Comparison of temperature and wind between ground-based remote sensing observations and numerical weather prediction model profiles in alpine complex topography: the Meiringen campaign

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Abstract. Thermally driven valley winds and near-surface air temperature inversions are common in complex topography and have a significant impact on the local and mesoscale weather situation. They affect both the dynamics of air masses and the concentration of pollutants. Valley winds affect them by favoring horizontal transport and exchange between the boundary layer and the free troposphere, whereas temperature inversion concentrates pollutants in cold stable surface layers. The complex interactions that lead to the observed weather patterns are challenging for Numerical Weather Prediction (NWP) models. To study the performance of the COSMO-1E (COnsortium for Small-scale MOdeling produced by the Kilometre-scale Ensemble Data Assimilation) model analysis, which is called KENDA-1, a measurement campaign took place from October 2021 to August 2022 in the 1.5 km wide Swiss Alpine valley called Haslital. A Microwave Radiometer and a Doppler Wind Lidar were installed at Meiringen, in addition to numerous automatic ground measurement stations recording meteorological surface variables. Near the measurement site, a low-altitude pass, the Brünig Pass, influences the wind dynamics similarly to a tributary. The data collected show frequent nighttime temperature inversions for all the months under study, which persist during the day in the colder months. An extended thermal wind system was also observed during the campaign, except in December and January, allowing an extended analysis of the winds along and across the valley. The comparison between the observations and the KENDA-1 data provides good model performances for monthly temperature and wind medians but frequent and important differences for single profiles, especially in case of particular events such as foehn. Modeled nighttime ground temperature overestimation is common due to missed temperature inversions resulting in a bias up to 8 °C. Concerning the valley wind system, modeled flows are similar to the observations in their extent and strength, but suffer from a too early morning transition time towards up valley winds. The findings of the present study allow to better understand the temperature distributions, the thermally driven wind system in a medium size valley, the interactions with tributary valley flows, as well as the performances and limitations of KENDA-1 in such complex topography.

Keywords. Complex topography, Remote sensing, NWP, Temperature inversion, Valley winds, Foehn

1 Introduction

In mountainous areas, interactions between the terrain and the overlying atmosphere favor horizontal and vertical transports of moisture and pollutants. The complex topography of the Alps consequently increases air masses exchanges along the valleys and between the boundary layer and the free troposphere (De Wekker and Kossmann, 2015; Rotach et al., 2022). Both theoretical studies and experimental campaigns demonstrated that complex topography creates circulations with small and large space and time pattern (Lehner and Rotach, 2018). In valleys, the superposition of the various processes leads to a complex vertical layering in the mountainous boundary layer, which strongly depends on the specific conditions of the surrounding terrain in each studied valley. For Numerical Weather Prediction (NWP) models, simulation of the atmosphere over complex terrain requires not only dense and accurate horizontal and vertical grids to parameterize the mountainous terrain (Sekula et al., 2019) but also good estimates of vegetation cover, soil characteristics, net radiation, and speed of the large-scale flow (Adler et al., 2021). Difficulties of models directly related to complex topography comprise, among others, the representation of ground-based temperature (T) inversions, thermally induced valley winds, and particularly Foehn events.

During calm clear nights, the air T in valleys can fall below the T measured across the surrounding hill tops leading to cold-air pooling and associated T inversions in mountainous regions (Miró et al., 2018; Joly and Richard, 2019). T inversions influence fog formation (Chachere and Pu, 2017), vertical dilution of pollutants (Duine et al., 2017; Diémoz et al., 2019) and the development of the boundary layer during daytime (Schnitzhofer et al., 2009). Such inversions often occur in complex topography (Joly and Richard, 2018) and are temporally more persistent in steep valleys compared to inversions over a plain, whereas wider valleys approach similar inversion characteristics as observed over plains (Colette et al., 2003).

However, the small-scale nature of inversions means that they are often poorly represented even in high-resolution operational NWP models (Vosper et al., 2013). Such stable conditions are controlled by complex small-scale circulations that depend on turbulent fluxes, short-wave and long-wave radiation, advection and subsidence. Therefore, the quality of the predictions is highly dependent on the representation of subscale processes. Deficiencies in the parametrization of the fluxes, especially during stable conditions, are well known (Hauge, 2006) and thus finer grid resolutions should be used for steep terrain (Sfyri et al., 2018). Simulations also underline the high sensitivity to the choice of the vertical grid in the prediction of cold pool formation and suggest that the vertical resolution near the surface is more important than the height of the lowest level (Vosper et al., 2013). However, assimilation of measurements, not only of surface data but also of profiling observations (Crezee et al., 2022), may improve the performance of NWP models for surface T inversions (Martinet et al., 2017).

Thermally driven winds primarily occur under fair-weather conditions (Zardi and Whiteman, 2013) and develop as a result of differential heating of adjacent air masses. The formation of thermally driven winds can partially be explained by the topographic amplification factor concept (Whiteman, 1990) and local subsidence in the valley center induced by up-slope flow (Schmidli and Rotunno, 2010) leading to an increased heating rate of the air masses in the valley than over the plain. The valley–plain T contrast then produces an along-valley pressure gradient that induces strong up-valley winds during the day and shallower down-valley winds during the night. Slope winds are air-mass movements parallel to the slope induced by buoyancy generated by a vertical temperature gradient. Slope winds move upward during the day and downward at night and play an important role in the morning and evening transition of along valley winds. However, slope winds evolve over shorter time scales than valley winds (Serafin et al., 2018).

The transition between up- and down-valley winds is mostly driven by the sunrise and sunset. Although minor changes in topography can lead to a significant change in flow regimes (Lang et al., 2015), some common characteristics are observed among existing studies. In general, the morning transition occurs with a certain delay with respect to sunrise caused by the time required for upslope winds and warm subsidence to erode the nocturnal T inversion. However, wind intensity can be heavily related to tributary valleys (Zängl, 2004) and therefore highly depends on the local topography. In the evening, as soon as the surface radiative balance becomes negative, the cold air forming at the surface moves down the slope and converges in the valley floor, which reverses the flow direction from up-valley to down-valley winds.

Synoptic winds coupled with either forced or pressure-driven wind channeling effects can superpose the above-described thermal mountain winds (Jacques-Coper et al., 2015; Whiteman and Doran, 1993). These large-scale flows do not have a defined diurnal cycle and are generally stronger than the thermal valley winds. Their effect on the valley wind system is highly variable and depends on the orientation of the synoptic flow with respect to the valley axis (Kossmann and Sturman, 2003; Rotach et al., 2015).

The capability of mesoscale NWP models to calculate the above described diurnal valley winds in real valleys has been investigated in a multiple studies (Chow et al., 2006; Langhans et al., 2013; Giovannini et al., 2017; Schmidli et al., 2018; Schmid et al., 2020; Adler et al., 2021; Schmidli and Quimbayo-Duarte, 2023). Globally, a good agreement between modeled and observed valley winds is achieved if the spatial resolution of the models and surface data (e.g., snow cover and soil moisture) are high enough (Rotach et al., 2015). The size of the valley has an impact on the accuracy of the modeled winds (Schmidli et al., 2018). Generally, a closer agreement between the models and measurements was found for higher spatial resolution, which allows a better representation of the topography (e.g. Skamarock, 2004; Skamarock and Klemp, 2008). (Wagner et al., 2014) shows that the grid resolution should be about 10 to 20 times higher than the relevant topographic feature to fully capture the different exchange processes. Hence, higher grid resolution generally improves the performance of numerical simulations, which is even more pronounced if surface and soil model fields are accurately initialized (Langhans et al., 2013; Schmidli and Quimbayo-Duarte, 2023).

Finally, models show poor performance to accurately simulate foehn events, a typical katabatic wind in Switzerland, with a cold bias in the lower profile ($<$1000 m) of the valleys (Jansing et al., 2022; Tian et al., 2022; Saigger and Gohm, 2022) and wind speeds generally overestimated, both above crest height and within the valley.

Although the surface measurement network is relatively well distributed over the Alps, operational T and wind profile measurements by remote sensing (REM) instruments are scarce within Alpine valleys. However, a precise knowledge of the T structure of the atmosphere in complex terrain is essential for NWP models and the use of REM observations is a solution to obtain sufficient space/time resolution of the fast varying meteorological conditions in valleys.

The campaign in the Haslital provides a unique set of observations providing a ten month period of continuous time series covering winter and summer months. A comprehensive measurement program with a MicroWave Radiometer (MWR), a Doppler Wind Lidar (DWL), a ceilometer and a mobile X-band weather radar was established. The selected location, situated in a narrow valley surrounded by mountain ridges of 2000-3000m, complements previous studies where measurements are predominantly collected in rather elongated and wider valleys.

The first objective of this study is to analyze the seasonal and diurnal cycles of T and wind in the vertical range containing the main topographical features (590-3000 m a.s.l.). The analysis is focused on both, seasonality and isolated events with a focus on T inversion and foehn events. In addition, a comprehensive description of along and cross valley winds during a heatwave event is performed, including a detailed analysis of thermal winds using data from three stations and two grid cells of the model along the valley. The second objective is to evaluate the ability of a convective-scale, operational NWP model to capture the observed atmospheric conditions in a highly complex Alpine Valley, such as the Haslital. To this end, we compare analyses of the operational Kilometer-scale ENsemble Data Assimilation system (KENDA-1) with the ground-based measurements and the profiling observations for both monthly averages and peculiar events.

2 Methods and Data

The campaign took place in Unterbach (MEE), a secondary site in the municipality of Meiringen (MER) in the Haslital valley from October 13, 2021 to August 24, 2022, located in complex topography. The DWL and data from the NWP model are available during the whole campaign, whereas the measurements from the MWR are only available from the end of January, ensuring observations during the winter, spring, and summer months (Fig. S1 for a global view of the instrumental setup).

Unless otherwise stated, the following conventions are valid throughout the rest of the document: data are always reported by the instrument or model name and the site; e.g. MWR/MEE correspond to MWR measurements at MEE and KENDA-1/MER to modeled data from KENDA-1 at the cell comprising the MER site, altitude given in meters (m) is equivalent to the altitude above sea level (m a.s.l.), wind speeds are given in km/h and direction in degrees according to north, times are in UTC. Local time corresponds to Central European Time (CET), which is one hour ahead of UTC time (UTC+1). The monthly averages are aggregated according to the median hourly values of the given parameter, and the median wind speed and direction are calculated by averaging the hourly wind vectors. To extend the wind analysis, the data are selected according to the directions of the longitudinal axis of the valley at both sites, allowing a total angle of 30° (± 15° around the valley axis) for along valley wind and a total angle of 60° (± 30° around the perpendicular to the valley axis) for across valley wind. For this analysis, positive wind speeds (red color) correspond to up-valley (westerly) winds for along valley winds (Fig. 1) and to northern wind from the Brünig Pass for across valley winds, and negative wind speeds (blue color) indicate opposite directions.

Finally, all profiles were linearly interpolated at a vertical resolution of 10 m to allow comparison between the observed and modeled data.

2.1 Site

The observational site is located in Haslital, an Alpine valley within the Swiss Alps in the Bernerse Oberland (Fig. 1). This 30 kilometer long valley extends from the Grimsel Pass (2164 m) to the Lake of Brienz (564 m). The up-valley 15 kilometers in the south of the measurement site are oriented in the SE-NW direction and present a middle size valley floor with steep surrounding slopes. The Haslital is then joined by a tributary valley and continues towards NW with a 1.5 km valley floor and a mean valley depth of 1600 m. In Meiringen, it is joined by a narrow, hanging tributary valley. At this point, the valley gradually bends from NW to SW as it reaches Lake Brienz. Five kilometers before the lake, the Brünig Pass (1008 m) is an important topographic feature that connects the Haslital to the Sarneraatal, a 30 km long valley oriented in the NE-SW direction (Fig. 1 presents a detailed map of the Sarneraatal and its connection to the Haslital). This pass interrupts the near constant ridge's height of about 2200 m in the north of the valley longitudinal axis.

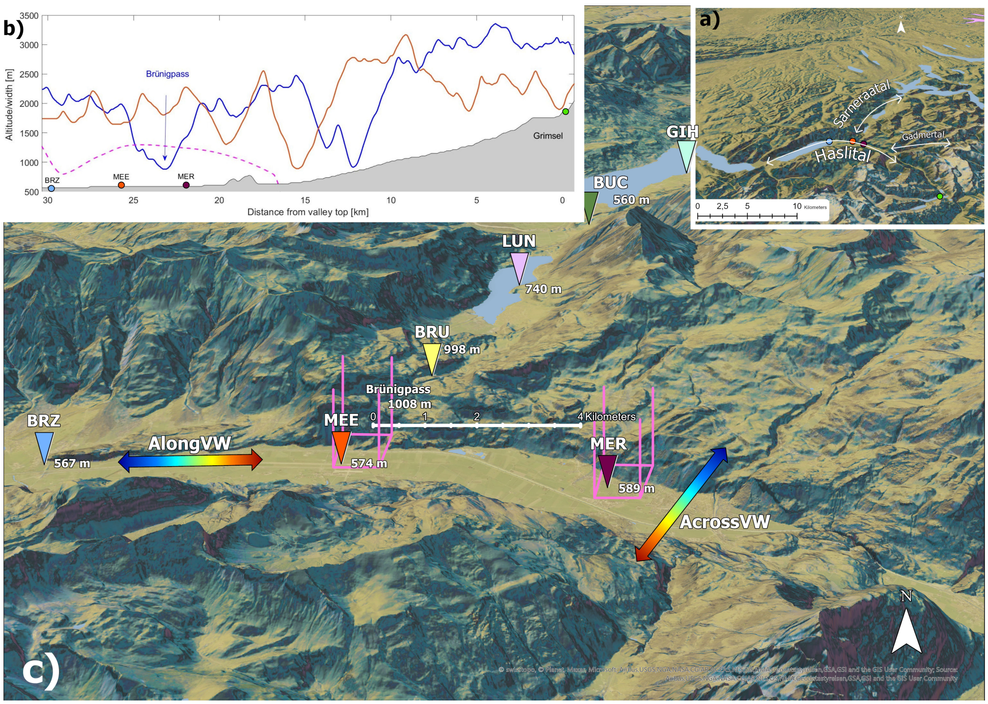

Figure 1. a) Map of the geographical situation in the lower Haslital, b) along valley altitude of the valley floor (shadowed) and of the two crests and c) a detailed view of the campaign sites, the Brünig Pass and of the ground stations in the Sarneraatal. The automatic measurement from the SMN in Meiringen (MER) is represented in purple, the campaign site in Unterbach (MEE) in red and the SMN station in Brienz in blue. The two cells of the model used are in pink. Arrows representing up/down valley winds and north-facing/south-facing slope winds are colored respectively in red/blue. The map was downloaded from Swisstopo (https://map.geo.admin.ch, last access: 12.01.2024)

In this study we use in-situ observations MER (46.732222°N, 8.169247°E, 589 m), a station of the automatic Swiss Measurement Network SwissMetNet (SMN) and REM observations from MEE (46.741344°N, 8.121453°E, 589 m) facing the Brünig Pass. These two locations are separated by 4 km on a height of 589 and 574 m a.s.l respectively. The main differences between these two sites are the valley longitudinal axis angle (ϕ_{MER} = 300°, ϕ_{MEE} = 270°) and the relative position of the surrounding connected valleys. Finally, model data is available for both sites with a 1.1 km grid resolution.

2.2 NWP model COSMO/KENDA-1

The NWP model data used in the study are taken from the operational MeteoSwiss KENDA-1 analyzes, produced by the Kilometre-scale Ensemble Data Assimilation system following Schraff et al. (2016) and the limited-area non-hydrostatic atmospheric model of the Consortium for Small-Scale Modeling Model (COSMO) (Baldauf et al., 2011) in the operational setup of MeteoSwiss. It uses a horizontal grid size of 1.1 km and 81 vertical levels with spacings from 20 m at the surface, 40 m at 1000 m, to 160 m at 3000 m and coarsening further up to the model top at 22 km. The lowest model level is 20 m above ground. The levels are terrain-following and a smooth level vertical (SLEVE) coordinate transformation is applied (Leuenberger et al., 2010). The terrain is filtered by a 2dx filter in order to dampen the high-frequency topography parts to ensure a stable model integration. The differences of KENDA-1 to the setup described in Schraff et al. (2016) include the modeling domain (central Europe covering the Alpine Arc), the grid size of 1.1 km and the observation errors tuned to the MeteoSwiss setup. KENDA-1 uses a 40 members ensemble of 1 hour model forecasts (first guess) and the following observations: SMN ground station measurements (2 m T, humidity and surface pressure), aircraft observations (T and wind from AMDAR and MODE-S), radio soundings (T, humidity and wind) and radar wind profiler (wind speed and direction). In addition, radar-based estimates of surface precipitation are assimilated in every member using the latent heat nudging method (Stephan et al., 2008). The first guess of the model and the observations are combined using the Local Ensemble Transform Kalman Filter (LETKF, Hunt et al., 2007) to obtain the best possible estimate of the current atmospheric state. The KENDA-1 analysis ensemble additionally uses lateral boundary condition perturbations and stochastic physics perturbations to optimize the spread-error relationship. Besides the ensemble analyses, a deterministic analysis member is calculated, which is close to the analysis ensemble mean (Schraff et al., 2016). KENDA-1 data refer to the deterministic analysis member, which are available in hourly time intervals but correspond to instant values.

Data from the two grid cells containing the MER and MEE stations were used. Both cells include part of the valley's north slope, inducing differences of 109 m and 130 m between the real topography and the model's terrain, respectively. The lowest data from the models are available at 705 m for KENDA-1/MER and 739 m at KENDA-1/MEE. The modeled valley floor is globally raised by a hundred meters (Fig. S2), whereas the ridges and the Brünig Pass are lowered with respect to their real altitudes. The altitude difference between the valley floor and the crests is thus reduced of several hundred meters. The Brünig remains a pass in the model terrain, but is only 200 m higher than the valley floor. In the modeled terrain, both the MEE and MER stations are located in the grid cell corresponding to the valley floor (Fig. S3). All in all, it has to be stated that the region under investigation is highly complex and the valleys are only marginally resolved in the NWP model. The Haslital is only less than 2 km wide, and KENDA-1 has a 1.1 km grid spacing. The Sarneraatal is even less resolved and the lakes located in this valley are not present in the model.

It should further be noted that in the region of interest, the observations of the SMN stations MER (2 m T and surface pressure) and Brienz (BRZ, 46.740719°N, 8.060864°E, 567 m) (surface pressure) in the Haslital, as well as Giswil (GIH, 46.849447°N, 8.190225°E, 471 m) (2m T and surface pressure) in the Sarneraatal are actively assimilated in KENDA-1. The assimilation system features a quality control algorithm which ensures that observations too far away from the model counterpart are rejected from the assimilation process. The relevant rejection criterion is based on a first guess check, where the absolute difference between the observation and the model first guess is compared against a threshold. The observation is rejected if the difference is larger than the threshold. The threshold is proportional to the square root of the sum of first guess spread squared and observation error squared. As an example, the observation error of the MER station is 1.18K and the model spread ranges from 0.1K to 2K, resulting in a threshold between 3.5K and 7K, depending on the weather situation. A statistical evaluation revealed that in March 2022 10% of the T observations at 2 m have been rejected, whereas only 1% have been rejected in July 2022. All rejections occurred during the night, suggesting that they occurred mainly in stably stratified atmospheres.

The wind profiles of the wind Lidar and the Microwave Radiometer are not assimilated and the distance between Meiringen and the closest assimilated radio-sounding at Payerne is 94 km, whereas the distances to the three assimilated radar wind profilers situated on the Swiss Plateau are between 75 and 110 km.

2.3 Instrumentation

2.3.1 In-situ meteorological data

The ground measurements in MER are part of the SwissMetNet (SMN) operated by MeteoSwiss and provide every 10 minutes near real-time data of T, humidity, surface pressure, precipitation amount, wind speed (mean and gust) and direction, global radiation, sunshine duration, snow height, and an operational foehn index (Dürr, 2008). Data from additional SMN stations in BRZ in the Haslital, GIH in the Sarneraatal and Frutigen (FRU, 46.599003°N, 7.657542°E, 756 m) are used in this study. BRZ and GIH allow assessing the influence of the winds originating from this tributary valley, while FRU is the nearest station with cloud amount estimation. Furthermore, wind observations from station operated by the Federal Roads Office (FEDRO) at the Brünig Pass (BRU), Lungern (LUN) and Buchholzbrücke (BUC) with similar temporal resolution are used.

2.3.2 Microwave Radiometer

A MWR (TEMPRO-G2 produced by RPG Radiometer Physics Gmbh) is used to obtain T profiles by measuring the emission of microwave radiation from atmospheric trace gases (Rose et al., 2005). It performs a scan every 5 minutes at 11 elevation angles and operates in 7 frequencies reception bands between 51 and 58 GHz. The device has an resolution of 3.5° (half power beam width) at 22 GHz. The data acquired during rainy conditions are discarded. The radiometer is measuring from 50 m above ground up to 2500 m, the first MWR level is then at 625 m. The spatial vertical resolution increases from 50 m at the bottom to 300 m at the top and corresponds to a related T accuracy between 0.25 °C to 1.00 °C, respectively, (Table S1). Löhnert and Maier (2012) compared T profiles based on MWR data and radiosonde data and reported an RMSE between 0.4 and 0.8 K in the lowest 500 m a.g.l., around 1.2 K at 1200 m and around 1.7 K at 4000 m above ground. However, the performance of an MWR is highly related to the retrieval algorithm and the training dataset (Rotach et al., 2015). During the Meiringen campaign, the retrieval developed for Payerne was used (Lohnert and Maier, 2012). This retrieval uses radiosonde data from Payerne to perform the multilinear regression and thus slightly higher uncertainties are expected if applied to observations in MEE. The instrument in MEE had a line of sight of about 10 km, which did not induce further additional uncertainty due to obstacles in the surrounding terrain (Löhnert et al., 2022).

2.3.3 Doppler Wind Lidar

A DWL can be used to infer wind speeds and direction even in complex topography (Wang et al., 2016). During the campaign, a Vaisala Leosphere Windcube 100S DWL was deployed in MEE to measure wind speeds with a vertical resolution of 100 m and a range from 200 m to theoretically 12000 m. For vertical scans, the first DWL level is at 775 m. There are three measurement modes: 120 second zenith scans performed each 10 min to measure vertical wind speed, Range Height Indicator (RHI) scans for two minutes every 10 minutes to measure radial wind speed along and perpendicular to the valley (not used in this study). In the remaining time, the instrument was performed Doppler Beam Switching (DBS) scans providing 7 independent wind profiles every 5 min to estimate the horizontal wind speed. In this analysis, the wind profiles were averaged for each 5 minute interval. Data collected during rain events and/or with confidence level $< 90\%$ are discarded. Moreover, data with wind speeds lower than 2 km/h were discarded for the wind direction analysis. The availability of data during the entire campaign is 80 % at 1000 m a.g.l. and 50% at 2500 m a.g.l.

3 Results

The measurement campaign at Meiringen allows a detailed description of the seasonality based on 6 months T and 10 months wind observations in the Haslital. Profile observations were performed at MEE and surface in-situ observations at MER, whereas the modeled surface and profile data are available at both sites. First we describe the seasonality of the profile observations and the model performances at MEE for the parameters T (sect. 3.1), wind speed and wind direction (Sect. 3.2). Surface observations are then used to study surface based T inversions and the heterogeneity of winds in the Haslital valley. The comparison between KENDA-1 data and observations from MER allows evaluating the model performance at a station, where the surface observations are assimilated into the model. Finally, the KENDA-1 performance during foehn events is described in the last section.

During the campaign, the mean T was 1°C below the 1991-2000 norm in December and January but clearly above the norm (1.5 to 2.5°C) in February, March and from May to August. More than 18 very clear days with at most 2 oktas of cloud cover during daytime were observed at in FRU in January, March, July and August, whereas less than ten very clear days occurred in November, December and May. In addition, three heat waves occurred, the first one lasting 6 days in mid-June, the second lasting 4 days around mid-July and the third one in the beginning of August. Additional important parameters are snow cover and precipitation since the surface albedo and the soil moisture affect the development of cold pools, subsidence, the atmospheric boundary layer and consequently thermal valley winds. Only 60% precipitation was observed compared to the 1991-2000 norm in November, but 120% in December. Snow covered the valley floor from the end of November until mid-December. Heavy liquid precipitation events reduced the snow cover to less than 15 cm by the end of winter. Strong precipitation deficits occurred in January and especially in March (35 and 15 mm). Furthermore, frequent foehn events were observed in March (95 h determined from the MeteoSwiss foehn index (Dürr, 2008)). From May to August, a precipitation deficit of at least 50%

was observed compared to the norm, except for June (96%). The full evolution of T, precipitation and sunshine duration is aggregated in the supplement (Tab. S2 and Fig. S4) and the wind features are fully described in the results section.

3.1 Temperature

3.1.1 Seasonality of temperature profiles at MEE

The evolution of T in MEE from February to July (Fig. 2.a) exhibits as expected a clear diurnal cycle with a vertical extent depending on the season. A layer with higher T develops gradually from sunset to sunrise, persists during the first half of the night, and fades out towards sunrise. The time of the T maximum as well as the persistence and the extent of the warm layer are enhanced during summer months. The maximum temporal T gradient generally follows sunrise and sunset (Fig. S5) and is limited to an altitude of less than 1500 m with values up to +5°C/h in the morning and between -4 and -6.5°C/h in the evening.

A thermal inversion layer is particularly visible from midnight to sunrise (Fig. 2.a) near the ground (590-1000 m) for all months of the study. The frequency of occurrence of these T inversions is highlighted by the positive vertical T gradient. A complete analysis of T inversion will be described in Sect. 3.1.3.

Fig. 2.b shows the differences between the observed MWR/MEE and modeled KENDA-1/MEE T profiles. In general, KENDA-1/MEE underestimates T at low altitude ($<$ 1500 m). In February, this underestimation lasts almost the whole day up to 2500 m, but is more enhanced ($<$ -1 °C) below 1500 m. March and April exhibit the same T underestimation below 1500 m, while a small T overestimation ($<$ 1 °C) is observed in March above the ridges in the morning. In May and June, underestimations are constrained to nighttime. In July a T underestimation at lower altitude ($<$ 1000 m) and a persistent T underestimation of up to -2 °C at the ridge level is observed. This was already partly present in May and June but the underestimation of 1-2 °C of KENDA-1/MEE is slightly larger compared to the MWR uncertainties ranging from 0.25 to 1°C as a function of altitude (see Sect. 2.3.2). However, the cold bias between the MWR and the radio sounding could suggest a larger error of KENDA-1.

3.1.2 Surface temperature comparisons

To better estimate the reliability of both the REM observations and the model, the lowest levels of MWR/MEE, KENDA-1/MEE and KENDA-1/MER are compared to the SMN/MER measurements used as a reference due to its low uncertainty ($\approx$ 0.2 °C). Differences in T between MWR/MEE and SMN/MER (Fig. 3. a) are normally distributed with a mean and median close to zero (-0.07°C) and RMSE equal to 1.45°C. Extreme differences (3σ) are larger than $\pm$ 4.35 °C.

The distribution of ground T differences between KENDA-1/MEE and SMN/MER (Fig. 3.b) is wider compared to the difference found for MWR/MEE (RMSE = 2.23 °C) and shows a positive skew (median = -0.27 °C and mean = +0.03 °C). Extreme values are significantly more frequent than for the MWR/MEE measurements, especially in the positive part of the distribution. KENDA-1/MEE T underestimations occur more often but with lower absolute differences than the overestimations, and the differences with the SMN/MER T reference can reach up to 9 °C. A similar distribution is observed for KENDA-1/MER (Fig. 3.c) with the same occurrence of extreme T differences (217 h). Differences below 2 °C represent 71.1 % at MER and 66.0 % at MEE which explains the slightly smaller RMSE for the cell over the SMN/MER station.

To check whether the differences in altitude between the stations and the first KENDA-1 level could explain the differences in T with SMN/MER, a standard correction of T with a mean environmental lapse rate (ELR) (-6.5 °C/km (Lute and Abatzoglou, 2021)) close to the mean measured lapse rate of MWR/MEE (-4.59 °C / km between 590 and 740 m) was applied to the modeled profiles. Considering the remaining T differences after the correction (grey in Fig 3.b and 3.c), we conclude that the horizontal and vertical distances between the SMN/MER station and the first level of KENDA-1/MEE are not the main causes of discrepancies in ground T estimation.

The median diurnal cycle of T differences between KENDA-1/MER and SMN/MER (Fig. 4) shows that KENDA-1 overestimates the T during nighttime (+1.5°C) in both cells and underestimates T during the day (-2°C in MEE and -1.5°C in MER). The interquartile range (0 to 3.5 °C) and the whiskers (-4 to 8 °C) of the differences are larger during the second part of the night for KENDA-1, when surface T inversions are more frequent. Thus, the presence of T inversions strongly influences the amplitude of the differences (see details in the next Sect. 3.1.3). One third of the daily bias can be explained by the altitude difference between the station and the KENDA-1 first level, since the median T correction during the day is around 0.65 °C. The T bias distributions of KENDA-1/MER and KENDA-1/MEE are similar during the cycle. However, the modeled daytime T over MER shows smaller differences to SMN/MER than over MEE, which can be explained by the reduced altitude bias or the reinforced assimilation. MWR/MEE shows no T bias from 21:00 to 6:00 and a negative T bias (> -1°C) from 6:00 to 15:00, followed by a slight overestimation from 15:00 to 21:00 (< + 0.5 °C). The MWR/MEE T differences have smaller whiskers and interquartile ranges during the second part of the night compared to KENDA-1/MEE, but they are similar during daytime. Overall, T observed at the lowest level of the MWR/MEE is closer to the T surface observation SMN/MER while modeled KENDA-1 T values shows higher deviations from the surface observations.

3.1.3 Surface temperature inversion

A comparison between the T inversions detected by two ground observations at different altitudes (MER and BRU), the REM MWR/MEE as well as the modeled KENDA-1/MEE allows a better estimation of the frequency of occurrence of cold pools, the sensitivity of REM observations, and the limitations of the model. The availability of the ground stations requires an altitude difference of $\simeq 400 m while T inversions could extend only up to 40 - 50 m a.g.l.. The frequency and amplitude of the ground - based T inversions are then underestimated within this analysis. An offset between the T inversions observed on the ground compare$ $valley wind from the Sarneraatal (3.3), which, however, also affects MWR/MEE and SMN/MER.$

Fig. 5.a shows the frequency of occurrence of negative T differences between MER at 576 m and BRU at 1000 m (horizontal distance = 3.7 km). It indicates that near-ground T inversions are common during the night for all months. The frequency of T inversions is 60% in December and January (all day), 40% and 30% during spring and summer nights, respectively. Daytime near ground inversions are common between November and February (20-60%), very high in December when the Haslital stays in the shade most of the time, but rare from March onwards. The foehn influence in March occurred mostly during daytime (8.1 % of daytime and 4.8 % of nighttime) and therefore did not directly influence the T inversion frequency. The observed T inversion amplitude follows a seasonal cycle with stronger inversions during winter months reaching up to 4 °C (Fig. 5.b). In summer, this

amplitude is reduced to about 2°C and constrained to nighttime. The erosion speed of the T inversion is independent of the month. However, the delay of the erosion onset to sunrise is smaller in summer (about 2h) than in winter (about 4h).

The same analysis between two similar elevations is performed on MWR/MEE and KENDA-1/MEE T profiles. MWR/MEE shows higher T inversion frequencies than both ground stations and KENDA-1/MEE, especially for June and July. MWR/MEE also presents a larger amplitude of the T inversion than the ground observations and KENDA-1/MEE with a maximum difference of +2°C and +4°C, respectively. As presented later on (3.3), the warmer MWR/MEE measurements in the free atmosphere (at 1000 m) than at BRU explains the higher frequencies and amplitudes of T inversions measured by MWR/MEE. From November to January, KENDA-1/MEE detects most of the near-ground T inversions, which last all day in winter, but their amplitude is always underestimated by 1-2°C (Fig. 5.b). From February to August, the presence of T inversions at the end of the night and in the first hours after sunrise is often underestimated by KENDA-1/MEE, which can affect the time of onset of the up-valley winds (Sect. 3.2.2). The underestimation of the T inversions by KENDA-1/MEE can be caused by the overestimation of T at ground level (Fig. 4) and the slight underestimation of T at higher altitudes between 850-1200 m (Fig. 2). Detailed examples of T profiles during a day with missed T inversion by KENDA-1/MEE (Fig. S6) show these opposite T bias with SMN/MER and MWR/MEE observations at several altitudes.

The analysis of the assimilation process for nights with strong ground KENDA-1/MER T overestimations shows that the model suffers from a systematic deficiency. During these nights, differences between the model's first guess and observations are mainly around 5 °C and can reach 10 °C in extreme cases (results not shown), so that observations are rejected due to differences exceeding the predefined threshold based on the ensemble's first guess, its spread, and the observation error. During these periods, the SMN/MER T is, therefore, not assimilated by the model analysis. Even if the observations are assimilated for some of the KENDA-1 time steps, assimilation has a very limited effect and allows only minor corrections towards the observations (< 1 °C) during some nights in both MEE and MER. It has to be noted that the KENDA-1 T overestimation during nighttime is similar at MEE and MER (Fig. 4).

3.2 Wind

During the campaign, the wind profiles were measured at MEE by the DWL, whereas ground-based 10 m wind is continuously measured at SMN/MER and at five other SMN and FEDRO ground stations (Fig. 1). First, the seasonality of the average measured wind profiles is described, followed by a more detailed analysis of the along and across valley components at MEE. The performance of KENDA-1/MEE is analyzed in each section. A comparison between the results for MEE and for other ground stations in the valley gives an insight in the complexity of the wind system caused by the peculiarities of the valley's topography.

3.2.1 Seasonality of wind profiles at MEE

Fig. 6.a presents the monthly median wind directions from the DWL/MEE observations for all weather conditions and correspond therefore to the overall effect of thermal wind generated within the valley combined with the influence of synoptic winds by topography or pressure channeling or downward momentum transport ((Whiteman, 1990)). The thermally induced valley winds

are characterised by a shift in wind direction after sunrise and sunset. In December and January, no clear presence of regular direction changes is observed at any altitudes. A clear shift in wind direction with a clear on-set of up-valley winds at sunrise and a gradually onset of down-valley winds at sunset is observed in February below 1200 m. Weaker diurnal cycles are observed in November and March from mid-day to around sunset. These shallow diurnal cycles can be explained by full snow coverage in November and by the channeled easterly winds due to frequent foehn events in March. A predominance of easterly winds is measured below 2000 m in November and below 1200 m in January, whereas a predominance of NW winds below 1500 m and of W winds at higher altitude is observed in December and February. The formation of a thermally induced wind is then clearly visible from April to August and will be further discussed in Sect. (3.2.2). From 10:00 to mid-afternoon, the direction at low altitudes (800-1000 m) is mainly from W, whereas flows from W-NW are measured in the upper profile up to the ridge height (see further explanation in Sect. 3.3). Above the ridge height, no diurnal cycle is observed but synoptic winds from NW to SW direction dominate in all months, with higher variability in January. In March, strong influence of foehn events can be observed. From April to August, NE winds from Sarneraatal (Sect. 3.2.3) are also observed from the ground to 1000 m from late midday to several hours after sunset. Fig. S7 presents the same monthly median of wind direction but restricted to fair-weather days with less than 5 oktas of cloud cover during daytime at the nearby FRU station. This selection of fair-weather days drastically restricts the number of days considered for some months. The general features are similar for March to August and the main difference is the absence of a clear feature in wind direction change in November and February.

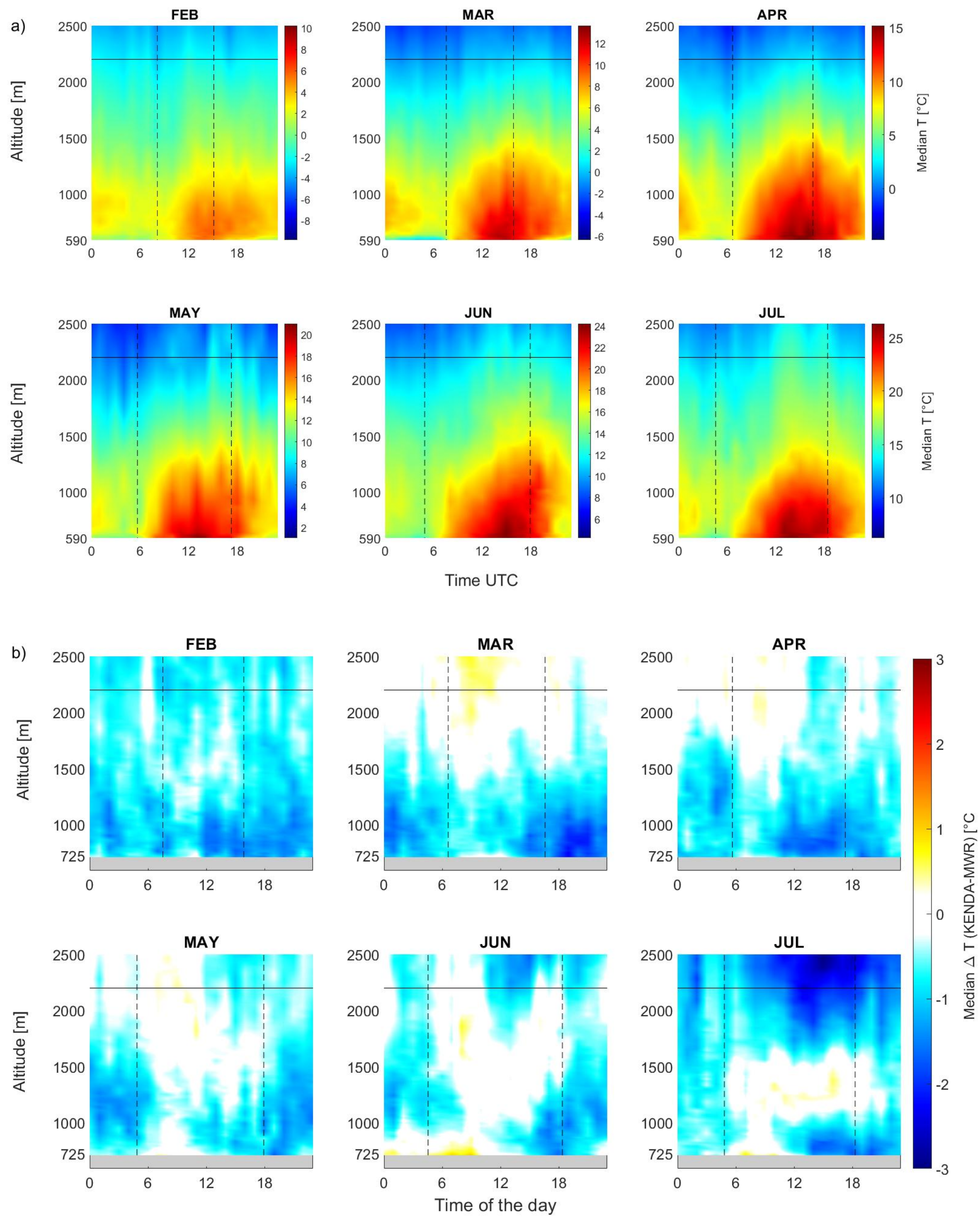

Figure 2. a) Monthly diurnal cycle of MWR/MEE T from February to July 2022. Monthly scales with a range of 20 °C but with minimum T based on the MWR/MEE profiles are used. b) Diurnal cycle of the median T profiles difference [°C] between KENDA-1/MEE and MWR/MEE for each month. The dashed vertical lines correspond to sunrise and sunset and the horizontal line to mean ridges' height.

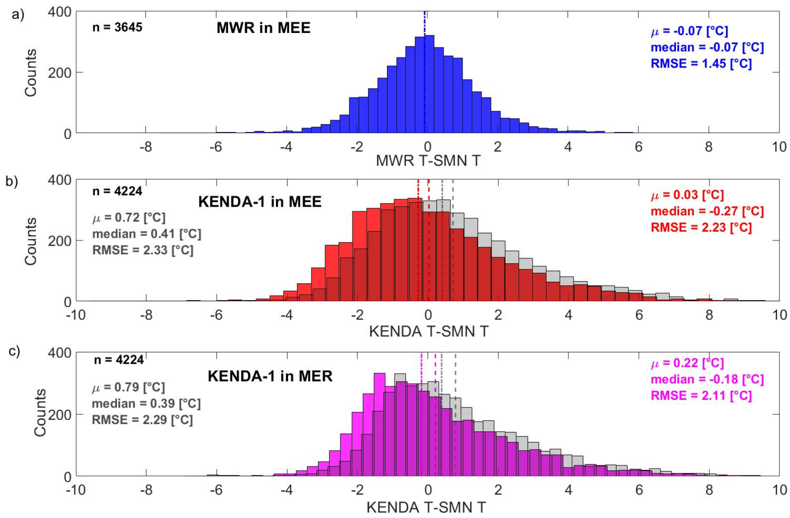

Figure 3. Distribution of the hourly T differences at the lowest level for a) MWR/MEE-SMN/MER b) KENDA-1/MEE-SMN/MER, c) KENDA/MER-SMN/MER. The lowest level corresponds to 576 m for SMN/MER, 625 m for MWR/MEE and 705 m for KENDA-1/MER and 739 m for KENDA-1/MER. The gray distributions indicate ground T differences after ELR corrections are applied. The dotted and dashed lines correspond to the median and the mean, respectively.

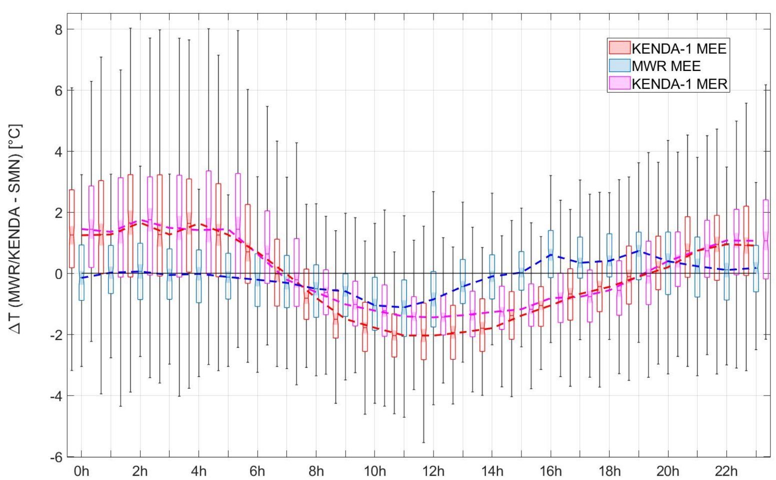

Figure 4. Box plots and whiskers of hourly ground T differences between the SMN/MER and the MWR/MEE (blue), the SMN/MER and KENDA-1/MEE (red), the SMN/MER and KENDA-1/MER (pink) as a function of daytime. The lowest level corresponds to 576 m for SMN/MER, 625 m for MWR/MEE and 705 m for KENDA-1/MER and 739 m for KENDA-1/MER. The dashed lines represent the median of the distributions. Only data present in all time series are used.

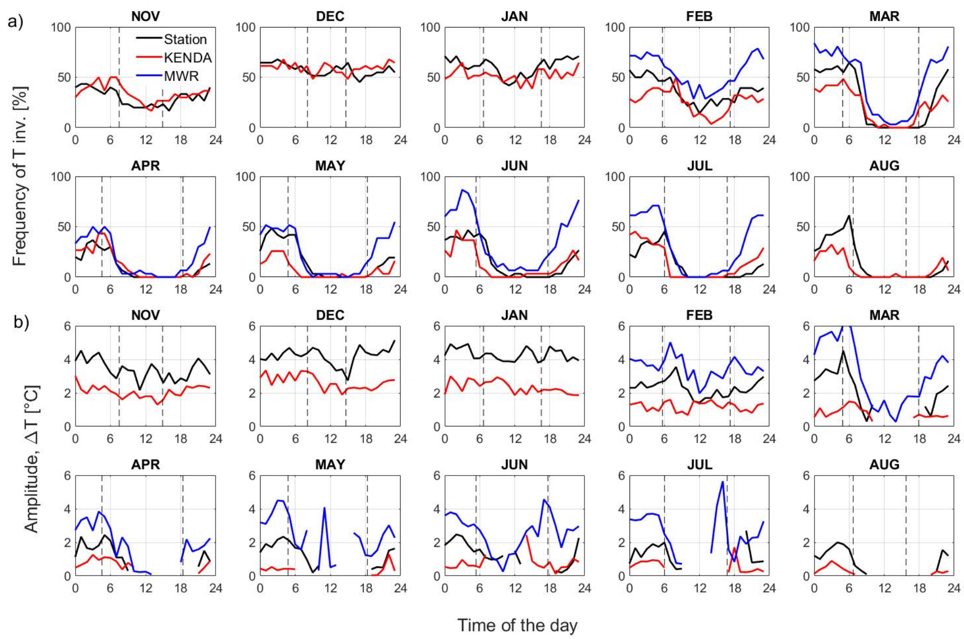

Figure 5. a) Diurnal cycle of the hourly T inversion frequency between T at SMN/MER (576 m) and FEDRO/BRU (998 m) ground stations (black), at the lowest level (625 and 739 m, respectively) and 1000 m of MWR/MEE (blue) and KENDA-1/MEE (red) profiles. b) Mean ΔT for the time where an inversion is detected. Sunrise and sunset are represented by dotted lines.

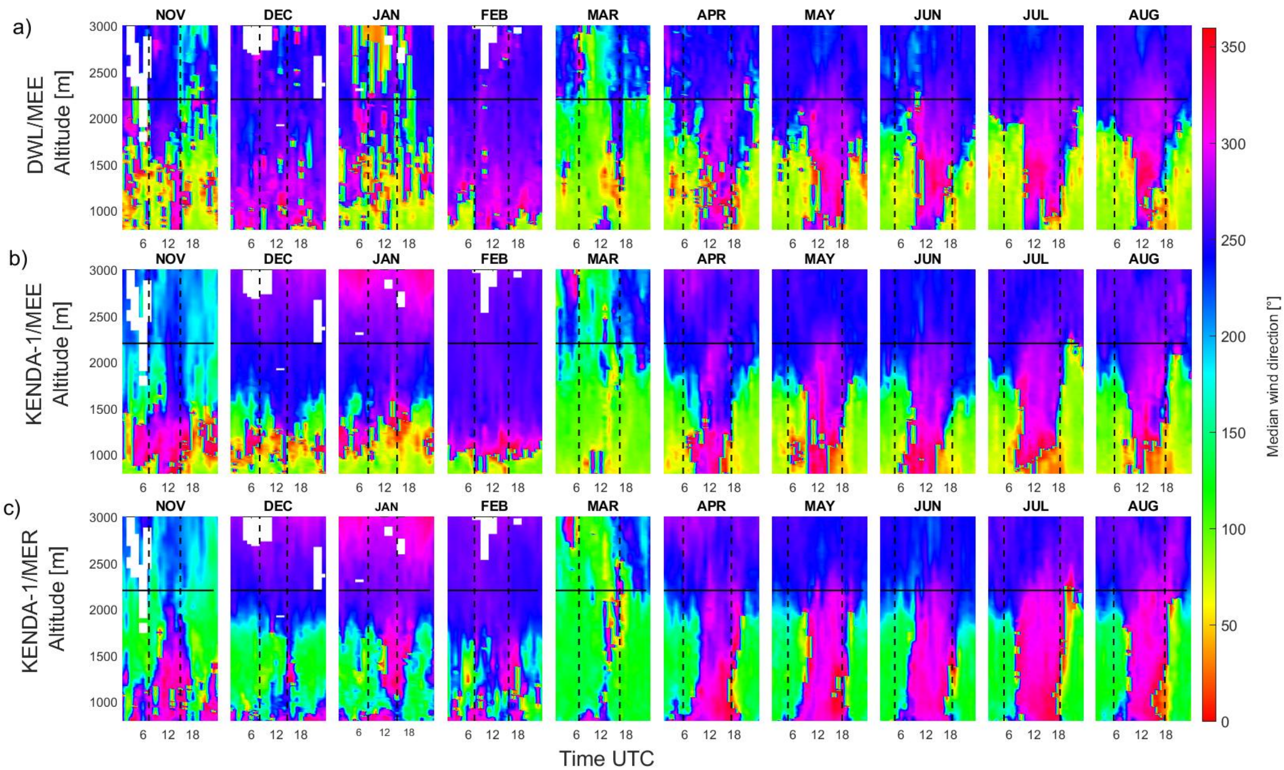

Figure 6. Monthly median wind direction [°] for a) DWL/MEE, b) KENDA-1/MEE and c) KENDA-1/MER (01.11.2021-23.08.2022). The vertical dashed lines correspond to sunrise and sunset and the horizontal line to the mean ridge height.

The KENDA-1/MEE wind profiles (Fig. 6.b) are generally very similar to the DWL/MEE observations. The good KENDA-1/MEE performances comprise first the influence of the foehn up to 3000 m and the valley wind pattern from April to August. The synoptic wind flows above the ridge height captured by model inputs and by assimilated measurements (e.g. RS, MWR and DWL profiles) from the Swiss plateau are also very well modeled with largest differences in November and January. A diurnal valley wind pattern is observed by DWL/MEE in February but is not modeled by KENDA-1/MEE, whereas it is modeled in November but only weakly observed. The presence of a shallow valley wind cycle in March is less visible in KENDA-1/MEE data. Apart from inaccuracies related to the valley wind transitions (see 3.2.2), the model and the measurements differ in the presence of frequent N flows from the Brünig Pass between the ground and 1200 m with increasing frequency toward sunset in KENDA-1/MEE, while N flows are found at higher altitude (1300-1700 m) in DWL/MEE. This characteristic is caused by the lower altitude difference between the topography (400 m) and the model terrain (200 m) and a smaller horizontal distance due to the 1.1 km cells (2.2). Finally, during winter months, KENDA-1/MEE exhibits continuous down-valley (E) winds between the ground and 1000 m that are absent in December. The discrepancy between KENDA-1/MEE and DWL/MEE is much lower for all months from November to February if only fair-weather days are considered (Fig. S7), leading to the expected conclusion that cloudy and precipitation days are less easily modeled.

3.2.2 Along valley winds

Fig. 7.a shows the seasonal and diurnal cycles of the wind speed along the valley at SMN/MER. The occurrence of thermally driven valley winds is confirmed by the diurnal cycle in November and from February to August. A 3-4 hour delay between sunrise and the onset of up-valley winds (> 10 km/h) is observed. February shows some early up-valley wind but their origin is rather linked to synoptic flow influence. The transition to down-valley winds occurs one hour before sunset in March and June and around sunset otherwise. The maximum of the monthly median speeds of the up-valley wind are between 15-20 km/h. Down-valley winds are weaker with a maximum of the monthly median speed of 2-7 km/h reached within the 2 to 3 hours after sunset. These results agree well with 10-year climatology (Fig. S8), which shows a clear wind speed maximum in July and an onset of dow-valley wind 1-2 hours after sunset in spring.

Similar seasonal and diurnal cycles of the valley wind are measured by the DWL/MEE on the first level at 775 m (191 m a.g.l. (Fig. 7.b). The onset of the up-valley winds occurs with the same delay to sunrise (4 h) during the summer months but their speed is of reduced maximum amplitude (10-15 km/h) than at SMN/MER. At 775 m, the up-valley wind intensity is also less regular than at ground with maximum speed around noon for May to August. The strongest down-valley winds are also measured in the first part of the night, with higher wind speeds (5-10 km/h) compared to the ground at SMN/MER where wind is slowed down by friction and T inversions that impede vertical transport. Furthermore, during August, DWL/MEE exhibits down-valley winds occurring two hours before sunset, whereas they are observed just after sunset at SMN/MER (Fig. 7.a), a difference probably linked to the flows from the Brünig Pass (Sect. 3.3).

In general, the modeled valley wind evolution of KENDA-1/MEE (Fig. 7.d) is consistent with the DWL/MEE measurements. The main differences can be seen in slightly higher up-valley wind speed, an underestimation of the down-valley wind speed and an earlier onset of up-valley winds. A comparison of the first level of KENDA-1/MER and SMN/MER (Fig. 7 b and a)

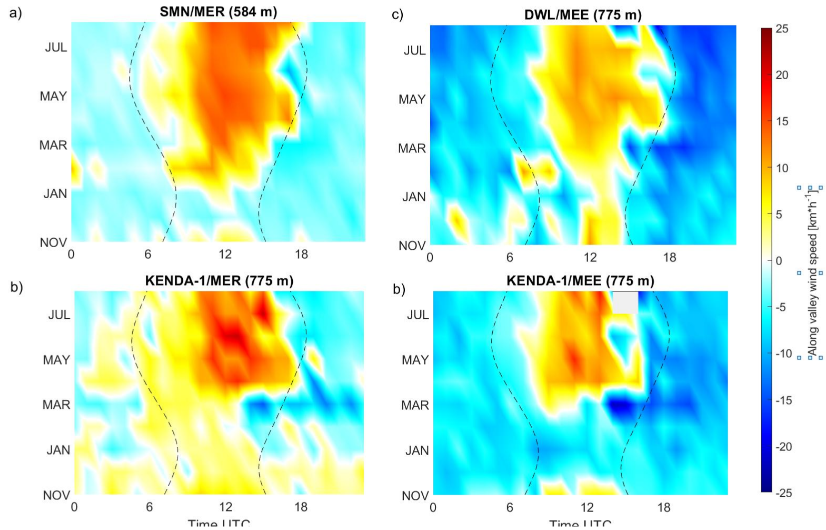

Figure 7. Monthly evolution of along-valley wind speeds [km/h] a) observed at the SMN/MER, b) observed at DWL/MEE, c) modeled at KENDA-1/MER and d) modeled at KENDA-1/MEE. Sunrise and sunset are represented with dashed lines.

indicates an underestimation of the wind speed along the valley by KENDA-1 / MER, leading to the absence of a diurnal cycle in November and December. Even in summer months, the along valley wind diurnal cycle is less pronounced in KENDA-1/MER due to the presence of weak up-valley wind in the second part of the night. The modeled data at MER and MEE also show distinct differences, a stronger presence of up-valley wind in MER during the whole campaign leading to stronger maximum up-valley, weaker down-valley wind speeds and the presence of weak up-valley wind during the entire days in winter.

The monthly diurnal cycle of DWL/MEE wind profiles (Fig. 8.a) allows a better visualization of the vertical extent of thermal valley winds. First, the height of thermally induced wind increases with increasing solar radiation, reaching 1000-1200 m in February, 1800 m in May and up to 2000 m in July and August. Second, the onset of an up-valley wind occurs simultaneously over the entire profile 3-4 hours after sunrise, whereas the onset of down-valley winds is not simultaneous throughout the profile. Near the ground, the onset is anticipated compared to higher altitudes so that up-valley winds can persist until 1-3 h after sunset above 1500 m. Third, the speed of down-valley wind decreases with altitude and with time after sunset. Finally, the daytime wind direction between 1000 m and 1500 m does not stay constant even during the summer months. This might be related to the interaction between synoptic flows and thermally driven flows, or to the influence of flows from the Sarneraatal. In spring and summer, the up-valley winds are stronger and more uniform at 1500 m than at 1000 m and persist longer in the afternoon probably under the influence of the synoptic winds.

The same representation for KENDA-1/MEE (Fig. 8.b) shows that the vertical extent of the modeled valley wind is comparable to the observation with differences of up to $\pm$ 250 m depending on the specific month. The main differences between KENDA-1 / MEE and DWL / MEE are an underestimate of the down-valley wind speed from ground to 1600 m, mostly in summer but also in January and February, and the too early onset of up-valley winds 1-2 h after sunrise between the ground and 1200 m. Finally, in winter, KENDA-1/MEE overestimates the influence of the synoptic winds leading to the presence of homogeneous up-valley winds down to 1000 m and models continuous down-valley winds underneath. The foehn influence in March is well modeled up to 2200 m after sunset, but KENDA-1/MEE extends its impact to 2750 m before sunrise.

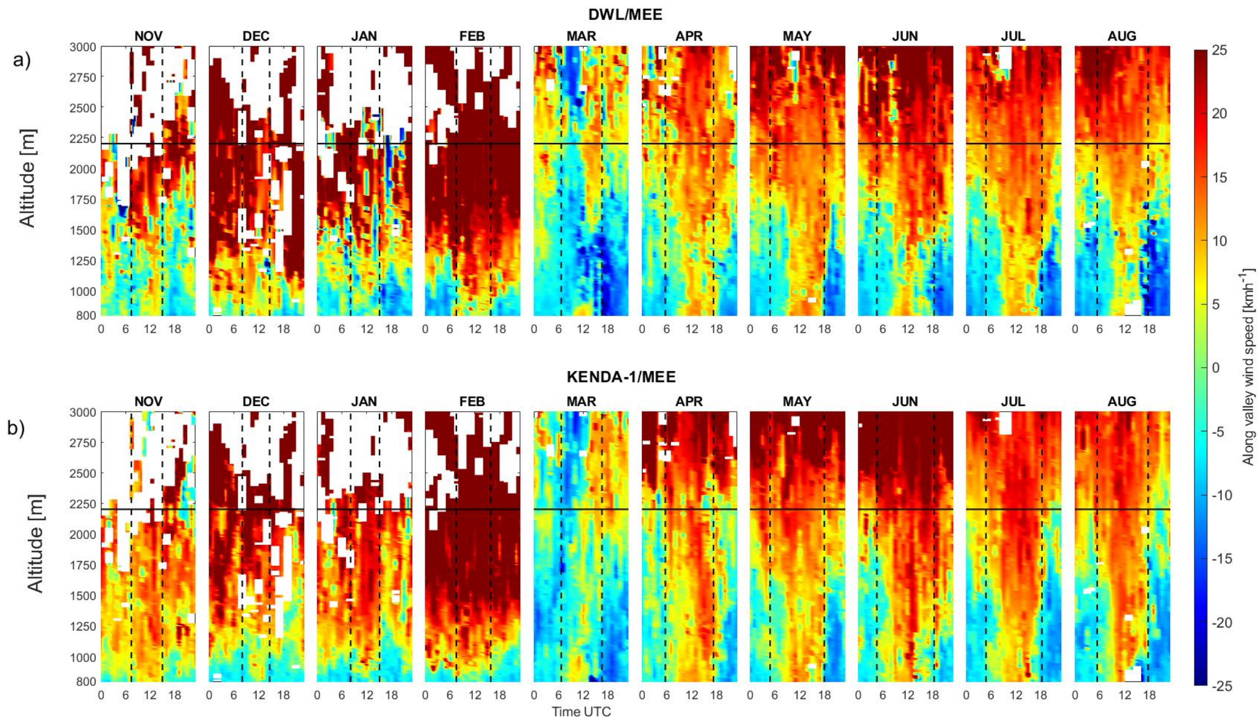

Figure 8. Monthly diurnal cycle of the along-valley wind component [km/h] as a function of altitude for a) the DWL/MEE observation and b) the KENDA-1/MEE data. Sunrise and sunset at ground level are given by dotted lines.

3.2.3 Cross valley winds

The cross-valley winds in MEE can originate from thermally induced slope winds in Haslital or from valley winds from the Sarneraatal passing over the Brünig Pass. Fig. 9.a shows the monthly diurnal cycle of the cross-valley wind measured by the DWL/MEE. During winter, the data are scarce and no particular pattern is visible except the presence of N winds from Brünig Pass at altitudes between 800 m to 1500 m in November, January and February. These N winds are strongest in January when 18 clear sky days were observed and nonexistent in December when only 3 clear sky days occurred. During all other months, strong cross-valley winds originating from Brünig Pass start between midday and sunset and stop around midnight. They are generally first measured near the ground and reach 1200-1500 m after sunset, where they reach wind speeds of 20-25 km/h and can extend up to 2500 m with lower speed. Intense downslope winds from the north-facing slope (> 25 km/h, in blue) are also observed between 1400 and 2000 m during some hours around sunset with a much lower intensity in May. This suggests a circular motion with North updraft winds (median vertical velocity of 1 km/h) that cross the valley at a low altitude, rise against the north facing slope and come back at higher altitude with a South downdraft component (median vertical velocity of -2 km/h). Fig. S10 shows radial winds perpendicular to the valley direction that clearly illustrate this circulation pattern observed in the presence of both up and down-valley winds around sunset.

KENDA-1/MEE also shows cross-valley wind patterns (Fig. 9.b) with strong winds from the Brünig Pass from March to August. These N winds also develop progressively from ground to 1400 m and stop around midnight. They are modeled earlier than measured, at the time (10:00) of the onset of up-valley winds in the Sarneraatal. Winds from the north facing slopes between 1400 and 2000 m are not present in KENDA-1/MEE despite being systematically observed with rather high intensities from April to August. This might be related to the model topography, where the height difference between the Brünig pass and the valley floor is underestimated and the lakes of the Sarneraatal are absent, leading to higher modeled T in the Sarneraatal and stronger winds from the pass.

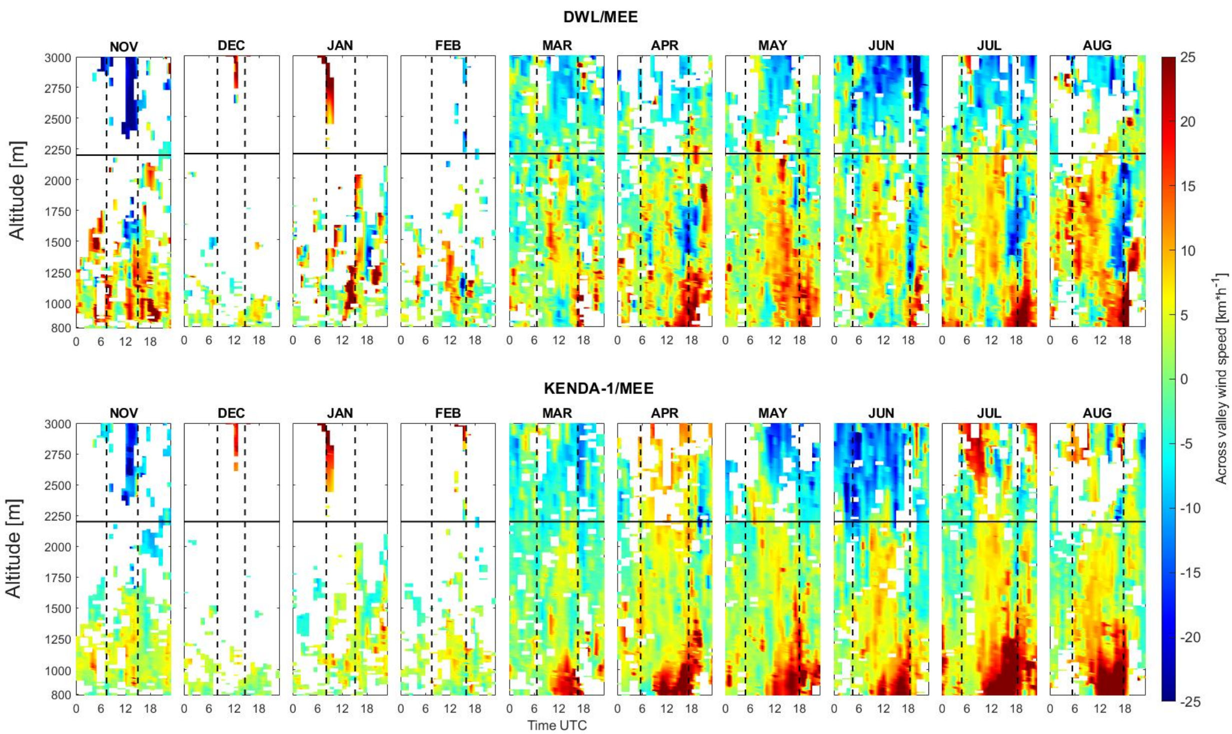

Figure 9. Evolution of the diurnal cycle of the cross-valley wind component [km/h] as a function of altitude for a) the DWL/MEE measurement and b) the KENDA-1/MEE. Winds coming from the south-facing slopes take a positive value (red), for the north-facing slope wind speeds values are negative (blue). Sunrise and sunset at ground level are given by dotted lines.

3.3 Heterogeneity of wind patterns in the Haslital

The different locations of the ground observations in Haslital allow a comparison of modeled data with observations at two different sites with different valley directions and different topographic features. Furthermore, a detailed analysis of the effect of the Brünig Pass during clear summer days is performed with the additional ground observations for wind in Haslital and in Sarneraatal. The modeled data provide some further insight in the difference of the thermal wind system from the lake of Brienz to the MER station.

A closer look at the SMN/MER and DWL/MEE wind speeds during a series of clear warm days in July with low cloud coverage (Fig. 11) shows some particularities relative to the previous analysis of along valley wind on the basis of a monthly median values. In SMN/MER (Fig. 11.a), a clear diurnal pattern of thermally induced winds is measured. The onset of up-valley winds occurs at 10:00 and the wind speed strengthens during the day (approximately +4 km/h per hour) to reach a maximum of 25-30 km/h at 15:00-16:00. The onset of down-valley winds occurs at 19:00. During night, down-valley winds are constant in direction and drop to 0-5 km/h. It has to be mentioned that the direction of up-valley winds at MER gradually shifts from the longitudinal axis of the Haslital towards an enhanced northern component on the 10 and 11 July during the afternoon.

In the lowest level of the DWL/MEE observations (190 m a.g.l.), up-valley wind is only measured on 10 July at 13.00-14:00 (Fig. 11.a, color bar). The wind direction switches thereafter to N and the wind speed increase gradually to reach 40 km/h at 20:00. The wind then weakens until midnight and changes direction afterward with a down-valley wind direction that persists occasionally (e.g. on 12 July) during the morning. Along valley wind patterns following the valley longitudinal axis (W/E) are only observed between 1300 m and 2000 m (not shown), which is higher than the Brünig Pass altitude. These along valley winds show a standard diurnal cycle with up-valley wind measured from 09:00-10:00 to 16:00-17:00 indicating wind speeds between 15 and 20 km/h.

In SMN/BRZ, the wind pattern varies during the three selected days (Fig. 11.a). July 10 and 12, up-valley wind begins at 8:00 and last until 14:00 with low wind speeds of 5-10 km/h. At 14:00, the wind direction switches towards down-valley winds (17-19 km/h), which last until 20:00. A small direction change towards the WSW occurs during the night. July 11, there is no up-valley wind phase with only down-valley wind (NE/N) during the entire day. Wind speeds are lower in the morning and strengthen to 20 km/h in the afternoon before weakening at 21:00.

The strong influence of the thermal winds from the Sarneraatal over the Brünig Pass during hot summer days is highlighted by this wind analysis at the three stations. An analysis of ground measurements from the BRZ, BRU, LUN, BUC and GIH (Fig. S9) automatic stations shows that flows measured at the Brünig Pass switch toward the Haslital (SSW) 2 to 3 hours earlier (5:00-6:00) compared the onset of up-valley wind at other stations in the Sarneraatal (08:00-09:00) and last much longer after sunset, up to 21:00-22:00. A further analysis of the monthly air pressure reduced at the sea level (QFF) at GIH and MER (10.a) shows higher QFF at GIH than at MER from March to August with a clear diurnal cycle. The QFF difference is maximal at noon, decreases slowly and becomes negative between the late evening and late morning, depending on the season. Air masses are consequently colder in the Sarneraatal than in the Haslital, which explains their fall from the Brünig Pass down to the Haslital floor. 10.b shows the difference between the potential temperature (x, y, θ) observed by MWR/MEE at the BRU altitude (1000

m) and at the automatic station in BRU. (x, y, θ) at BRU and MWR/MEE are computed from pressure data of GIH and MER, respectively. (x, y, θ) at BRU is 2-6°C colder than at the same height above MWR/MEE for all months analyzed in this study. The diurnal cycle of T shows the opposite behavior compared to QFF, which can be explained by a faster warming of air masses near the ground at BRU compared to 500 m above the ground in the free atmosphere over MEE. Finally, this observed difference in air temperature can be explained by the valley volume effect. The larger volume of Sarneraatal ($\approx 304\text{km}^3$) compared to Haslital ($\approx 177\text{km}^3$) needs more energy to warm up and results in a colder T.

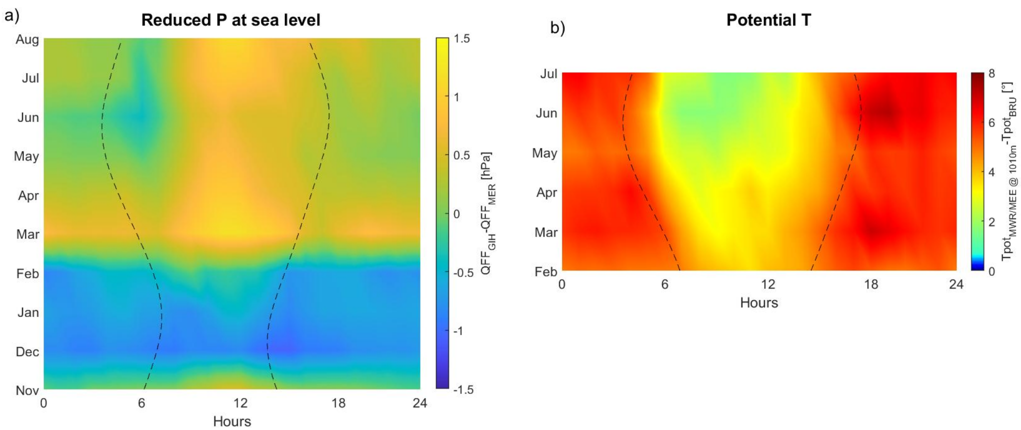

Figure 10. a) Seasonal and diurnal cycles of the difference in pressure reduced at sea level between SMN/GIH and SMN/MER and b) Seasonal and diurnal cycles of the difference in potential temperature between MWR/MEE at 1000 m and BRU. Sunrise and sunset time are given by dotted lines.

The occurrence of wind from the Brünig Pass is driven by the strength of the thermal wind in both the Haslital and the Sarneraatal. It can explain the N wind observed in MEE during the afternoon, the early evening and even sometimes the morning (e.g., on July 11). It also strongly influence the diurnal cycle at BRZ leading to the onset of down-valley winds in the early afternoon or even by suppressing up-valley winds (July 11). Finally, their influence at MER is weak with only a slight shift of the wind direction towards N in the late afternoon. During these summer days, a standard thermal wind diurnal cycle is observed in MER and in MEE at altitudes higher than the Brünig Pass (not shown).

Concerning the modeled data (Fig. 11.b), the influence of the Sarneraatal thermal winds is well captured by KENDA-1 / MEE, so the differences between MER and MEE are important. At MER, the wind speed and direction follow a clear thermally driven valley wind diurnal cycle whereas a relatively stable wind direction from NE during nighttime and N-NE during daytime is modeled at MEE. Wind speeds for MEE are always equal to or higher than those of MER and show a weaker diurnal cycle.

The major differences compared to the observations are an overestimated influence of the valley winds from the Sarneraatal leading to the absence of down-valley winds in the model data at MEE during the night and the morning, as well as a shift in wind direction toward N at MER. The differences in the diurnal cycle of the wind speed at MER and MEE are well modeled by KENDA-1, but the wind speed is overestimated at both sites with differences up to +30 km/h.

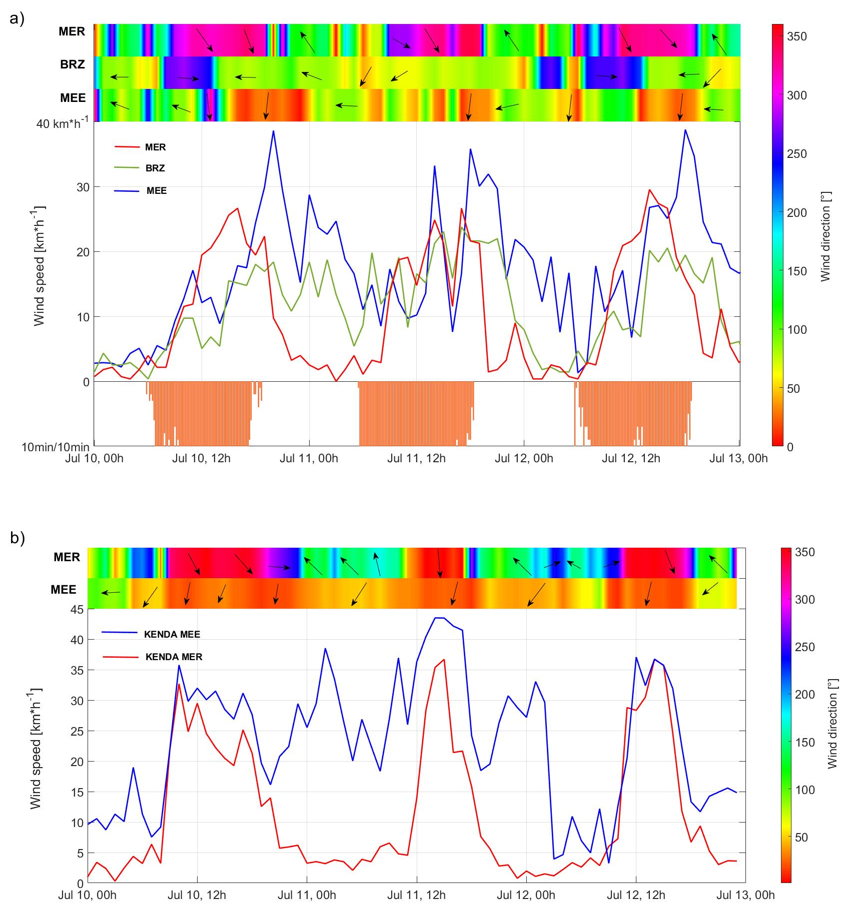

Figure 11. a) Measured and b) modeled wind speeds (solid lines), wind direction (colored bands and arrow) and sunshine duration (orange bars) for a) the DWL/MEE (775 m), the SMN/BRZ (577 m), the SMN/MER (584 m) and b) KENDA-1/MEE (775 m) and KENDA-1/MER (775 m).

Strong heterogeneities in the wind pattern along the Haslital are observed not only in this detailed analysis of thermal wind during summer, but also in the previous analysis of the median monthly wind. The comparison of KENDA-1/MER and KENDA-1/MEE (Fig. 6 b and c) wind profiles confirms the perturbation of the thermal wind diurnal cycle in the Haslital by the

Sarneraatal winds. In MER, the diurnal cycle of along valley winds is more pronounced with a clear extension up to 2000 m in November, March and April, a more delayed onset as a function of altitude in spring and a more constant wind direction during all months (Fig. 7). Generally, the onset of down-valley wind is better defined in MER because of the weaker influence of winds from the Sarneraatal and the overall higher wind speeds. It has to be noted that up-valley winds modeled at MER take almost the same direction (300-310°) as at MEE (290-300°), even if the valley bends (≈ 30°) between the two sites, except in the early morning (sunrise to 10:00) when up-valley winds come from W at low altitude. This near ground direction difference is similar to the observed winds at MEE, but happens earlier (from sunrise) and disappears at 10:00. Modeled down valley winds in KENDA-1/MER always follow the main longitudinal valley axis, as in KENDA-1/MEE.

3.4 Foehn events

Foehn is a katabatic wind leading to clear weather conditions on the northern side of the Alpine ridge. At MER, the foehn wind blows from the Grimsel Pass and follows the Haslital. The study of the T during foehn events combines all the periods where foehn was identified at SMN/MER, according to the foehn index in MER. It represents 117 hours of foehn during clear weather in March and slightly overcast sky (50-70% of maximum global radiation) in April and June. A detailed study on the wind is then only performed for three selected events (10-16 March 2022/19-22 March 2022/26-24 April 2022).

3.4.1 Temperature during foehn events

During foehn events, the MWR/MEE tends to measure 0.5-1.5°C lower T than the SMN/MER (Fig. 12.a), which can be partially explained by the different sites locations and altitudes. In contrast, a significant KENDA-1/MER and KENDA-1/MEE T underestimation of -2 to -4 °C is observed regardless of the time of day. Furthermore, the differences categorized according to the measured wind speed (Fig. 12.b) show that higher wind speeds (> 20 km/h) induce higher median T underestimations. Saigger and Gohm (2022) performed simulations in the Inn valley with the Weather Research and Forecasting (WRF) model and observed a similar bias at low altitudes during an intensive foehn event. In addition, Tian et al. (2022) also report significant cold and moist biases in the model during foehn hours. Note that KENDA-1/MER is in better agreement with SMN/MER than KENDA-1/MEE, which can indicate significant differences in the influence of foehn at the two stations.

The comparison of T profiles during foehn events in March (Figs. S11 and S12) shows that KENDA-1/MEE and KENDA-1/MER underestimates the T not only at the surface but up to 900-1400 m depending on the event. In some cases, KENDA-1 missed the T increase due to foehn and in other cases KENDA-1 follows the T evolution but with a smaller T. The median T bias of 2-4°C observed at the surface is also measured along the profile and is reinforced when a T inversion missed by KENDA-1/MEE precedes the foehn event. The increase in T due to the foehn breakthrough measured by the MWR/MEE is delayed by less than one hour compared to the SMN/MER detection. A similar one hour's delays from SMN/MER are modeled by KENDA-1, with shorter delay at MER than at MEE as expected by the orientation of the Haslital and the provenance of foehn.

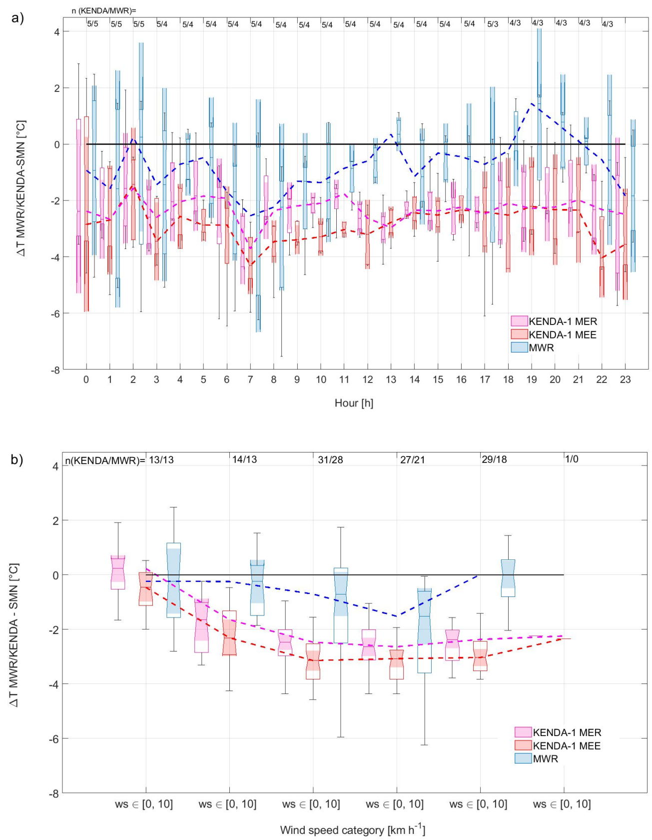

Figure 12. Box plots and whiskers of ground T differences between MWR/MEE and SMN/MER (blue), KENDA-1/MEE and SMN/MER (red) and KENDA-1/MER and SMN/MER (pink) as a function of a) the hour of the day and b) the 10 m measured wind speed at SMN/MER for all foehn events during the campaign. The lowest level corresponds to 584 m for SMN/MER, 625 m for MWR/MEE and 775 m for KENDA-1/MEE and KENDA-1/MER. The dashed lines represent the median of the different distributions and n is the number of cases in each of the categories.

3.4.2 Wind during foehn events

DWL/MEE measurements (Fig. 13.a) shows the extend of higher wind speeds induced by the foehn from ground to 1800-2000 m for a selection of three cases in March and April 2022. The foehn breakthroughs are nearly simultaneous at ground (SMN/MER) and up to 1000-1500 m at DWL/MEE for the events of March 11 and April 23. For March 20, an important delay of ≈ 3 his measured between 800 and 1300 m, whereas foehn winds are measured from 1300 m to 2000 m at the same time as at SMN/MER. The maximum wind speed (60-75 km/h) of DWL/MEE is observed at 800 m and is much higher than that at the SMN/MER (45 km/h) on March 11. The wind speed at the lowest level of the DWL/MEE is usually similar to that at SMN/MER. Thus, these three analyzed events exhibit some similarities but also large differences.

Concerning KENDA-1 data, the foehn breakthrough is modeled too early on March 11 at both stations, on time on March 20 at both stations and on April 23 at MER and too late on April 23 at MEE. The foehn arrival and end is modeled sometimes on time by KENDA-1, but positive and negative time shifts of up to 4h at both stations. The modeled wind directions are also often shifted by more than 100° (Fig. S13a). The foehn speed is often overestimated or underestimated by 20-30 km/h at all altitudes by KENDA-1/MEE (Fig. S13b). KENDA-1/MER models very high speeds of 75 to 110 km/h from ground level up to 1500 m, which is twice as fast compared to the DWL/MEE observations located only 5 km further down in the valley. Even though the Haslital is narrower just before MER (1.b), such difference in wind speeds suggests a potentially large overestimation of the foehn speed at this location. Finally, the simultaneous wind speed overestimation and the T underestimation by KENDA-1 during foehn events are difficult to explain since a stronger foehn should allow for a greater T increase.

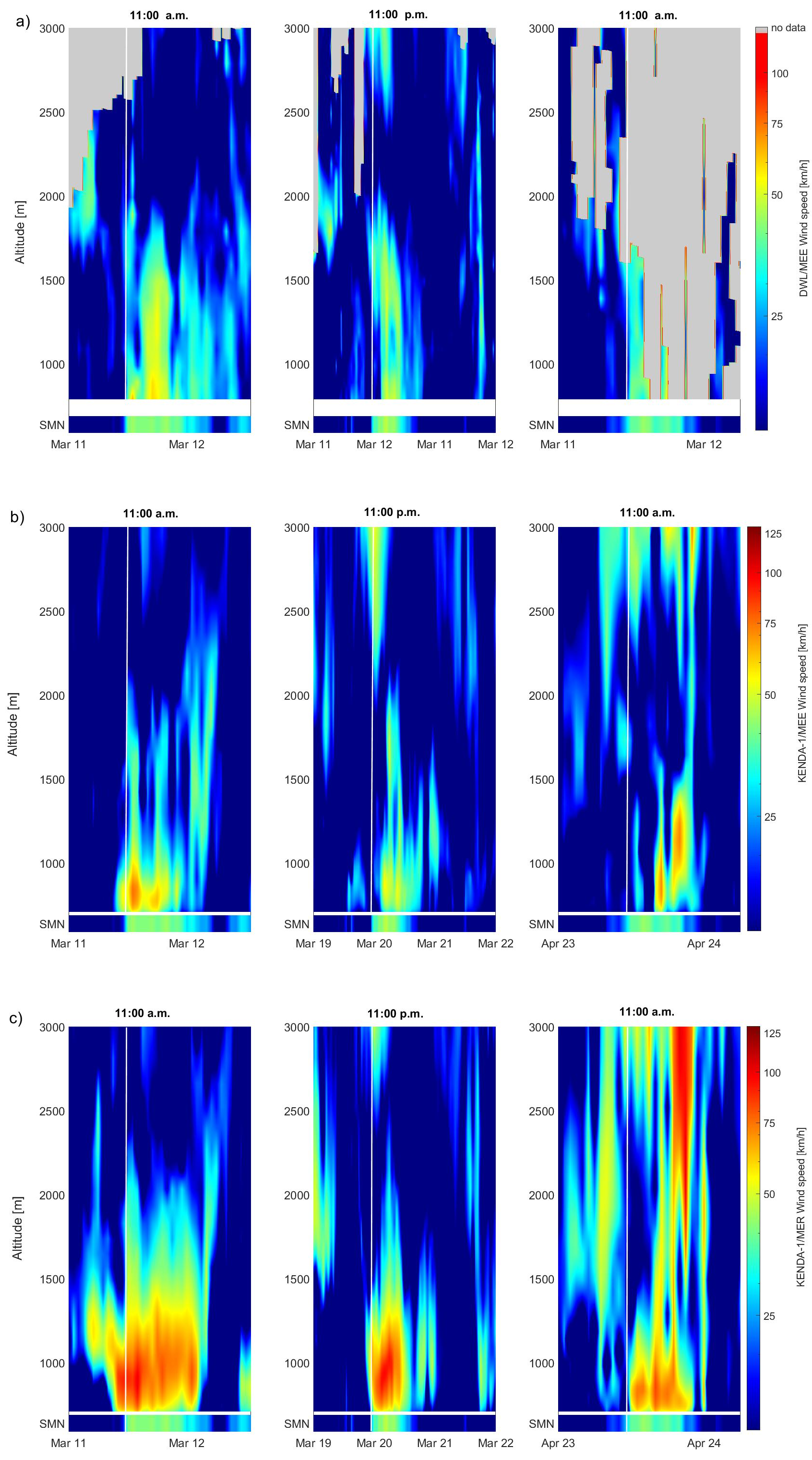

Figure 13. Wind speed profiles [km/h] time series from a) DWL/MEE, b) KENDA-1/MEE and c) KENDA-1/MER during a selection of 3 foehn events: left 11-12.03.2022, middle 19-22.03.2022 and right 23-24.04.2022. Wind speeds [km/h] from the SMN/MER are given in the lower part of each figure. The solid line represents the foehn breakthrough.

4 Discussion

Complex topography, landscape heterogeneity, and specific thermal wind regimes challenge the spatial and temporal resolution of the models, their performance in data assimilation, and the parameterization of multiscale processes. The discussion will therefore focus on three points, the characteristics of the terrain around the campaign site, the comparison of the observed wind and T profiles with previous observations in the Alps and the model performance in MER and MEE.

4.1 Topographical and methodological challenges

The Haslital presents several peculiar topographic and landscape characteristics, particularly in the vicinity of the campaign site (Fig. 1). Its junction with the Sarneraatal via the Brünig Pass links the two valleys with an angle of $\sim 90°$, 400 m above the valley floor. As described in 3.3, the valley volume effect explains that colder air from the Sarneraatal tends to fall into the Haslital from the Brünig Pass. It allows winds from the Sarneraatal to easily reach the Haslital with a cross-valley wind component similar to downslope winds and to disturb its along-valley wind system. This phenomenon is enhance in case of Bise situation, a N-NE synoptic winds that occurred on 35 days in the January-August 2022 period. The location of MEE directly below the Brünig Pass is therefore essential for comparison between MEE and MER results. Based on numerical simulations in the Alpine Inn Valley, Zängl (2004) suggests that variations in wind intensity are mainly related to tributary valleys, which increase or decrease the mass flux in the main valley. In this regard, low passes can have similar effects as tributaries. Moreover, the model grid cell containing MEE is situated on the valley's floor, but the Brünig Pass is only 200 m above MEE in the model terrain and the model does not consider the presence of lakes in the Sarneraatal. DWL/MEE, on the other hand, only observes winds in the middle of the Haslital with lower influence of the south facing slope. Consequently, the differences between the modeled T/wind averaged values and the observations cannot be considered as model errors only.

In addition, the curving of the valley between MER and MEE implies that the valley side faces different orientations along the Haslital leading to differential heating by the incoming solar radiation. The presence of large lakes covering the entire valley floor on its down valley side, in a distance of 5 km to the west of MEE, modifies the heat exchange between the surface and the atmosphere due to their high thermal inertia. Their influence on T along the valley can affect the pressure difference and, consequently, the time, vertical extent, and strength of the thermally induced valley winds. When comparing observed phenomena with similar studies, the combination of the above mentioned peculiar features gives explanatory hints for the observed differences. Finally, this study is based on monthly median values, so that the averaging artifacts have to be considered, e.g. for the analysis of maximum wind speed, the onset time of valley wind or wind directions. In that sense, this analysis focused on climatology and not on the forecast skills of the KENDA-1 model.

4.2 Comparison of observed phenomena with other studies

4.2.1 Occurrence of surface based T inversion in valleys

T patterns in MER follow a classical seasonal and diurnal cycle. The most important characteristic in the context of this study is the presence of frequent ground T inversions (Fig. 2.a, 5.a). According to a 3 year study in the French Jura performed over 16 station pairs at different altitudes (Joly and Richard, 2019), T-inversions are equally common in winter and summer (60% of the time), but with a larger amplitude (3 °C) in winter than in summer (2 °C). Additionally, temperature inversion occurred also more than 50% of the time in a 13 years T climatology in the Cascade Range, USA, at comparable altitudes (Rupp et al., 2020), with the formation and dissipation of inversions consistently having an approximately four hours time difference from sunset and sunrise. Finally, a 56-year climatology in the Austrian Alps (Hiebl and Schöner, 2018), shows that T inversions occur throughout the year with a frequency of about 30% from October to January and 15% from April to August. The intensity, magnitude and thickness of these surface T inversions follow a similar seasonal pattern as observed in the Haslital. Inversions are more frequent in eastern Austria, less frequent in the wide western valleys and basins, and almost vanishing in the high-Alpine summit area. This campaign in Haslital (Fig. 5.a) shows a similar occurrence of near-ground T inversions, i.e. 30% between the two ground stations (MER-BRU) and 40% in the MWR profiles. The amplitudes are similar to the results of Joly and Richard (2019) with slightly higher values during the winter months (+ 1°C). The seasonality of the phenomena is mainly characterized by the frequency of T inversions during the day in winter and the onset of the erosion process.

4.2.2 Characteristics of valley winds in the Alps

Previous studies on diurnal valley winds in Alpine valleys were carried out in the Rhone (length = 140 km, floor width = 4-5 km, ridge-to-ridge width = 15 km, Schmid et al. (2020)), in the Adige (L = 140 km, BW = 2-3 km, RRW = 8 km, Giovannini et al. (2017)) and in the Inn valley (L = 140 km, BW = 4-5 km, RRW = 20 km, Adler et al. (2021)). These three valleys are relatively long and wide compared to the Haslital (L = 30 km, BW = 1.5 km, RRW = 5 km), which can induce differences in the thermal valley wind systems. All three studies make a selection of valley wind days by using threshold on minimum global solar radiation or up valley wind speeds and selected global weather type.

Similarly to the observations in the Haslital, the change in wind direction in the Rhone valley (Schmid et al., 2020) occurs for altitudes up to about 2 km a.g.l. with diurnal pattern undergoing significant changes during the course of the year. During summer, maximum up valley wind speeds of 30-35 km/h are found above the Rhone valley during the early afternoon at 200 m a.g.l. Similar timing for maximum up-valley winds are found at both MER and MEE, but with reduced speeds both at ground (SMN/MER, 20-30 km/h) and at 200-300 m a.g.l. (DWL/MEE, 15-20 km/h) which can be related to some extend to the absence of clear-sky day selection in this study. At MEE, the highest wind speeds of 30 to 45 km/h are found later on, at 18:00 and 19:00, between 800 and 1400 m and correspond to valley winds from the Sarneraatal. The topographic difference between the Brünig Pass and the standard tributaries' inlet at the campaign site in Sion can also explain the time and altitude differences of the strongest winds. Concerning down valley wind speeds, Schmid et al. (2020) report their presence between 500 and 1000 m.a.g.l with wind speeds of about 15-20 km/h. They occur in the second part of the night in spring and summer, and during the entire night in winter. Several differences are observed in the Haslital: 1) down valley winds reach the ground even in summer (Fig. 7) and extend up to 800-1000 m.a.g.l., 2) their speed gradually decreases during the night with almost no wind between 00:00 and the new onset of up-valley winds, and 3) at MEE, maximum down valley wind speeds are measured from March to July at the same altitude as in the Rhone valley but with lower wind speeds (10-15 km/h). If the last difference can also be explained by the applied monthly average, the timing and extent of the down valley winds probably relates to topography differences.

In the Adige valley in the Italian Alps, a campaign in May-August (Giovannini et al., 2017) observed maximum up-valley wind speeds between 15:00 and 16:00 that are stronger near the valley outlet (20-30 km/h) and gradually weaken (8-10 km/h) towards the highest valley parts located 100 km further up. Surface down-valley wind speed appears to be very weak (0-5 km/h), and nearly constant in the entire valley. However, in contrast to the Haslital and the Rhone valley, the down valley wind onset is delayed to 00:00. Wind profiler data from the outlet of the Adige valley show that the strongest up-valley winds are recorded in the late afternoon, similarly to the observations at MEE (Fig.8.a). In contrast to both Schmid et al. (2020) and this study, the down-valley winds of the Adige valley gradually weaken toward higher altitudes around midnight. For the rest of the night, stronger wind are also found between 500 and 1000 m.a.g.l. similarly to the observation in the Rhone valley (Schmid et al., 2020).

Finally, both the time and the pattern of the onset of up valley wind are similar in the Rhone, the Adige and the Haslital valleys. The onset occurs 3-4 hours after sunrise with flows that move almost simultaneously between 0 and 1500 m.a.g.l from June onward due to a rapid warming by short-wave solar radiation. During the evening transition, the down-valley wind begins at the ground due to the progressive cooling of the lowest atmospheric layer (Zängl, 2004) and thickens during the night. Note that, Schmid et al. (2020) reported a delayed onset as a function of altitude in autumn but unfortunately, no data were acquired during this period in the Haslital.

The CROSSINN campaign (Adler et al., 2021) was carried out from August to October in the lower part of the Inn valley and focused on cross-valley winds. For two days in September, the wind field in the vertical plane across the valley shows an enhanced cross-valley wind circulation in the second part of the afternoon (15:00-17:00). Over the south facing slope of the valley, subsidence prevails, while over the north facing slope upward motion is measured. This flow pattern forms a closed circulation cell with a clear cross-valley component comprising a northerly component in the lower 700 m.a.g.l. and a southerly component above. Similarly to the Inn valley, the Haslital at MEE also lies in the E-W direction and the valley bends between MEE and MER. A cross-valley circulation is also observed from March to August (Fig. 9. a), with a change in wind direction from N to S between 450 and 850 m a.g.l. and a stronger pattern in summer. However, contrary to the CROSSIN campaign's results, valley winds from the Sarneraatal are probably the main drivers of this cross valley circulation in MEE.

4.3 Model performance

According to the presented results, KENDA-1 is generally able to capture the main features of the observed atmospheric conditions. This is remarkable given that the complex topography in the region of this study is only marginally resolved by KENDA-1. It is thus not surprising that some meteorological phenomena specific to mountainous regions and/or particular synoptic conditions are hard to capture by the model.

4.3.1 KENDA-1 skill in temperature estimates

The analysis of the diurnal cycle shows that the majority of ground T differences with respect to observations lays between ± -3 °C (Fig. 4) with a nighttime overestimation and a daytime underestimation by KENDA-1. In a study over complex topography (Alpine arc and particularly Switzerland and northern Italy) Voudouri et al. (2021) found a similar diurnal cycle in ground T mean error in COSMO-1E forecasts, but of reduced amplitude (-0.5 °C bias during day and a +0.5 °C bias during night). Despite the complex topography in the vicinity of MER and the induced elevation bias, the modeled climatology of ground T is satisfactory, even if differences of up to 8 °C are found in some periods. The main explained source of ground T differences is caused by missed surface T inversions. The frequency of this phenomenon is partially missed by KENDA-1 from March to August (Fig. 5.a) and its amplitude is underestimated for all months. In particular, KENDA-1/MEE missed the strong T inversions at the end of March (results not shown), which are enhanced by night-time radiative cooling and daytime surface heating due to very low cloud coverage and deficit in precipitation (3). The observed differences in amplitude are mainly due to an underestimation of T at the ground level (Fig. 4). A work carried out by Sekula et al. (2019) on the nonhydrostatic model CY40T1 AROME CMC (2km horizontal resolution) showed the same general overestimation of the minimum T at the bottom of the valleys. The largest differences were measured during strong high-pressure systems, which favors the formation of cold air pools, leading to T overestimations of up + 7 to 9 ° C for 10 days in March.

A preliminary analysis on KENDA-1 behavior during these strong T inversions shows that the observed differences are probably due to a too low ensemble spread of model first guess. The model is too much trusted in the model-observation weighting scheme and measured T at MER are therefore not used in the model assimilation step, what on the other hand is necessary to avoid instabilities in the data assimilation step. Another hypothesis is that a too large observation error is assigned to the station of MER (1.17K end of March). Furthermore, in this period, the difference between the observed and modeled ground relative humidity (RH) remains within $\pm$ 5% during the day, but during the night the model is much drier (-20 to -30 % RH, not shown). Westerhuis et al. (2021) showed, particularly during conditions favorable for surface T inversion. The KENDA-1 vertical coordinates follow the terrain. Therefore, in complex topography, numerical artifacts may originate from the intersection between T inversions and the surface of the vertical grid used by the model. The systematic T underestimation during night can also be driven by an overestimated modeled cloudiness involving underestimated out-going long-wave radiation. Further investigations have to be performed using ceilometer and/or DWL observations to estimate the model skill with respect to cloud cover. Finally, it is hypothesized that the differences with observations can also originate from a modeled ongoing turbulent mixing whereas in reality a cold pool with a full or partial decoupling from the above flow is present in the valley.

For the T profile comparison, MWR/MEE T is used as reference, but the uncertainties regarding its reliability, especially at high altitude, must be considered in evaluating the KENDA-1 results. Löhnert and Maier (2012) performed a MWR-RS comparison and showed that the random error inherent to the measurement principle can be important in some cases. They showed that the random error range increases to 1.7 K at 4 km height, due to a 95% influence of the profile used as apriori. KENDA-1/MEE and MWR/MEE T profiles differences are constrained to $\pm$ 1 °C for all altitudes between 1400 and 2200 m both day and night except in June and July (Fig. 2.b). Differences of up to -3 °C can occur near the ground in winter or at ridge level in July. The overall negative bias can be explained mainly by two factors: first, the MWR is susceptible to errors, especially at higher altitudes with RMSE between 1 and 1.5 ° C (Liu et al., 2022), and second, the MWR/MEE has been trained with sounding profiles from Payerne, so that the difference in altitude between both stations (+100 m) and in the atmospheric conditions could induce a larger RMSE or even a bias in the MWR measurements. Despite these uncertainties, the differences in T up to -3 °C are probably a clear underestimation of KENDA-1 T. The hypothesis of cloud amount overestimation mentioned before can also explain this T profile bias.

4.3.2 KENDA-1 skill in wind estimate

The monthly valley wind reveals a good performance of the model. Up and down-valley winds are in good agreement with the observations from March to July and, to a lesser extent, in November and February. KENDA-1 is also able to get the seasonal evolution of the vertical extent of the valley wind system. However, the onset of up-valley winds is predicted too early after sunrise (Figs. 6 and 8). This 1-2 hour difference from the observations is partially explained by the absence of surface T inversion in the model (sect. 3.1.3), so the time that allows for erosion of the stable layer is not taken into account.

The capability of COSMO models to estimate the diurnal along-valley winds in real valleys has been investigated by Schmidli et al. (2018) for 3 summer weeks with weak synoptic forcing and intense solar heating. The model results are compared to observations at the MeteoSwiss ANETZ stations, the automatic monitoring network preceding the present-day SMN. They showed that the wind diurnal cycle is well represented by COSMO1-E in large valleys such as the Rhine Valley at Chur (base width of 3 km and width at half height of 8 km) and medium valleys (e.g. the Rhone Valley at Visp with base width of 1 km and width at half height of 4 km). For smaller valleys, e.g. the Maggia Valley in Cevio (base width of 500 m, width at half-height of 3 km), the valley wind amplitude was underestimated. Despite an underestimation of the maximum valley wind speed, the onset of up and down valley winds was correctly modeled. The results of the modeled wind speed and direction at MEE are comparable to the analysis in Visp (Fig. 8), a valley with a similar cross section. However, the onset of up and down valley winds shows lower agreement with the observations at Meiringen, probably due to the four-time shorter length of the Haslital and its topographic peculiarities.

The differences between KENDA-1 and the observed cross-valley wind climatology (Fig. 9) can be interpreted as a overestimated influence of the Sarneraatal thermal winds in the model world or as an effect of grid cell overlap on the north-facing slope. The presence of strong down slope winds at the Brünig Pass may have a direct influence on the along valley wind diurnal cycle. In a recent study in the Rhone valley in Sion, Schmidli and Quimbayo-Duarte (2023) reports a correctly modeled evening transition but an inadequate representation of the morning wind reversal by COSMO-1E. Like in the Haslital (Fig. 9), the overestimated cross-valley wind in the model reaching the valley floor interrupts the formation of the up-valley flows for certain days. In Sion, the cross-valley flow is restricted to upper levels so that the stronger lower valley atmosphere stratification protects the up-valley flow.

According to (Schmidli et al., 2018), the horizontal resolution required for a accurate wind representation along the valley requires at least 1-2 grid cells in the base cross section of the valley. A more important feature is the altitude bias of the model at the ground. For the MER station, the width of the valley can contain 1.5 grid cells (Fig. 1) but the fact that no cell contains only the valley floor leads to a disfavouring bias in altitude. Surface atmospheric moisture is a key factor of stratification, which in turn favors the cross valley winds influence. Simulations performed by Schmidli and Quimbayo-Duarte (2023) show that a 30% increased soil moisture relative to KENDA-1 data leads to better along valley wind modeling. Even though stronger smoothing of the topography improves the stratus cloud simulations, it also decrease the quality of forecasts of valley winds and orographically induced convection (Westerhuis et al., 2021).

Finally, despite the fact that KENDA-1 agrees well with with the observations in respect to monthly median values, the case-by-case analysis shows important differences from observations. No systematic differences are observed in most profiles. Even though these differences show regular patterns in the case of foehn or valley winds, it is common that unpredictable behavior affects the model.

5 Conclusion

The measurement campaign comprised two sites in the middle size Alpine valley of the Haslital. Ground measurements are operationally performed at SMN/MER, whereas REM instrumentations (MWR, DWL and a ceilometers) were located at MEE. The Brünig Pass north of MEE is situated only 400 m over the Haslital floor and open to the bigger valley of the Sarneraatal. This 10 months campaign (November 2021 and August 2022) yields valuable information on the diurnal and seasonal cycles of wind and T profiles that were not available in this region and that are rather sparse in Alpine middle size valleys. In parallel to these observations, the data of two grid cells of the KENDA-1 assimilation model has been analyzed and compared to the measurements.

Regarding the observed and modeled T, the main results concerns the surface based T inversion. Nighttime T inversions are commonly observed during all the months under study with bigger amplitudes during December and January and a persistence during daytime from November to February. The frequency of occurrence and the amplitude of the surface T inversions are both underestimated in the T profiles of KENDA-1. This results in a systematic overestimation of the ground T during the presence of surface based inversions. In extreme cases it reaches up to 8 °C. This large model error has an important consequence, since the discrepancies between the model first guess prevents the SMN/MER observations to be assimilated. Apart from this, the differences between MWR/MEE and KENDA-1/MEE profiles are small with a T underestimation of -2 to -3 °C under 1500 m that is more frequent during nighttime.

Thermal valley winds are observed clearly from April to August, slightly in November, February and March, but are absent in December and January. This diurnal flow pattern develop in a more distinct way for the summer months (June to August). The vertical extent of down-valley winds after sunset increases from February to August: from 600 m a.g.l. to 1600 m a.g.l. respectively. The morning transition to up valley wind is delayed by about 3-4 hours compared to sunrise and takes place nearly simultaneous for the entire the profile. The onset of down-valley winds happens less than an hour before sunset and propagates from ground to ridge height in some hours. In addition, this thermal wind system can be influenced by external factors such as synoptic wind intrusions or perturbation from adjacent valleys wind system. At MEE, N winds from the Sarneraatal through the low altitude Brünig Pass are observed regularly from mid-afternoon to sunset and from ground to the altitude of the pass. They are due to colder air masses from the Sarneraatal. This valley has in fact a 1.7 higher volume than the Haslital, leading to a slower warming by insolation. At MEE, these flows affect the evening transition and sometimes even the along valley wind pattern during daytime below the altitude of the pass. If these N flows only slightly modify the up valley wind direction at MER, they are able to suppress the up valley winds at BRZ. In summer, a cross valley circulation is measured around sunset (19:00-20:00) at MEE with a separation between north and south facing wind between 700 and 1000 m a.g.l. The formation of the cross-valley circulation is influenced by the strong wind from the Sarneraatal.

Comparison with observations shows that KENDA-1 was able to simulate the median directions and speeds of the thermally driven valley winds. The vertical extent of the thermal winds, the onset time of down valley winds and the interaction with synoptic winds are also appropriately modeled. However, KENDA-1 shows a too early (1-2 hours) onset of up-valley winds that can be partially explained by the absence of the near-surface stable layer caused by the nighttime inversion. Moreover, the observed cross-circulation in MEE at sunset is not captured by KENDA-1.

Unlike monthly values, the analysis of single profiles shows important differences between the model and the measurements. This is particularly true during foehn events with a near systematic underestimation of 2 to 4°C by KENDA-1 in both the ground and the profile temperatures. Wind speeds simulation during foehn show significant difference over MEE and MER: the KENDA-1/MEE show a good match up to 1000 m a.g.l. whereas KENDA-1/MER reports wind speed twice as high (120 km/h). A detailed analysis of three clear sky summer days also allows to underline distinct differences between the observations and the model concerning the wind direction (up to 90°), the wind speed (up to 30 km/h) and the timing (up to 4-6 h) of the along valley transition.

The results nicely illustrate the complex interaction of various meteorological processes in an Alpine valley. Despite the descriptive approach used in this study it highlights many open questions and reveals that further effort is needed by the community to deepen our knowledge regarding meteorological processes in complex terrain and the interaction of processes at various scales. One example of such a complex interaction is the wind that falls from the Sarneraatal to the Haslital's floor through the Brünig Pass. However, many observed phenomena are not yet satisfactorily characterized and modeled and require further investigation. A better understanding of the exchange processes in complex topography and the ability of the model to take them into account are an essential conditions to improve the prediction capacity of NWP in complex mountainous terrain.

Data availability. Data are available on request

Author contributions. AB did the analysis, AB and MCC prepared the manuscript. MH and SM operated the instruments during the campaign. DL and MA provided the model data. All co-authors contributed to the manuscript online.

Competing interests. The authors declare that they have no conflict of interest.

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