of potential temperature, the wind vectors, and TKE across the glacier valley (Figure 4). On the SW day at 0600 UTC, the potential temperature field shows stable stratification directly over the glacier surface with large local differences (Figure 4a). The minimum in potential temperature coincides with the body of potentially colder air in Figure 3b,c. These local stable layers directly over the glacier surfaces remain present for the next hours (Figure 4b,c), but the
stratification weakens subsequently, which is also visible in the spatial structure of potential temperature in Section 3.2. TKE maxima are present above glacier surfaces, suggesting locally generated turbulence, but also at upper levels, related to possible interaction of the larger-scale flow with the topography. On the NW day, the atmosphere in the glacier valley is strongly stably stratified (Figure 4e,f). The northwesterly synoptic flow passes over the ridge located northwest of HEF, leading to a subsequent steepening of the isentropes and the formation of a gravity wave over the glacier valley with a visible elevated maximum of TKE around 200 m over the glacier (Figure 4e,f). Around 1200 UTC, the gravity wave breaks (Figure 4h), and strong mixing and severe turbulence with TKE values higher than 10 m²·s⁻² are present over the glacier tongue. The breaking gravity wave in the lee of the upstream steep topography (Figure 4h) is the main reason for the continuous change in wind speed and direction in the time series of the NW day (Figure 2d,h).

The along-glacier cross-section shows that gravity-wave-like structures are also present on the SW day from the early morning at upper levels (Figure 5a). However, the atmosphere is less stratified (than on the NW day) and the along-glacier slope is less steep than the northwest ridge; therefore, conditions for gravity wave formation are less favourable. At 1000 UTC (Figure 5c), gravity wave breaking is visible at levels 1,000 m above the glacier surface; however, HEF is not affected by these upper-level gravity waves, and the local glacier boundary layer remains stably stratified with minima of potential temperature over the glacier surface (Figure 5a–d). On the NW day, the along-glacier cross-section reveals strong stability over the accumulation area of HEF above 3,000 m (Figure 5e,f), which is not present at the glacier tongue. The major reason for this weakening is likely the strong cross-glacier flow related to the gravity wave, eroding the local glacier boundary layer at the glacier tongue. When the gravity wave is breaking, the associated strong mixing entirely dominates the local boundary layer over the glacier tongue, while a weak stable layer remains above the accumulation area of HEF (Figure 5g).

These two case studies show that background flow direction, atmospheric stability, and the upstream topography have a relevant impact on the formation and strength of gravity waves over the glacier valley. Furthermore, these factors determine whether the glacier tongue is affected by gravity-wave breaking and the related severe turbulence.

**FIG URE 4** Vertical cross-section along the light-green line in Figure 1 for the SW day (upper row) and the NW day (lower row). White lines along the topography indicate glaciated areas. Black contour lines (every 0.25 K) show the potential temperature, wind vectors indicate the cross-valley wind speed, and colours indicate turbulence kinetic energy [Colour figure can be viewed at wileyonlinelibrary.com]
3.3.2 Vertical profiles

The cross-sections from the previous section gave an overview of the atmospheric processes contributing to the glacier boundary-layer structure. In the following paragraphs, we have a more detailed look at the thermal stratification and wind speed with vertical profiles (Figure 6) from specific points over the glacier (highlighted in Figure 5). On the SW day at 0800 UTC (Figure 6a), the potential temperature profile reveals a stable stratification at the lowest 20 m at all chosen points, while the upper points on the glacier (p1 and p2) show a stronger stability than the glacier tongue points (p3 and p4). The stable stratification remains present until 1200 UTC at all points (dashed lines), while a neutral layer is present above the near-surface stable layer. The vertical profile of horizontal wind speed (Figure 6b) suggests mainly constant wind speeds with height, although a very weak local maximum is present at around 20 m, at the same height as the stable layer in the potential temperature profile. The synoptic wind direction is the same as the down-glacier flow; therefore, we can assume that the down-glacier wind is supported by the synoptic flow. The strongly stable stratification and the very weak jet maximum hint that the down-glacier flow might have partly katabatic origin.

Judging from the vertical profiles, we cannot conclude without a doubt whether the forcing of the local layer above the glacier is katabatic in the model. The simulated local stable layer is not entirely comparable to observed katabatic flows during the HEFEX campaign, which had a wind-speed maximum of 2 m above ground (Mott et al., 2020, their figure 4). The lowest model half-level is located at 7 m, and the lowest 100 m of the atmosphere in the model consists of seven model levels. Thus, it is likely that the model is not able to resolve the observed shallow katabatic glacier winds with the vertical grid. However, the model is able to simulate a weak jet maximum together with a stable layer over the glacier, suggesting that the modelled katabatic layer is much deeper than in reality.

On the NW day (Figure 6c,d), the vertical profiles show a different structure: According to the potential temperature profile at the two lower points p3 and p4 (0800 UTC, glacier tongue), the atmosphere above the glacier surface is neutrally stratified, whereas at the two points in the accumulation area of HEF (p1 and p2) it remains stable, as on the SW day. Atmospheric stability decreases with time, so that at 1200 UTC a slightly unstable layer is present above the glacier tongue. The vertical profiles show that a stable boundary layer over the glacier (profiles from p3 and p4 at 0800 UTC) coincides with a weak jet maximum.
While the stable layer in the accumulation area of HEF remains persistent until 1200 UTC, the gravity wave and the induced cross-glacier flow erode the glacier boundary layer and strong turbulent mixing leads to a neutral stratification over the glacier tongue (p3 and p4). This shows that even dynamically induced atmospheric processes such as the gravity wave can be very localized, and therefore atmospheric conditions at the glacier tongue are not automatically representative for the entire glacier.

The vertical structure of resolved TKE and its budget terms also gives insights into the vertical ABL structure over the glacier tongue at hefex-3 (Figure 7). The TKE budget equation can be written after Stull (1988b):

\[
\frac{\partial \overline{e}}{\partial t} + \overline{U_j} \frac{\partial \overline{e}}{\partial x_j} = \frac{\partial}{\partial x_j} \left( \frac{g}{\theta_v} \left( u_i' \theta_v' \right) \right) - \frac{u_i' u_j'}{\overline{U_j}} \frac{\partial \overline{U_i}}{\partial x_j} - \frac{\partial (u_i' \overline{e})}{\partial x_j} - \frac{1}{\rho} \frac{\partial (u_i' p')}{\partial x_j} - \varepsilon,
\]

where capital letters with overbars denote mean quantities, and small letters with primes refer to turbulent fluctuations; \( \overline{e} \) is TKE, \( U \) is the mean wind speed, \( g \) is the acceleration due to gravity, \( \theta_v \) is virtual potential temperature, \( \rho \) is air density, and \( p \) is pressure. On the left-hand side are the local TKE tendency and advection with the mean flow, and on the right-hand side we find the thermally driven buoyancy production/consumption term, the mechanical shear production term, the turbulent transport, the pressure correlation term, and the TKE dissipation rate \( \varepsilon \). For our analysis, we consider the TKE advection, the vertical buoyancy production, the shear production, and the dissipation rate from the averaged model output. We have to note that, in sloping terrain, horizontal contributions to the buoyancy production might be non-negligible (Oldroyd et al., 2016), but we cannot assess them with our present model set-up. Turbulent transport and the pressure correlation term mainly serve as a redistribution mechanism of TKE, but we do not consider them in our analysis.

On both case-study days, the TKE magnitude increases between 0800 UTC and 1200 UTC (Figure 7a,c). On the SW day, the vertical profile of TKE is almost constant with height at 0800 UTC, while at 1200 UTC, a local TKE maximum is present at around 20 m, coincident with the stable layer (Figure 6a,b). The corresponding TKE budget terms
suggest that the negative buoyancy term acts as a sink of TKE close to the ice surface, and that shear production is the major contributor to the TKE budget, counteracted by the dissipation, implying that the processes contributing the TKE are mostly local.

On the NW day, the vertical profile of TKE at 0800 UTC is also constant with height, and the TKE budget structure is similar to the SW day. However, at 1200 UTC the TKE values increase dramatically up to $6 \text{ m}^2 \cdot \text{s}^{-2}$, also at upper levels. The TKE budget terms suggest again that shear is the dominant production term, with values up to $0.02 \text{ m}^2 \cdot \text{s}^{-3}$, whereas the buoyancy term is closer to zero. However, the TKE advection is of almost equal magnitude to the shear production term at upper levels. As shown in other studies of strong dynamically induced situations with breaking gravity waves (e.g., Večenaj et al., 2010; 2012), the TKE advection is a non-negligible term of the TKE budget. The NW-day TKE structure is therefore not only influenced by local processes acting in the vertical but also by larger-scale three-dimensional dynamics, visible in the generally high TKE values, but also in the TKE budget structure with TKE advection, with the mean flow being a relevant term.

4 | BOUNDARY-LAYER PROCESSES

The general model evaluation (Section 3) showed that the model is, in general, able to simulate the overall atmospheric structure over HEF. Shortcomings of the model include the too deep down-glacier wind on the SW day and the overly strong gravity wave on the NW day, visible in the shifted wind directions in Figure 2f. However, the model can simulate mesoscale processes contributing to the small-scale ABL structure over the glacier. Therefore, we continue with studies of the boundary-layer processes over the glacier, and in the following we will see that the relevant physical processes are simulated successfully.

4.1 | Sensible heat flux structure

The previously described meteorological phenomena are spatially heterogeneous, and the wind speed and direction might especially have an impact on sensible heat flux variability over the glacier. However, according to the energy balance, sensible heat flux is mostly determined
by the net radiation and the repartition with the latent heat flux. Time series of net radiation from the SW day suggest some cloudiness, whereas the model simulates an almost cloud-free day (Figure 8a). This also manifests itself in the sensible heat flux observations, which are generally smaller than the simulated sensible heat flux. However, a direct quantitative comparison between observed and modelled surface fluxes is a challenge because the energy balance is closed in the model but not in the observations (Mauder et al., 2020). However, the sign of the sensible heat flux is negative both in the observations and model output, suggesting that the ice surface is represented correctly in the model. The net radiation on the NW day shows a better agreement between the model and observations. The weather conditions are cloud free before noon, but clouds establish in the afternoon. The sensible heat fluxes are smaller than on the SW case-study day in both the observations and the model, but the sign is also negative over the glacier surface on both case-study days.

The animation from 1 min model output (see Video S1) shows a very high temporal variability of the sensible heat flux. For the SW day, with increasing daytime turbulence, plumes of local sensible heat flux maxima travel down the glacier tongue with the mean wind. The existence of such plumes is observed for the rest of the day, suggesting a high spatio-temporal variability. As for the NW day, the breaking gravity wave is the largest driver of high temporal variability in sensible heat flux. This raises the question whether this non-stationarity is also present in the observations, and if it has a diurnal cycle. For this purpose, we divide the 1 min sensible heat flux time series (from both model output and observations) into 15 min intervals.
and calculate the non-stationarity ratio (NR) after Mahrt (1998):

\[ NR = \frac{\sigma_{k,\text{btw}}}{RE_k}, \]

(2)

where \( \sigma_{k,\text{btw}} \) is the within-record standard deviation (btw) of the 1 min time series within the 15 min intervals, and \( RE_k \) is the standard error of all the 15-min-averaged flux results (\( k \) is the number of intervals). When the non-stationarity ratio is approximately unity, the time series can be assumed to be stationary, because the variances of the 15-min averages do not exceed their standard error. For values larger than 2, non-stationarity can be assumed. We are aware that stationarity depends on the choice of the averaging period, but similar averaging choices were made for the non-stationarity ratio of sensible heat fluxes under gravity wave influence during the T-REX experiment (Večenaj and De Wekker, 2015).

The 1 min instantaneous values of the non-stationarity ratio (Figure 8) for both days from the model and observations show mainly values around or above 2, suggesting non-stationary behaviour. Whereas the non-stationarity ratio gradually increases with increasing daytime turbulence on the SW day, the non-stationarity ratio on the NW day is, in general, higher without a diurnal cycle: Non-stationarity ratio maxima are already present at 0800 UTC, possibly related to the strong gravity-wave activity, which continues until 1200 UTC (see Figure 4f). Furthermore, since the non-stationarity ratio is higher in the model than in the observations, the model seems to simulate a stronger gravity wave influence. The higher the non-stationarity, the more mesoscale contributions influence the surface sensible heat flux. The high non-stationarity in the time series might partly explain the large differences between observed and modelled sensible heat fluxes.

Aside from temperature, the horizontal wind speed also influences the strength of sensible heat fluxes (Figure 9e,f). The calculated linear correlation coefficient \( r \) between the sensible heat flux and the wind speed suggests a strong negative linear correlation in both the model and observations of the SW day (Figure 9e). Although the linear correlation coefficient is smaller on the NW day (Figure 9f), the general pattern remains: The higher the horizontal wind speed, the stronger the sensible heat flux is; it has to be mentioned at this point that the model tends to overestimate the horizontal wind speed (Figure 2), leading to higher sensible heat fluxes than

**Figure 9**  Averaged model output of the surface sensible heat flux (colours and dashed lines) and horizontal wind vectors from the lowest model level over glaciated surfaces. Sensible heat fluxes over ice-free surfaces are not shown. The top and bottom rows show data from the SW day and the NW day, respectively, at four different times [Colour figure can be viewed at wileyonlinelibrary.com]
observed (Figure 8a,b). These findings are in accordance with the bulk formulation of the sensible heat flux (Stull, 1988a, their equation 7.4.1d): The central assumption is that the sensible heat flux is linearly proportional to the wind speed and to the temperature difference between the surface and the chosen height. Over a melting glacier, the surface temperature is 0°C, and the height-dependent temperature difference becomes very small (due to the prevailing glacier boundary layer). Therefore, sensible heat flux strength also depends on wind speed. This is visible in the different linear correlation coefficients on the two different days: On the SW day, glacier boundary-layer processes are dominant, leading to a stronger correlation between sensible heat flux and wind speed, whereas on the NW day, the correlation is less pronounced due to the erosion of the local glacier boundary layer. Despite the differences in absolute magnitude, the linear relationship between the wind speed and the sensible heat flux is similar in the model and observations, suggesting that the model is able to reproduce the relevant physical processes.

The observations only show the sensible heat flux at a point, but the model output suggests that the averaged sensible heat flux over the glacier surface is spatially highly heterogeneous on both case-study days. On the SW day at 0600 UTC (Figure 9a), the spatial sensible heat flux structure is connected to the horizontal wind speed and direction, and the weak cross-wind with potentially warmer air from the slopes is also visible with the higher sensible heat flux over the glacier tongue. This changes at 0800 UTC (Figure 9b), visible with weaker sensible heat fluxes over the glacier tongue and streaks in its spatial structure at the same location as the body of potentially colder air in Figure 3b. At 1000 UTC, the streaks are still visible, but the sensible heat flux is now stronger; this is connected to the weakening stratification over the glacier tongue. At noon (Figure 9d), together with the increasing horizontal wind speed, the sensible heat flux over the glacier tongue reaches its maximum. On the NW day, the sensible heat fluxes are generally smaller than on the SW day, but they show an atypical pattern: At 0600 UTC, a local sensible heat flux maximum is already present at the glacier tongue downwind from the south-facing slope (Figure 9e). This structure is visible for the next 6 hr (Figure 9f-h). When, at 1200 UTC, the gravity wave breaks and severe turbulence is present over the glacier tongue, the sensible heat flux is strongest (Figure 9b,h). While on the SW day the persistent down-glacier flow leads to a specific spatial structure (streaks) over the glacier tongue, the strong mesoscale influence with the gravity wave also leads to localized high sensible heat flux values over the glacier tongue only. To conclude, the magnitude of the sensible heat flux is mostly governed by the net radiation and the surface properties; the horizontal wind speed has an essential influence on the spatial pattern of the sensible heat flux over the glacier.

4.2 Temperature advection patterns and heat budget

The heterogeneous spatial structure of the wind field and potential temperature determine the spatial patterns of the horizontal temperature advection (Figure 10). On the SW day, the body of potentially colder air (Figure 3) coincides with horizontal cold air advection over the glacier. At 0600 UTC, when the potentially colder air is pushed towards the North-facing slope, the glacier tongue is under the influence of weak warm-air advection. This changes at 0800 UTC, when the down-glacier wind speed increases and the glacier tongue is under persistent cold-air advection. The streamlines show that the source of the cold air is mainly located at the glacier accumulation area. Until 1200 UTC, the thermal stratification decreases (Figure 6a), but the glacier tongue remains under the influence of cold-air advection. This suggests that if the wind direction is down-glacier, the glacier remains in the presence of cold-air advection. The NW day shows a different temperature advection pattern: The cross-glacier circulation leads to advection of warmer air over the glacier tongue. The absolute values of the warm-air advection over the glacier are, however, much smaller than the absolute values of the cold-air advection of the SW day. The wind direction determines the type of advection over the glacier tongue: The upstream source region is the heated slope, and, supported by the gravity wave, the glacier tongue is under the influence of, albeit weak, warm-air advection. On the other hand, the glacier accumulation area, where local stable stratification is present (Figure 6d) remains under very weak cold-air advection.

We can now examine whether the horizontal temperature advection is driven mostly by wind speed or temperature gradients. Therefore, we compare the model output with observations from the EC stations. Dependent on the case-study day, we either investigate the impact of along-glacier (SW day) or cross-glacier (NW day) flows on horizontal temperature advection. We calculate the observed horizontal temperature advection between stations hefex-3 and hefex-4 (for the SW day) and between stations hefex-3 and hefex-2 (for the NW day):

$$T_{ADV} = U \frac{\Delta T}{\Delta s}. \quad (3)$$

where $\Delta s$ is the distance between the stations, $\Delta T$ is the temperature difference, and $U$ is the average horizontal wind speed. The observations (Figure 11a) suggest that the glacier tongue is under the influence of horizontal
FIGURE 10  Averaged model output of the averaged horizontal temperature advection (colours) and streamlines derived from the wind field from the lowest model level. The top and bottom rows show data from the SW day and the NW day, respectively, at four different times [Colour figure can be viewed at wileyonlinelibrary.com]

FIGURE 11  Scatter plots of the SW day (blue) and the NW day (orange) from observations (upper row) and model output (lower row) of location hefex-3 for (a,c) horizontal temperature advection and wind direction and (b,d) horizontal temperature advection and wind speed [Colour figure can be viewed at wileyonlinelibrary.com]
cold-air advection, when the wind direction is predominately down-glacier (200°). However, when the wind direction shifts towards 250°, the temperature advection sign changes. The same patterns are visible in the model output (Figure 11c), although there is a more gradual shift in wind direction; this is likely related to the synoptic wind influence. Since there is likely a katabatic glacier wind present in the observations, and the model struggles to resolve the glacier wind adequately, the synoptic influence on the glacier ABL is larger than in the observations, leading to the wind direction shift. However, both the type and the magnitude of the observed and modelled horizontal temperature advection are similar. On the NW day, both observations and the model suggest that the glacier tongue is under continuous horizontal warm-air advection, together with the shift in modelled wind direction associated with the strong gravity wave (Figure 4). These results suggest that if a down-glacier wind is present (corresponding to a wind direction 200°), horizontal cold-air advection dominates. However, if the wind direction shifts and the source area changes from the ice surface to the surrounding, heated, rocky terrain, horizontal warm-air advection dominates on the glacier tongue. Aside from the wind direction shift in the model (discussed in Section 3), the magnitude of the horizontal temperature advection is similar in the observations and the model. In general, the horizontal cold-air advection connected to the down-glacier flow on the SW day is stronger than the horizontal warm-air advection under intermittent conditions on the NW day (Mott et al., 2020).

The relation between the wind speed and the horizontal temperature advection is shown in Figure 11b,d. The pattern of the scatter is similar in the observations and the model, but the overestimated wind speeds in the model lead to a shift in the pattern towards higher wind speeds. However, the strength of the advection seems to be related to the wind speed strength; for example, strong cold-air advection the SW day is related to higher wind speeds. On the other hand, the type of advection (cold-air or warm-air advection) is then rather governed by the horizontal temperature gradient.

Horizontal temperature advection is only one term of the temperature tendency equation in the boundary layer (Wyngaard, 2010; Haid et al., 2020):

$$\frac{\partial \theta}{\partial t} = -U \frac{\partial \theta}{\partial x} - V \frac{\partial \theta}{\partial y} - W \frac{\partial \theta}{\partial z} - \frac{\partial \theta' \theta'}{\partial z},$$

(4)

consisting of temperature advection with the mean wind in all three directions (ADVx, ADVy, ADVz) and the vertical heat flux divergence (vHFD). The radiative flux divergence is neglected, because it is considered to be small. Vertical profiles of the three temperature advection terms (Figure 12a) from the SW day at 0800 UTC and 1000 UTC suggest that cold-air advection dominates over warm-air advection. Together with the decreasing vHFD over hefex-3, this results in a net cooling of the atmosphere over the glacier tongue (Figure 12b). However, at 1200 UTC, the net positive advection contribution in the heat budget (Figure 12b) leads to a net warming of the atmosphere over the glacier. This heat budget analysis shows that a down-glacier wind with horizontal cold-air advection (Figure 11b) is not a guarantee for general cooling of the atmosphere over the glacier. On the NW day, cold-air advection dominates the glacier tongue at 0800 UTC (Figure 12c). As on the SW day, this leads to a net cooling at 0800 UTC (Figure 12d). Two hours later (1000 UTC), the advection part of the heat budget turns positive, leading to an overall warming tendency over the glacier tongue. At 1200 UTC, the situation changes: Whereas the overall advection turns positive, the overall heat budget remains negative. The breaking gravity wave mixes potentially colder air from aloft (originated from the potentially colder air over the large glacier upstream of HEF) into the glacier boundary layer. This heat budget analysis shows that it is not possible to judge only from horizontal advection components on the atmospheric warming or cooling above the glacier, because both the vertical temperature advection and the vHFD are terms of equal magnitude. Therefore, our numerical modelling approach gives useful insights to the individual components of the heat budget over the glacier.

5 | DISCUSSION

The simulation of boundary-layer processes in complex terrain is still a major challenge for numerical weather prediction models (Edwards et al., 2020) due to insufficient terrain representation (Wagner et al., 2014), too coarse land-surface datasets (Golzio et al., 2021), or turbulence parametrizations developed based on assumptions for horizontally homogeneous and flat terrain (Goger et al., 2018; 2019). These problems can partly be avoided by increasing horizontal and vertical resolution towards horizontal grid spacings where the largest eddies can be expected to be already resolved (Chow et al., 2019). Our LES set-up at Δx = 48 m has a realistic representation of the topography surrounding the glacier (within the scope of numerical stability); and furthermore, the horizontal mesh size also allows a realistic representation of the land use and the ice surface (Figure 1).

One of the largest questions when setting up the model was whether WRF is able to simulate small-scale glacier boundary-layer processes. Cuxart (2015) states that
A horizontal grid spacing of $\Delta x = 5$ m might be sufficient to simulate processes in the stable mountain boundary layer. Although we use a coarser grid spacing ($\Delta x = 48$ m) in our simulations, our results suggest that the representation of the glacier boundary layer is partly possible: On both days, the model simulates a stably stratified layer over the ice surfaces, which remains persistent during the daytime (SW day) or is eroded by the gravity wave at the glacier tongue (NW day). The outer domain ($d03, \Delta x = 240$ m) is not able to simulate the stable layer over the ice surfaces (not shown), although the vertical grid spacing is the same. This implies that the horizontal mesh size has an essential influence on the correct simulation of the spatial variability of ABL processes over the glacier.

Our current set-up is not able to resolve all relevant length scales equally well. For example, on the NW day, the synoptic wind direction is well simulated at the ridge station IHE (Figure 2f), and the model simulates a strong, breaking gravity wave over the glacier tongue. At the glacier tongue, though, the modelled gravity wave easily erodes the glacier ABL in the simulation, whereas in the observations a weak but strongly disturbed katabatic wind layer is present. This would explain the wind direction shift between modelled and observed wind direction in (Figure 2f). The simulated gravity wave influence is stronger than in reality, visible in the high TKE values (more than $10 \text{ m}^2\text{s}^{-2}$ in Figure 4h) and the higher non-stationarity of the sensible heat flux than in the observations (Figure 11d). The HEFEX stations did not record such high TKE values over the glacier tongue on the NW day, but longer TKE time series from station IHE reveal that comparably high TKE values (around $10 \text{ m}^2\text{s}^{-2}$) are possible in connection with northwesterly synoptic flows. Furthermore, aircraft observations by Jiang and Doyle (2004) over the Ötztal Alps suggest that our simulated TKE values (around $10 \text{ m}^2\text{s}^{-2}$) are realistic. The predictability of mountain waves decreases with increasing nonlinearity and terrain steepness (Doyle et al., 2011), and strong gravity waves are often a challenge for the simulation of dynamically driven flows over mountainous terrain (Gohm et al., 2004; Umek et al., 2021).
The different scales in the ABL over the glacier are not isolated from each other, and scale interactions are common in both case-study days. Small-scale processes over the glacier, such as local stable layers, are still underrepresented in the model, and therefore less resilient towards mesoscale influence, although they might be more persistent in reality. For example, a down-glacier katabatic wind during summer might have a jet maximum of only 2 m above ground—as in Mott et al., 2020—which cannot be resolved by the model’s vertical grid (lowest model half-level at 7 m). However, especially on the SW day, the down-glacier wind is supported by the synoptic flow direction, resulting in a deeper flow with a weak jet at around 20 m. Though this might be partially attributed to the model’s vertical grid, Smeets et al. (1998) also observed deeper down-glacier winds when the katabatic glacier wind was aligned with the synoptic flow, while low-frequency disturbances in the turbulence spectra were present. The influence of the synoptic flow direction on the turbulence structure was also observed over the Saint-Sorlin Glacier in the French Alps (Litt et al., 2017). Our results from the SW day agree with the aforementioned studies and show that the synoptic flow direction governs the spatial structure of the glacier boundary layer. Owing to the glacier’s high altitude, synoptic flows interact with the topography and plunge into the comparably shallow glacier valley, eroding the glacier boundary layer. This raises the question of whether conditions a glacier is able to maintain its own microclimate. Purely thermally driven situations (i.e., with an undisturbed glacier wind) as described in Smeets et al. (2000), Nicholson and Stiperski (2020), and Mott et al. (2020) are still possible under fair-weather conditions with weak background wind, but the synoptic wind direction and the local topography are major factors for the spatial sensible heat flux structure on the glacier and should not be disregarded.

Other studies comparing LES studies with high-resolution observations report good model performance, so that the WRF model can be used as a tool for better process understanding (e.g., Muñoz-Esparza et al., 2017; Babić and De Wekker, 2019; Hald et al., 2019; Conolly et al., 2021), and for long-term simulations of more than 1 week (Wagner et al., 2019; Udina et al., 2020). However, in our complex terrain setting, the model still overestimates the horizontal wind speeds on both days, in accordance with other LES studies with WRF over truly complex terrain with a similar bias (Gerber et al., 2018; Liu et al., 2020; Umek et al., 2021). One reason for this is the differences in measuring heights of the instruments (2 m) versus the lowest model level at 7 m. Although Wagner et al. (2014) point out that relevant boundary-layer processes are resolved when the valley is represented by at least 10 grid points, model topography still differs from reality due to smoothing. The realistic simulation of the thermally induced circulation, katabatic winds, boundary-layer separation, and gravity waves also depend on the topography steepness and/or shape (Smith and Skyllingstad, 2005; Doyle et al., 2011; Wagner et al., 2015; Prestel and Wirth, 2016). Furthermore, the restriction of model topography not having slope angles over 40° also reduces the complexity of the ABL flow (e.g., no bluff-body boundary-layer separation in the lee of very steep mountains). This well-known problem of vertical grids over complex terrain might be solved with the implementation of immersed boundary conditions (Lundquist et al., 2010).

Although there are still shortcomings and biases present in the model (e.g., wind direction, overestimated wind speeds and sensible heat fluxes), the simulation output provides information about the mesoscale physical processes influencing the glacier ABL structure. This is possible because the processes affecting the glacier ABL structure have a larger scale and are related to topography; for example, the gravity wave on the NW day. The information is often missing from the aforementioned purely observational studies (Smeets et al., 2000; Litt et al., 2017; Nicholson and Stiperski, 2020) and is one of the major advantages of employing this high-resolution LES set-up. Furthermore, the simulation results agree well with the findings from the HEFEX campaign (Mott et al., 2020): The wind speeds over the glacier are higher when a down-glacier wind is present than during disturbed situations (Figure 2), the sensible heat flux is linearly dependent on the wind speed (Figure 8), and the connection between the wind direction and the horizontal temperature advection is also present (Figure 10). The large benefit of the simulations is, however, the information of spatial fields (both horizontal and vertical) of boundary-layer variables such as the sensible heat flux, TKE, and temperature advection, which cannot be observed by point measurements. Nonlinear processes such as gravity waves strongly contribute to non-stationarity and spatial heterogeneity of the sensible heat flux over the glacier, governing the spatial heterogeneity besides the elevation dependence (Greuell et al., 1997). This raises questions as to whether linear relationships, single point measurements, and flux-profile methods basing on Monin–Obukhov similarity theory are representative for surface energy balance modelling of mountain glaciers (Denby and Greuell, 2000; Sauter and Galos, 2016).

6 | SUMMARY AND CONCLUSIONS

We conducted the first real-case LESs with the WRF model at Δx = 48 m over the HEF Glacier located in the Ötztal Alps, Austria, for two case studies with different synoptic
forcings: August 7, 2018, when southwesterly flow was present; and August 17, 2018, when northwesterly flow prevailed. The model results are validated with EC towers on the glacier surface (HEFEX) and permanent stations on surrounding ridges and/or slopes. Dependent on the direction of the synoptic flow, different wind patterns are present over the glacier and affect the spatial structure of the glacier boundary layer and the sensible heat flux. This leads us to the following conclusions:

1. On the SW day, the model is able to simulate the temperature and wind patterns over the glacier in a satisfactory way. Though the horizontal wind speed is generally overestimated by the model, the wind direction is simulated well for the SW day. This allows a further process-oriented analysis of the model output. On the NW day, the model simulates a strong cross-glacier wind, whereas the observed wind direction suggests a heavily disturbed down-glacier wind. However, the down-glacier wind from the observations is so shallow that the model cannot resolve it with its current horizontal and vertical grid spacing.

2. On the SW day, the simulated potential temperature structure reveals that a prevailing body of potentially colder air is present over the ice surfaces, accompanied by persistent down-glacier winds. On the NW day, strong cross-glacier flow leads to a more homogeneous potential temperature structure without horizontal temperature gradients between the glacier and surroundings.

3. Cross-sections along and across the glacier from the SW day reveal that small-scale processes such as differential heating between the glacier and surroundings are present and sources of TKE are likely local. Furthermore, the synoptic wind direction supports the down-glacier flow. On the other hand, on the NW day, a strong gravity wave forms on the upstream steep slope, resulting in severe shear-driven turbulence and an erosion of the boundary layer on the glacier tongue, though the accumulation area of the glacier is less affected.

4. Vertical profiles of potential temperature and wind speed reveal that a stably stratified layer is present over the glacier surface on the SW day. TKE values are generally low, and weak shear is the only TKE production source. On the NW day, the layer above the glacier tongue is neutrally stratified without a distinguishable local layer, and the TKE values are high, especially also at upper levels. This is connected to strong shear production; but TKE advection also acts as an important contributor to elevated TKE maxima.

5. The model overestimates the sensible heat flux magnitude over the glacier tongue compared with the observations. However, model and observations agree well on the linear dependency of the sensible heat fluxes on the wind speed and the high non-stationarity of sensible heat fluxes over the glacier. On the SW day, non-stationarity is mostly related to a gradual increase in daytime turbulence. On the NW day, non-stationarity is higher than on the SW day owing to the strong mesoscale influence by the breaking gravity wave. The sensible heat flux structure over the glacier is very heterogeneous and strongly influenced by the wind field.

6. On the SW day, together with the down-glacier wind, mainly horizontal cold-air advection is present over the glacier tongue, whereas on the NW day the cross-glacier flow induces mainly warm-air advection. The source region (i.e., either glacier accumulation zone or the surrounding slopes) influences whether cold-air or warm-air advection is present over the glacier tongue, whereas the horizontal wind speed mostly determines the advection strength. However, an overall analysis of the heat budget suggests that the horizontal advection terms have a small influence on whether overall cooling or heating of the near-glacier atmosphere is present, because the vertical temperature advection and vertical heat flux divergence play a relevant role.

7. The synoptic flow, the upstream topography, and atmospheric stratification have a major influence on whether a local glacier boundary layer is able to form (SW day) or if it is completely eroded (NW day), whereas even dynamically induced processes can be highly heterogeneous in space due to the underlying complex terrain.

8. The horizontal and vertical resolution of our current model set-up is not yet high enough to resolve very shallow katabatic down-glacier winds. However, the model is successful in simulation of a daytime stable boundary layer over the glacier surfaces and the mesoscale processes that contribute to the glacier wind's spatial structure or possible erosion.

The HEF-LES set-up is a valuable tool to investigate boundary-layer processes over the glacier. The high-resolution LES can deliver three-dimensional spatial fields of meteorological variables, which allows us to potentially understand the processes contributing to the strong heterogeneity in surface fluxes over the glacier. The synoptic wind direction strongly governs the local glacier ABL formation, and therefore it might be able to predict local conditions on the glacier when only the large-scale wind direction is known. Based on the findings from the simulations, an updated wind climatology over the glacier might give more insight into when a katabatic glacier wind is present or when synoptic winds dominate. Our simulations showed that this model set-up can be used
for other case studies, and also for other seasons (i.e., winter), to deliver high-resolution wind fields for the analysis of wind-driven snow redistribution patterns (Voor-dendag et al., 2021) or for studies of WRF simulations coupled with energy balance models (Sauter et al., 2020).

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AUTHOR CONTRIBUTIONS
Brigitta Goger: conceptualization; formal analysis; investigation; methodology; project administration; software; validation; visualization; writing – original draft; writing – review and editing. Ivana Stiperski: conceptualization; data curation; formal analysis; methodology; resources; software; writing – original draft; writing – review and editing. Lindsey Nicholson: conceptualization; data curation; formal analysis; methodology; resources; writing – original draft; writing – review and editing. Tobias Sauter: funding acquisition; methodology; project administration; software; supervision; writing – original draft; writing – review and editing.

CONFLICT OF INTEREST
The authors declare no conflict of interest.

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SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section at the end of this article.

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APPENDIX A. TURBULENCE REPRESENTATION IN THE MODEL

The turbulence representation in real-case LESs of the atmosphere faces multiple challenges; for example, realistic inflow boundary conditions and correct turbulence development without unphysical, spurious patterns. A method to avoid unrealistic turbulence development is the introduction of cell perturbations of potential temperature at the inflow boundaries (Muñoz-Esparza et al., 2014). This is a common practice in (semi-)idealized LESs, where sources of turbulence generation (e.g., complex topography, surface heterogeneity) are absent.

In real cases, the cell perturbation method was only used in a few recent studies to our current knowledge (Conolly et al., 2021) and has not been fully implemented yet in the WRF version we are using (WRF 4.1). Therefore, we do not employ the cell perturbation at the LES domain borders. However, we have to ensure that turbulence is generated in a realistic way in our innermost domains. There are multiple reasons why we assume that this is the case in our innermost LES domain (d04):

- We regard the first 3 hr of model simulation as spin-up and exclude them from our analysis of the results to make sure that turbulence develops accordingly.
- Our region of interest is located in highly complex terrain, and at the fine grid spacing of \( \Delta x = 48 \) m the terrain and land use are already well represented. Therefore, we can expect that the thermal structure caused by the lower boundary condition leads to enough perturbations for the generation of a realistic turbulence structure in the LES domains. Aside from small-scale processes, larger-scale, dynamically induced gravity waves also play an important part in turbulence generation at both upper levels and near the surface.
Aside from the two larger domains (d01 and d02), run with an ABL parametrization, the two LES domains d03 ($\Delta x = 240$ m) and d04 ($\Delta x = 48$ m) are run without an ABL parametrization. Since the topography is already well resolved in d03, we can expect that d04 receives realistic inflow fields from d03.

The innermost domain falls into the range where an ABL parametrization is not feasible any more, because it can be expected that the largest turbulent eddies in the model are already resolved (Cuxart, 2015). The simulation set-up can be considered as an LES, while a small portion of the modelled turbulence remains subgrid scale. The split of overall turbulence in resolved and subgrid scale can be calculated from the output from the WRF LES Diagnostics module as described in Umek et al. (2021). We calculate the resolved covariances from the wind components,

$$
\overline{u'^2}_{\text{RES}} = \overline{\tilde{u}\tilde{u}} - U^2, \overline{v'^2}_{\text{RES}} = \overline{\tilde{v}\tilde{v}} - V^2, \overline{w'^2}_{\text{RES}} = \overline{\tilde{w}\tilde{w}} - W^2,
$$

(A1)

and resolved TKE follows after Stull (1988b):

$$
\text{TKE}_{\text{RES}} = \frac{1}{2}(\overline{u'^2}_{\text{RES}} + \overline{v'^2}_{\text{RES}} + \overline{w'^2}_{\text{RES}}).
$$

(A2)

Furthermore, subgrid-scale TKE can be derived from the subgrid-scale variances provided by WRF LES Diagnostics.

We validate the simulated TKE structure with TKE observations from the station hefex-3 (Figure A1). The resolved part of TKE is much larger than the subgrid-scale TKE, suggesting that the term “large-eddy simulation” is valid for our current set-up. However, the model overestimates the resolved TKE on both case-study days, related to the overestimated wind speeds in the model. Otherwise, there is a spike in the modelled TKE on the NW day around noon, related to the breaking gravity wave over the glacier. Although the TKE is overestimated by the model due to the aforementioned reasons, we assume that turbulence develops accordingly in the model after initialization.