The Influence of the Resolution of Orography on the Simulation of Orographic Moist Convection

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(Manuscript received 25 July 2019, in final form 8 March 2020)

ABSTRACT

Currently, major efforts are under way to refine the horizontal resolution of weather and climate models to kilometer-scale grid spacing (Δx). Besides refining the representation of the atmospheric dynamics and enabling the use of explicit convection, this will also provide higher resolution in the representation of orography. This study investigates the influence of these resolution increments on the simulation of orographic moist convection. Nine days of fair-weather thermally driven flow over the Alps are analyzed. Two sets of simulations with the COSMO model are compared, each consisting of three runs at Δx of 4.4, 2.2, and 1.1 km: one set using a fixed representation of orography at a resolution of 8.8 km, and one with varying representation at the resolution of the computational mesh. The spatial distribution of precipitation during daytime is only marginally affected by the orographic details, but nighttime convection to the south of the Alps—triggered by cold-air outflow from the valleys—is very sensitive to orography and precipitation is enhanced if more detailed orography is provided. During daytime, the onset of precipitation is delayed. The amplitude of the diurnal cycle of precipitation is reduced, even though more moisture converges toward the Alpine region during the afternoon. The hereby accumulated moisture sustains precipitation during the evening and nighttime over the surrounding plains. For these differences, the effects of changes in orographic detail are more important than changes in grid spacing. In addition, the individual convective cells are weaker, but their number increases with higher resolved orography.

1. Introduction

Moist convection is one of the key mechanisms controlling the vertical redistribution of heat, water vapor and momentum in the atmosphere. Thereby, it interacts with other atmospheric processes and regulates the

climate system. It is thus important to have reliable simulations of moist convection in the contexts of both numerical weather prediction and climate simulations. Several processes may lead to the formation of precipitating convective cells. They can be triggered by frontal ascent (Hogan et al. 2002; Rasp et al. 2016) or by planetary boundary layer (PBL) processes (Crook 1996; Trier et al. 2004) like for instance convergence of air in the PBL. In the latter case, shallow cumulus clouds that form at the top of the PBL may experience a transition to deep convection after which they can produce substantial amounts of rain. This transition can be fostered

Denotes content that is immediately available upon publication as open access.

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DOI: 10.1175/MWR-D-19-0247.1

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by upper-level destabilization (Juckes and Smith 2000) and it can be hindered by the dilution of moist-adiabatically rising air parcels through clear-air entrainment (Kirshbaum 2011). Cloud organization, on the other hand, reduces the effects of clear-air entrainment (Kirshbaum and Grant 2012) and can help convection to deepen. Also, a suitable moisture distribution within the PBL with patches of high moisture content may locally enhance the moisture available to the convective cells (Schlemmer and Hohenegger 2014, 2016). These patches can for instance be generated by convergence of cold pools from existing cells.

This study focuses on moist convection originating over and in the vicinity of mountainous terrain. Over complex orography, additional mechanisms may favor the initiation of convection. On fair-weather days, thermally induced wind systems form, consisting of a mesoscale mountain–plain wind system and local valley and slope wind systems (e.g., Wagner 1932; Egger 1990; Langhans et al. 2013). The mesoscale mountain–plain wind system is driven by differential heating between the air over the mountains over the surrounding plains. Previous studies found that the concept of the valley volume effect (Steinacker 1984; Whiteman 1990) gives an upper limit for the additional bulk warming over mountains compared to the surrounding plains, which may be reduced by local thermal circulations like the cross-valley circulation (Serafin and Zardi 2010, 2011; Schmidli 2013; Rotach et al. 2015). Additionally, turbulent heat exchange at the surface was found to enhance heating over orography due to the thermally driven winds that formed on sloping surfaces (de Wekker et al. 1998). The temperature anomaly over mountains generates a thermal low during the day and a thermal high during the night. The mountain–plain wind system evolves accordingly—it converges to (plain-to-mountain winds) and diverges from (mountain-to-plain winds) the mountains during the day and night, respectively. In the European Alps, this circulation has been illustratively called “Alpine pumping” (Lugauer and Winkler 2005; Langhans et al. 2013). The horizontal inflow during the day generates a surplus of moisture and heat over the mountains compared to the surrounding plains. The valley and slope wind systems transport the incoming air up the mountain valleys and slopes eventually causing convergence of PBL air and moisture over the mountain tops (e.g., Van Baelen et al. 2011; Adler and Kalthoff 2014). This convergence locally generates updrafts, which help overcoming the convective inhibition (CIN). The supply of moisture and heat additionally favors the initiation of deep convection by destabilization (e.g., Demko and Geerts 2010) and by counteracting the dilution from clear-air entrainment (e.g., Kirshbaum 2011). Because of such wind systems,

the flanks of mountain ranges like the European Alps exhibit an increased fair-weather daytime convective activity compared to the surrounding plains (e.g., Kuo and Orville 1973; Frei and Schär 1998; Weckwerth et al. 2011). During the night, on the other hand, convection is inhibited over the mountains due to subsidence caused by the thermal high. The diverging mountain-to-plain winds transport moisture away from elevated regions to the surrounding plains fostering evening and nighttime convective activity over these areas (Yu and Jou 2005; Sato and Kimura 2005; Mazon and Pino 2013). Also, forced lifting of the preexisting flow by drainage winds has been found to trigger convection (Barthlott et al. 2016).

Atmospheric models begin to explicitly resolve moist convection at a model grid spacing (Δx) of approximately 4 km or less (Ban et al. 2014; Prein et al. 2015). For larger Δx , convection is generally treated as a subgrid-scale process and parameterized. The value of convection-resolving models for the purpose of numerical weather prediction has been recognized for several years (e.g., Hohenegger and Schär 2007; Lean et al. 2008; Weusthoff et al. 2010). In recent years, studies have demonstrated the added value of explicitly resolved convection also for the simulation of precipitation in climate simulations (e.g., Hohenegger et al. 2008; Kendon et al. 2014; Ban et al. 2014; Leutwyler et al. 2017). Recent advances in computing power and architecture made it possible to increase the domain size and integration time allowing for decade-long continental-scale climate simulations at convection-resolving resolution (e.g., Leutwyler et al. 2016; Prein et al. 2017; Hentgen et al. 2019) and even global-scale convection-resolving simulations have been conducted over week-long periods (e.g., Miyamoto et al. 2013; Bretherton and Khairoutdinov 2015; Fuhrer et al. 2018; Neumann et al. 2019).

Within the convection-resolving range, scale dependencies of prognostic model variables on Δx have been observed. For instance, Langhans et al. (2012) studied summertime convective activity over the Alps. They ran simulations between 4.4 and 0.55 km Δx and found that the simulated convective cells scale with Δx . The clouds get smaller and more numerous with smaller Δx implying that Δx imposes a visible scale on the simulated fields. Model grid spacing Δx also determines the details in the representation of orography, which—apart from influencing the generation of gravity waves and orographic form drag—potentially affects the surface forcing of orographic moist convection. For instance, a smaller Δx gives a more accurate representation of steep valley slopes which was found to influence the simulated valley winds (Serafin and Zardi 2010; Wagner et al. 2015a,b; Schmidli et al. 2018). Langhans et al. (2012) and Panosetti et al. (2019), who studied convergence across

different Δx in real-case simulations over the Alps, kept the representation of orography fixed in their studies to circumvent such changes with higher Δx . Other studies explicitly smoothed orography to investigate its influence on surface precipitation. For instance, [Schumacher et al. \(2015\)](#) compared the location of rainbands in a lee cyclone over the Rocky Mountains between simulations with strongly smoothed orography and a nonsmoothed control run and found small differences only. They could attribute the formation of the rainbands to ascent up the large-scale orography, rather than ascent along small-scale topographic features. In a similar analysis, [Picard and Mass \(2017\)](#) and [Prein et al. \(2013\)](#) also found only small differences when comparing precipitation fields resulting from simulations using smoothed against nonsmoothed mountain ranges. Other studies, however, found that small-scale details in orography play a decisive role in the formation of rainbands or embedded convection ([Cosma et al. 2002](#); [Fuhrer and Schär 2007](#); [Kirshbaum et al. 2007](#)). These studies describe cases of orographic convection that is forced by mechanical lifting. Less is known about the effect of terrain detail on thermally driven orographic convection. In 5-yr-long simulations over Greece—and thus containing both mechanically and thermally forced cases—[Tselioudis et al. \(2012\)](#) found increased precipitation over mountains in simulations at a Δx of 11 km compared to 50 km. They traced this difference back to the higher surface elevation in the 11 km simulation caused by the higher resolved orography. In idealized simulations of exclusively thermally driven convection, [Panosetti et al. \(2016\)](#) and [Imamovic et al. \(2019\)](#) both found an earlier onset of deep convection and surface precipitation for higher Gaussian-shaped mountains.

The goal of this study is a systematic assessment of the sensitivities related to the resolution of orography in a real-case study of summertime moist convection over the Alpine area. The following research questions are addressed: How does the resolution of orography affect the processes driving orographic moist convection? How does it affect the resulting life cycle of convective cells (onset time, intensity and duration, as well as accumulated surface precipitation)? Do small-scale variations of orography affect the horizontal scales in the prognostic fields (vertical velocity, cloud field)? If so, what is the relative importance of these 'external scales' compared to the scale imposed by Δx ? We study these questions using two sets of numerical experiments, one with a resolution-dependent representation of orography, and one with fixed orography across a range of computational resolutions. In the following section, the model and the simulations are described. The results

of the study are shown and analyzed in [section 3](#) and discussed in [section 4](#).

2. Methods

a. Model description

The simulations are run with the model of the Consortium for Small-Scale Modeling (COSMO) version 5.0 ([Baldauf et al. 2011](#)). The COSMO model is a fully compressible, nonhydrostatic limited area model. It is used by several weather services around the globe for operational numerical weather prediction and has also been further developed into a regional climate model ([Jaeger et al. 2008](#); [Rockel et al. 2008](#)). A rotated latitude–longitude grid is used in the horizontal directions. The vertical direction is discretized using a pressure-based hybrid vertical coordinate. The time integration is performed with a third-order Runge–Kutta scheme ([Klemp and Wilhelmson 1978](#); [Wicker and Skamarock 2002](#)). Temperature, pressure and horizontal and vertical winds are discretized with a fifth-order advection scheme. A second-order scheme ([Bott 1989](#)) is used for the horizontal discretization of moist quantities. Radiation is parameterized by the δ -two-stream approach after [Ritter and Geleyn \(1992\)](#). Cloud microphysics are represented by a single-moment bulk scheme after [Reinhardt and Seifert \(2006\)](#). Subgrid-scale turbulence is parameterized with a 1D TKE-based model after [Raschendorfer \(2001\)](#) in the vertical. Horizontal diffusion is computed using a Smagorinsky–Lilly closure after [Baldauf and Zängl \(2012\)](#). No convection parameterization is employed (neither for deep nor shallow convection). The second-generation land surface model TERRA_ML ([Heise et al. 2003](#)) with 10 layers is coupled to the atmospheric part of the model.

b. Experimental setup

The model is run at Δx of 4.4, 2.2, and 1.1 km. The vertical grid is the same in all simulations and consists of 60 levels with a spacing ranging from around 20 m at the surface to around 1.2 km at the model top located at an altitude of around 23.5 km. Rayleigh damping is applied above 11.5 km to minimize the reflection of gravity waves. The runs are initialized and driven by a simulation with Δx of 12 km which is run over continental Europe [see [Ban et al. \(2014\)](#), their Fig. 1]. The driving simulation uses the same physical parameterizations as the nested models, except that convection is parameterized and that the horizontal Smagorinsky–Lilly eddy viscosities are replaced by numerical diffusion. A relaxation zone of 35 km is located at the domain boundaries of the nested simulations. The initial and lateral boundary data for the driving simulation stem from ERA-Interim reanalysis data ([Dee et al. 2011](#)). To make sure that the soil

FIG. 1. Orography within the simulation domain represented on a grid of $\Delta x = 1.1$ km. The full analysis domain (Alpine region) and the Po Valley subdomain are shown by the white rectangles A and E, respectively. The domain limited by the white rectangle B, as well as the red line C are both used to calculate latitude–altitude cross sections. The red line D delimits the area that is covered by the CombiPrecip dataset. Two geographical regions that are important for later discussions are labeled in orange.

moisture and the atmosphere are well equilibrated at the beginning of the simulations, the soil moisture profiles are not taken from the comparably coarse ERA-Interim reanalysis dataset. Instead, equilibrated soil moisture profiles from a 10-yr-long ERA-Interim-driven COSMO climate simulation (Ban et al. 2014) are employed.

Orographic moist convection over the Alpine region is studied during the nine days between 11 and 19 July 2006. This period has already been analyzed in similar studies on orographic convection [see, e.g., Langhans et al. (2012) and Panosetti et al. (2019) for the same days and Hohenegger et al. (2009), Langhans et al. (2013), or Schmidli et al. (2018) for a longer time period in July 2006]. It is characterized by relatively weak synoptic forcing and a strong diurnal cycle of convective activity. The nine days can therefore statistically be considered as an ensemble of quasi-stochastic realizations of orographic convection. The synoptic situation with the presence of a persistent high pressure system is a typical driver of summertime fair-weather thermally driven orographic convection over central Europe. The simulation domain covers an area of 1153×1082 km² and is centered over the European Alps as shown in Fig. 1. The analysis of Alpine-wide convective activity will be performed on the subdomain A shown in Fig. 1. In the

following, this region will be referred to as “Alpine region.” The subdomains B, D, and E, as well as the vertical section C are used for specific analyses that will be presented in section 3.

Two groups of simulations (represented by the rows in Fig. 2) are conducted to assess the influence of the resolution of orography on the convective activity. Each group consists of three simulations (represented by the columns in Fig. 2) which are identical, except for Δx which is 4.4 (left), 2.2 (middle), and 1.1 km (right). In the first group (top row—in the following referred to as SM, for smooth), the orography has identical resolution independent of Δx . In the second group (bottom row—in the following referred to as RAW), the orography is represented at the respective model resolution. The simulations will in the following be referred to as, for example, RAW1 (1.1 km raw simulation) or SM2 (2.2 km smooth simulation). By comparing the simulation output of the SM and RAW groups, it is possible to isolate the effect that the degree of detail in the represented orography has on the simulated orographic moist convection.

The procedure to obtain the SM and RAW orography datasets is described in the following. The underlying dataset, the Global Land One-kilometer Base Elevation Project (GLOBE; Hastings et al. 1999) is filtered using a fifth-order Raymond (1988) low-pass filter to derive orography datasets at 1.1, 2.2, 4.4, and 8.8 km resolution. The former 3 constitute the orography used for the RAW1, RAW2, and RAW4 simulations. The SM orography is generated like in Panosetti et al. (2019): The Raymond-filtered orography at 8.8 km is interpolated onto the SM1, SM2, and SM4 grids. This interpolation gives the same degree of detail between the different SM orography datasets in physical space, while introducing differences in spectral space. To ensure a smooth and consistent representation of orography across resolution also in spectral space, a 2D Gaussian-weighted moving window averaging low-pass filter is applied to all SM orography datasets. This gives the benefit of equal effective resolution (see, e.g., Skamarock 2004) across the different grids, but in turn, it introduces new very small differences in physical space. The latter is illustrated by the labels in Fig. 2, showing the maximum surface elevation over the Alpine region. Note that the RAW4 orography is resolved at roughly twice the resolution of the SM4 orography because the latter results from interpolation of the 8.8 km dataset. With each increment in model Δx , the difference in resolution between RAW and SM doubles. Consequently, the RAW1 orography is roughly eight times higher resolved than the SM1 orography. The interpolation and filtering applied to the SM orography conserves the mountain volume to a high

FIG. 2. Representation of orography within the analysis domain A in each of the six simulations. The simulations consist of two sets referred to as (top) SM (smooth orography) and (bottom) RAW (raw orography). Each set consists of three simulations with a grid spacing of Δx of 4.4, 2.2, and 1.1 km, represented by the three columns. The labels in the lower-right corner of the panels indicate the maximum elevation (m MSL) within the analysis domain.

precision (it differs by no more than 0.05% between corresponding SM and RAW simulations). The differences in orography are thus merely related to the represented degree of detail, not to a different mountain volume, similarly as in [Imamovic et al. \(2019\)](#). To retain a consistent representation of the terrain, the surface fields of the SM simulations are adjusted to the new orography by applying the same interpolation and filter as for orography. For categorical surface fields, a nearest-neighbor interpolation is used, and no filter applied. Elevation differences between RAW and SM arising from the treatment of orography are considered at model initialization by adjusting the atmospheric profiles and the soil temperature to the represented elevation. Temperature is extra or interpolated assuming the lapse rate of the closest points in the vertical column. There is no need to adjust snow cover since at this time of the year, no snow is left in the simulations.

c. Observational data

Two observational datasets of surface precipitation at hourly resolution are used to evaluate precipitation in RAW and SM.

First, the CombiPrecip dataset ([Sideris et al. 2014](#)), which is based on a combination of precipitation radar measurements from the Swiss weather service MeteoSwiss

and rain gauge measurements at the surface to calibrate the radar measurements. The radar covers an area around Switzerland as indicated by subdomain D in [Fig. 1](#).

Second, measurements from 104 surface observation stations over Switzerland coming from the MeteoSwiss ANETZ (automatic monitoring network) are used. For each measurement station, the closest model grid point to the station location is selected using nearest-neighbor interpolation and then compared to the measured precipitation.

3. Results and analysis

a. Convective features

1) MEAN DIURNAL CYCLE

[Figure 3](#) compares the mean diurnal cycle of precipitation between SM ([Fig. 3a](#)) and RAW ([Fig. 3b](#)). Precipitation is averaged over the Alpine region (subdomain A in [Fig. 1](#)). In SM, the diurnal peak precipitation and the total accumulated precipitation decreases with increasing model resolution. This is in agreement with the findings of [Panosetti et al. \(2019\)](#) for a similar domain and the same simulation period. Compared to the corresponding SM simulations, the peak intensity of the RAW simulations is further reduced. In the RAW1

FIG. 3. Mean diurnal cycle of rain rate (mm h^{-1}) (a)–(c) averaged over the analysis domain A; (d)–(f) averaged over the intersection of the analysis domain A with the radar domain D; (g)–(i) averaged over the grid points closest to the measurement stations. Shown are the results of the (left) SM and (middle) RAW simulations, as well as (right) the difference between RAW and SM. The model grid spacing Δx is 4.4 (black), 2.2 (blue), and 1.1 (red) km. Gray lines show the observed precipitation (d),(e) from the CombiPrecip dataset, and (g),(h) from 104 Swiss precipitation measurement stations. The labels in the left and middle columns indicate the mean daily accumulated precipitation over the corresponding area or grid points. In all panels, the day-to-day variability is indicated by the semitransparent red sector for the 1.1 km simulations. It indicates the value range between the 15th and 85th percentile, thus roughly representing the spread of seven out of the nine simulated days excluding the day with the smallest and the day with the largest value/difference. Stars in the right column indicate hours when the difference between RAW1 and SM1 is significant ($\alpha = 0.05$).

and RAW2 simulations, the diurnal cycle of precipitation is delayed compared to RAW4 and the SM simulations. The onset time and maximum intensity is delayed, and precipitation does not end at 2200 UTC, as in the RAW4 and the SM simulations. Instead, there is a substantial amount of nighttime precipitation until 0100 and 0400 UTC in RAW2 and RAW1, respectively. The longer duration of precipitation in these

simulations overcompensates for the reduced peak intensity, such that the total accumulated precipitation increases with higher resolution in RAW. Figure 3c shows the difference in the mean diurnal cycle of precipitation between RAW and SM with the semitransparent red sectors indicating the day-to-day variability (15th to 85th percentile) for the 1.1 km simulations. The hours with a significant difference between RAW1

FIG. 4. Mean diurnal cycle of cloud water content (green shading) (g kg^{-1}) as a function of time and altitude (km MSL), averaged over the Alpine region (subdomain A in Fig. 1). Grid points within topography are not considered in the horizontal average. Values below 0.001 g kg^{-1} are masked in the plot. Cloud water is the sum of the specific cloud liquid water content (q_c , shown by the red contour lines) and the specific cloud ice content (q_i , shown by the black contour lines). Contour lines are shown for 0.001, 0.005, and 0.01 g kg^{-1} . Shown are the results of the (top) SM and (bottom) RAW simulations. Results are shown for both sets of simulations with decreasing Δx from (left) 4.4 to (right) 1.1 km.

and SM1 are indicated by a star (based on a two-sided Wilcoxon–Mann–Whitney test with $\alpha = 5\%$). Figure 4, showing the domain-average specific cloud water content (green color shading), also indicates a delayed onset, a reduced maximum intensity, and a longer duration of convective activity in RAW. Cloud water is the sum of cloud liquid water (q_c , red contour lines) and cloud ice (q_i , black contour lines). In both RAW and SM, increasing model resolution results in higher domain average q_i and lower q_c .

In Figs. 3d and 3e, the diurnal cycles of precipitation in RAW1 and SM1 are compared against the observed precipitation from the CombiPrecip dataset. Surface precipitation is averaged over the intersection of the analysis domain A with the radar domain D. The SM simulations show a too early timing of maximum precipitation intensity and decay in the afternoon, insensitive to model resolution. The longer duration of precipitation in the RAW simulations is thus in better agreement with the observations, but the delayed onset is not. While RAW2 captures the timing of maximum intensity relatively well, the diurnal cycle in RAW1 is delayed compared to the observations. Compared to the CombiPrecip dataset, all models clearly underestimate the peak intensity

and accumulated amount of precipitation. However, because the precipitation amount in the CombiPrecip dataset is calibrated only locally using surface observations, the spatial extrapolation could lead to errors in the dataset. The direct comparison of the simulated precipitation with surface measurements (Figs. 3g and 3h) indicates much better agreement between simulated and observed precipitation intensity compared to the evaluation with CombiPrecip.

2) SPATIAL PATTERN

Figures 5 and 6 show the total daytime (0800–2000 UTC) and nighttime (2000–0800 UTC) precipitation, respectively, accumulated over the nine simulated days. The figures show that the spatial pattern of daytime and nighttime precipitation is fundamentally different. Figure 5 reveals no major systematic differences in the spatial pattern of daytime precipitation between RAW and SM, except for a stronger reduction in total accumulated precipitation from low to high model resolution in SM compared to RAW (cf. Fig. 3). Precipitation falls primarily over the mountains, which is what we expect because the mountains favor the formation of deep convection during the day. The highest amounts of

FIG. 5. Total accumulated daytime (between 0800 and 2000 UTC) precipitation (mm) during the nine simulated days over the Alpine region. Values below 10 mm are masked to provide spatial orientation with the underlying elevation map (gray shading). Shown are the results of the (a)–(c) SM and (d)–(f) RAW simulations. The columns show the simulations of each group with decreasing Δx from (left) 4.4 to (right) 1.1 km. Panel (g) shows the observed daytime precipitation in the CombiPrecip dataset.

precipitation fall over the southwestern parts of the Alpine region owing to the larger water vapor content taken up over the Mediterranean Sea. Figure 5g shows observed precipitation in the CombiPrecip dataset, which is only available for the intersection of the Alpine region with domain D. The models capture a precipitation maximum along the main Alpine ridge and the south perimeter but miss a second maximum along the North perimeter. Also, the models miss the precipitation north to the Alps, in particular over the adjacent Jura mountain range. Looking at the spatial pattern of nighttime precipitation in Fig. 6,

we see that virtually all the additional nighttime precipitation in the RAW simulations falls over the southern Alpine flanks and over the Po Valley (see label in Fig. 1). The amount increases systematically with higher resolution in RAW. This is not observed in SM, implying that the nighttime precipitation over these areas in RAW is a consequence of the higher resolved orography. The finding of overall enhanced nighttime precipitation in the RAW2 and RAW1 simulations is supported by observational evidence. However, the observed pattern is less concentrated over the Po Valley as compared to the RAW1 simulation.

FIG. 6. As in Fig. 5, but for accumulated nighttime precipitation (between 2000 and 0800 UTC).

FIG. 7. Histograms of vertical velocity at an altitude of 4 km over the Alpine region (subdomain A in Fig. 1) calculated from hourly velocity snapshots using all model grid points during the nine simulated days. The histograms are computed for a bin width of 0.5 m s^{-1} . Shown are the results of the (a) SM and (b) RAW simulations, and (c) the relative difference in the frequencies between RAW and SM [i.e., $(\text{RAW} - \text{SM})/\text{SM}$]. The model Δx is 4.4 (black), 2.2 (blue), and 1.1 (red) km.

Instead, the observed nighttime precipitation dominates over the Alpine flanks. In particular, a precipitation hot spot is observed to the northwest of the Alps which is absent in the SM simulations and only hinted at in the RAW simulations. Note that due to the chaotic nature of convection, the sample of nine days is relatively small for an evaluation of the precipitation field and these findings need to be considered with care.

3) UPDRAFTS

Histograms of vertical velocities at 4 km altitude—roughly corresponding to the altitude where the convective mass flux peaks—are shown in Fig. 7. Both, the SM (Fig. 7a) and the RAW (Fig. 7b) distributions are approximately symmetric for frequencies $>10^{-3}$, but below, there is a pronounced kink in the distribution for positive velocities (i.e., the distribution is skewed for low frequencies). This skewness is the result of latent heating which enhances positive vertical motions (e.g., Schumacher and Pauluis 2010). The position of the kink in the diagram does not depend on the particular meteorological case considered, but appears to be characteristic for simulations of the European summer climate (e.g., Leutwyler et al. 2017). Figure 7c shows the relative change in frequencies between RAW and SM normalized by SM as a function of vertical velocity. A value of 1 thus means that the frequency of a given velocity in RAW is twice the frequency in SM. Focusing on positive velocities (updrafts), the RAW velocity distribution is characterized by a higher frequency of weak updrafts and lower frequency of strong updrafts compared to SM. Values around $2\text{--}3 \text{ m s}^{-1}$ are twice as frequent in RAW, whereas updrafts around $>10 \text{ m s}^{-1}$ occur considerably less frequently. These differences are of interest and

quite surprising. Intuitively one would expect that the narrower mountains in RAW would force narrower and stronger updrafts, but rather we find that the frequency of strong updrafts is smaller. Narrower updrafts would, however, also be more affected by dilution from entrainment which could have a weakening effect on convection. Even though not shown in Fig. 7, it is remarked here that the results shown are qualitatively the same at all altitudes above 2 km MSL.

For additional insights into how the scale of the updrafts is affected by the resolution of orography, power spectral densities (PSDs) of vertical velocity at 4 km are shown in Fig. 8. The PSDs are shown for all resolutions, but this discussion will focus on the SM1 and RAW1 simulations (Fig. 8c), as the coarser resolved simulations do not (SM4, RAW4) or only partially (SM2, RAW2) resolve the spectral energy peak. The SM1 (black line) and RAW1 (red line) PSDs peak at a wavelength of 6 km—roughly at the effective resolution of the RAW1 and SM1 simulations—indicating that even for this relatively high model resolution, the spectra could still be affected by Δx . The spectral energy peak at wave lengths around 6 km is more pronounced in RAW1 than in SM1, whereas wave lengths $>10 \text{ km}$ show stronger variance in SM1 than in RAW1. These findings indicate a relative shift of energy toward smaller scales in RAW1, forced by the small-scale details in the orography which is in line with the results from the vertical velocity histograms in Fig. 7.

4) CLOUD SIZE

Figure 9 shows the size frequency distributions of all convective clouds that form during the nine simulated days. The term size here refers to the horizontal area

FIG. 8. Normalized power spectral densities (PSDs) of vertical velocity at an altitude of 4 km, averaged over all simulated hours. The PSDs are computed as an average between their latitudinal and longitudinal components over the Alpine region (subdomain A in Fig. 1). The curves are normalized by the total variance so that their integrals equal one. The results are shown for the (a) SM4 and RAW4, (b) SM2 and RAW2, and (c) SM1 and RAW1 simulations.

covered by a coherent cloudy patch. The cloudy patches are separated using an algorithm that finds common grid cell edges according to a threshold value of the liquid water path ($>0.01 \text{ kg m}^{-2}$, with the liquid water path computed between an altitude of 0–10 km). Ice clouds are excluded from the analysis because interest lies on convective clouds rooted in the PBL. The distributions created from the SM and the RAW simulations are shown in Figs. 9a and 9b, respectively. As the minimum size of a cloud is limited by the area of one grid cell, smaller clouds form in the simulations with higher resolution and their number increases. The total number of clouds also increases with higher resolved orography, along with a change in total cloud cover. For the total

number of clouds, the ratios RAW4/SM4, RAW2/SM2 and RAW1/SM1 are 1.07, 1.17, and 1.33, respectively, whereas for the change in cloud cover fraction, the corresponding ratios are 0.90, 0.95, and 1.04. The increase in the number of clouds from SM to RAW is thus systematic across resolutions and largest at 1.1 km. The cloud fraction only increases from RAW1 to SM1 and this increase is smaller in relative terms than the increase in the number of clouds. This implies that the RAW simulations produce smaller clouds on average. For a given size bin, the relative difference in the number of clouds between RAW and SM is shown in Fig. 9c. Small clouds are more abundant in the RAW simulations compared to the SM simulations. Larger clouds on the

FIG. 9. Histograms of cloud size over the Alpine region (subdomain A in Fig. 1) during the nine simulated days calculated from hourly snapshots of cloud structures. The size of the clouds represents the horizontal extent of cloudy patches with a liquid water path greater than 0.01 kg m^{-2} . Shown are the results of the (a) SM and (b) RAW simulations, and (c) the relative difference in the frequencies between RAW and SM [i.e., $(\text{RAW} - \text{SM})/\text{SM}$]. The difference is only shown for bins containing more than 30 clouds in RAW and SM. The model Δx is 4.4 (black), 2.2 (blue), and 1.1 (red) km.

FIG. 10. (a)–(c) Alpine pumping represented by the mean diurnal cycle of the lateral water vapor convergence (mm h^{-1}) within the lowest 2.5 km over the analysis domain A (see Fig. 1). The vertical extent is chosen to represent the average PBL depth over the Alpine region. (d)–(f) Mean diurnal cycle of the surface pressure anomaly averaged over domain A. (g)–(i) Mean diurnal cycle of the PBL temperature anomaly averaged over the lowest 2.5 km of domain A. Shown are the results of the (left) SM and (right) RAW simulations, and (right) the difference between RAW and SM. The grid spacing Δx is 4.4 (black line), 2.2 (blue line), and 1.1 (red line) km. In all panels, the day-to-day variability is indicated by the semitransparent red sector for the 1.1 km simulations. It indicates the value range between the 15th and 85th percentile, thus roughly representing the spread of seven out of the nine simulated days excluding the day with the smallest and the day with the largest value/difference. Stars in the right column indicate hours when the difference between RAW1 and SM1 is significant ($\alpha = 0.05$).

orders of $O(100\text{--}1000) \text{ km}^2$, on the other hand, are less frequent in RAW.

b. Alpine pumping

To understand the evolution of the thermally driven convective activity, the strength of the Alpine pumping is compared between SM and RAW. A key quantity in this context is the lateral convergence of water vapor because it fuels moist convection. Integrating the horizontal water vapor flux ($\rho q_v \mathbf{v}$) pointing into the analysis domain A between the surface and an altitude of 2.5 km MSL (roughly corresponding to the altitude of the

domain-average cloud base) gives the lateral moisture convergence over the Alpine region within the PBL. Figure 10 shows the mean diurnal cycle of moisture convergence in SM (Fig. 10a) and RAW (Fig. 10b) and the difference between RAW and SM (Fig. 10c). It should be noted that the patterns and differences shown in Fig. 10 are dominated by the mass flux ($\rho \mathbf{v}$) and not by the moisture content (q_v). Daytime convergence systematically decreases from SM4 to SM1 which is expected since precipitation intensity is also reduced at higher model resolution (see Fig. 3). More surprising is that daytime convergence and nighttime divergence in

RAW increase with higher resolution—even though this goes along with a further reduction in precipitation intensity. The Alpine pumping is thus substantially enhanced in RAW compared to SM even though daytime precipitation is less intense. This enhancement is closely related to a stronger amplitude of the diurnal surface pressure anomaly (Figs. 10d–f) and goes along with a stronger amplitude of the diurnal PBL mean temperature anomaly (Figs. 10g–i) over the Alpine region. The pressure and temperature anomalies are averaged over the Alpine region after subtracting the spatial and temporal mean in each simulation separately. Like this, average differences between the simulations arising from the different orographies are crossed out. For the PBL mean temperature, a PBL top at an altitude of 2.5 km MSL is assumed.

Since the Alpine pumping is a fundamental process relating to many aspects of fair-weather orographic moist convection (e.g., accumulation of moisture and heat over the mountains, thermally driven wind systems) it is a good quantity to assess the robustness of our results. The semitransparent red sectors in Fig. 10 indicate the day-to-day variability spanning the values between the 15th and 85th percentile, thus roughly representing the spread of seven out of the nine simulated days (excluding the day with the smallest and the day with the largest value). Applying a two-sided Wilcoxon–Mann–Whitney test reveals that the differences are significant ($\alpha = 5\%$) for most hours of the day (marked by the stars in the right panels of Fig. 10).

c. Delayed onset of convection

The first precipitating convective cells form over the mountain tops around 1000 UTC in SM1, roughly 1 h earlier as in RAW1. The reason appears to be related to a higher abundance of near-surface water vapor (q_v) over the mountain peaks in SM1. This is illustrated by the comparison of Figs. 11e,f, showing a latitude–altitude q_v cross section across the Alpine arc (line C Fig. 1) at 0900 UTC. In RAW1, the moisture is less concentrated at the surface and more abundant at higher levels (>3 km altitude). When the first convective cells are initiated over the mountain tops in SM1, they can be expected to be grow stronger than in RAW1 due to the more abundant q_v at the surface (cloud structures are shown by the purple contour lines in Fig. 11). The differences in the q_v pattern likely relate to the degree of detail in the represented orography, but it turns out difficult to do a comprehensive analysis and to provide a definitive conclusion. Still, we will in the following outline two mechanisms that may contribute to the differences in the observed q_v pattern between SM1 and RAW1.

First, the meridional profile of the SM1 Alpine range, as shown in Fig. 11, reflects one single mesoscale slope (resembling an idealized Gaussian mountain) whereas the RAW1 profile has several superimposed small-scale mountain peaks and valleys. The SM1 Alpine range allows for near-surface q_v -loaded air to glide up its smooth slope while remaining close to the surface until it converges at the central ridge of the Alpine arc. The additional orographic structures in RAW1 1) force the near-surface converging flow to follow along their deviations in the vertical and 2) generate their own thermally driven circulations toward their associated small-scale mountain peaks which can interfere with the mesoscale converging flow (as illustrated, e.g., in Fig. 11f by the elevated patch of high q_v above the small ridge North of the Alps at $y = 520$ km). The latter was also found in Lang et al. (2015) who studied the effect of a superimposed valley on an extended mountain slope in an idealized setup. The result of both effects 1) and 2) is to vertically redistribute q_v away from lower to higher altitudes in RAW1 while it converges toward the center of the Alpine ridge resulting in a slower accumulation of near-surface q_v on top of the Alpine arc. This effect is also illustrated by Figs. 12a and 12b showing a latitude–altitude cross section of the 9-day zonal average meridional water vapor flux ($\rho q_v \mathbf{v}$) during the morning hours (0800–1200 UTC) over the Po Valley within domain B (see Fig. 2). The converging flow toward the Alps (red/blue patch south/north of the Alps) is concentrated on a narrower vertical band around the mean surface elevation (solid thick black line) in SM1 (Fig. 12a), compared to RAW1 (Fig. 12b) where it is vertically dispersed downward into the mountain valleys and upward into the lower troposphere.

The second mechanism, although difficult to quantify, is based on the idea that cumulus detrainment of water vapor (see, e.g., Kirshbaum 2011) may be more important in RAW1 because the smaller cloud structures (see Fig. 9) have a higher surface-to-volume ratio. More moisture can thus be expected to be mixed into the free troposphere slowing down the initiation of deep convection in the morning.

d. Secondary nighttime precipitation

Virtually all the additional precipitation in the RAW simulations falls during the evening and night over the Po Valley and the southern Alpine slope (see Fig. 6). The responsible cells appear to be triggered by evening/nighttime cold-air outflow from the Alpine valleys into the Po Valley. This process is not unique to the RAW simulations but the resulting moist-convective activity lasts longer than in SM.

FIG. 11. Snapshots of specific water vapor content (q_v) (g kg^{-1}) and wind arrows on 12 Jul 2006, shown as latitude–altitude cross sections on a meridional cut through the Alpine arc (line C in Fig. 1). Shown are the results of the (left) SM1 and (right) RAW1 simulations, and the rows show the results during the morning hours at 0500, 0700, 0900, and 1100 UTC. The wind arrows are normalized to the white wind arrow shown in (a). The thin black lines show the contours of $q_v = 8 \text{ g kg}^{-1}$ to highlight locations of high q_v . The thick purple contour lines indicate cloudy patches (specific cloud liquid water content $> 1 \text{ g kg}^{-1}$).

Figure 13 compares the vertical integral of water vapor between RAW1 and SM1 at 2000 UTC—shortly before the onset of the evening and nighttime convection in RAW1. The signal is separated into the contribution from within the PBL (between the surface and 2 km altitude; Figs. 13a and 13b) and above the PBL (between 2 and 10 km altitude; Figs. 13c and 13d). The Po Valley PBL is more humid in RAW1 than in SM1. When triggered, this can help convective cells to become more vigorous. Also well visible in Fig. 13b, the better-resolved mountain valleys in RAW1 contain lower-lying air masses with higher temperature and

higher density that can store additional water vapor. The contribution from above the PBL shows a substantially moister free troposphere in RAW1 than in SM1 which can also be favorable for the development of deep convection by reducing the drying caused by clear-air entrainment in existing convective cells. The enhanced moistening of the free troposphere in RAW1 happens over the Alpine arc, as indicated in Fig. 13d. Given a weak synoptic on average northeasterly background flow, this signal is partially advected over the southern Alpine flanks and the Po Valley by 2000 UTC.

0800–1200 UTC

1600–1800 UTC

2200–0600 UTC

FIG. 12. Latitude–altitude cross sections of the meridional water vapor flux ($\rho q_v \mathbf{v}$) ($\text{g m}^{-2} \text{s}^{-1}$). The meridional water vapor flux is averaged over the nine days during the indicated hours: (a)–(c) 0800–1200 UTC, (d)–(f) 1600–1800 UTC, and (g)–(i) 2200–0600 UTC. The values are obtained by averaging in the zonal direction of subdomain B (see Fig. 1) and considering only grid cells that are not within orography. Shown are the results from the (left) SM1 and (right) RAW1 simulations. The solid black lines indicate the mean surface elevation averaged in the zonal direction whereas the black filled areas show the minimum elevation in the zonal direction. It lies lower in RAW1 because the mountain valleys are deeper. The water vapor flux between the minimum and the mean elevation can be considered the transport within the mountain valleys. (right) The difference in the vapor flux between RAW1 and SM1. The thick (thin) solid black line indicates the mean (minimum) surface elevation of the SM1 simulation in the zonal direction. The filled black shape indicates the minimum surface elevation of the RAW1 simulation. Note that for grid cells located below the SM1 minimum elevation, the difference between RAW1 and SM1 equals the value of RAW1 because the values of SM1 equal zero (within topography).

FIG. 13. Mean vertically integrated water vapor (kg m^{-2}) at 2000 UTC averaged over the nine simulated days and integrated (a),(b) between the surface and a slab at 2 km and (c),(d) between slabs at 2 and 10 km. Shown are the results of the (left) SM1 and (right) RAW1 simulations.

The more humid Po Valley PBL in RAW1, found in Fig. 13b results from stronger lateral moisture convergence during the late afternoon and early evening. The diurnal cycle of lateral moisture convergence into the PBL of the Po Valley (domain E in Fig. 1) is shown in Figs. 14a–c. The Po Valley PBL is dried during the morning and moistened during afternoon in both SM and RAW. However, the moistening during the late afternoon lasts longer in RAW (Fig. 14c). Figures 14d–f show, for the 1.1 km simulations, the contributions to the net convergence of the separate moisture fluxes pointing into the Po Valley from the different spatial directions. The longer-lasting moisture convergence in RAW1 during the evening results from stronger fluxes into (or weaker fluxes out of) the domain from South and from East (Fig. 14f).

This finding is illustrated in Figs. 12d,e showing the meridional water vapor flux ($\rho q_v \mathbf{v}$) between 1600 and 1800 UTC. The northward branch of the plain-to-mountain circulation (red banners at $y = 250\text{--}350$ km) advects moist air from the Ligurian Sea (see Fig. 1) toward the Alps (qualitatively the same happens in the

zonal direction). This northward moisture flux is stronger in RAW1 than in SM1, as shown in Fig. 14—likely driven by the enhanced low pressure anomaly in RAW over the Alpine region (see Fig. 10). It could be that the orographic representation of the coastal mountain range additionally favors the development of a stronger northward flux since the minimum elevation of the mountain range is lower in RAW1 than in SM1 (cf. orography for $y = 260$ km between Figs. 12d and 12e). The northward moisture flux converging toward the Alpine range collides with a southward moisture flux propagating out of the mountain valleys into the Po Valley (blue banners close to the surface at $y = 350\text{--}400$ km in Figs. 12d and 12e). The southward flux represents cold-air outflow from existing convective cells over the Alpine range. It forces the relatively warm air from the Mediterranean to rise. This process may be responsible for the triggering of the first deep-convective cells over the Po Valley during late afternoon.

To complete the picture, Figs. 12g and 12h show the situation during nighttime (averaged between 2200 and

FIG. 14. (a)–(c) Mean diurnal cycle of the lateral water vapor convergence (mm h^{-1}) within the lowest 2 km of the Po Valley domain E (see Fig. 1). The vertical extent is chosen to represent the average PBL depth over the Po Valley. Shown are the results of the (left) SM and (middle) RAW simulations, and (right) the difference between RAW and SM. The grid spacing Δx is 4.4 (black line), 2.2 (blue line), and 1.1 (red line) km. The day-to-day variability is indicated by the semitransparent red sector for the 1.1 km simulations. It indicates the value range between the 15th and 85th percentile, thus roughly representing the spread of seven out of the nine simulated days excluding the day with the smallest and the day with the largest value/difference. Stars in the right column indicate hours when the difference between RAW1 and SM1 is significant ($\alpha = 0.05$). (d)–(f) Contribution of the fluxes entering the Po Valley domain from the four directions (N: North, S: South, W: West, E: East) to the net moisture convergence shown in (a)–(c). The results are shown exclusively for the 1.1 km simulations.

0600 UTC). The southward moisture transport is enhanced in RAW1 compared to SM1, reflecting the stronger nighttime moisture divergence found for RAW in Figs. 10 and 14.

Summarizing, we propose an interplay of two effects to drive the stronger evening and nighttime convection over the Po Valley and the southern Alpine flanks in RAW1, as compared to the SM simulations and the coarser resolved RAW simulations: The enhanced low-level moisture convergence over the Po Valley during the late afternoon and evening supplies moisture and sustains the convective cells while the more humid free troposphere may additionally favor deep convection by reducing the drying of the clouds by clear-air entrainment.

4. Discussion and conclusions

The aim of this study is to assess the impact of high-resolution orography on kilometer-scale simulations of orographic moist convection. To this end, the COSMO model is used to simulate nine days of reoccurring thermally driven convection over the Alps. Two sets of

simulations are compared, each consisting of three runs at a grid spacing of $\Delta x = 4.4, 2.2$, and 1.1 km: one with fixed low-resolution orography (SM), and one where the representation of orography is refined together with Δx (RAW).

A higher resolution in the orography causes a later onset of surface precipitation with reduced peak intensity and longer duration. The total accumulated precipitation increases with the additional amounts falling during the evening and night over the Po Valley and the southern Alpine slopes. Some features specific to the RAW simulations can be seen in observations like for instance the longer duration of precipitation during the evening and night, whereas the delayed onset in the morning is not in agreement with observations. With higher orographic resolution, a stronger low pressure anomaly over the mountains builds up during the day. Accordingly, the mesoscale plain-to-mountain flow (Alpine pumping) is stronger, resulting in stronger moisture convergence over the Alps during the day. At the same time, afternoon precipitation intensity is reduced resulting in an accumulation of water vapor

such that the atmosphere contains substantially more water vapor in the late afternoon. This fosters convective activity during the evening and night. Our findings thus show that the overall simulated convective activity is enhanced with higher resolved orography, not because it intensifies but because its duration is prolonged. The intensified Alpine pumping provides the required additional moisture. The formation of precipitating convective cells on top of the Alpine arc in the morning is delayed with higher detail in the orography. We did not expect this as previous studies (Panosetti et al. 2016; Imamovic et al. 2019) found an earlier onset of convection for high mountains (RAW) compared to low mountains (SM). We relate the delay to the deeper valleys and higher peaks which delay the convergence of moisture at the center of the Alpine arc. Thus, it is rather the dynamical impact of the additional detail in the orography instead of the higher-elevated mountain peaks controlling the onset time of convection in our simulations. Finally, the convective cells become weaker, the clouds become smaller and more numerous along with the larger number of resolved mountain peaks. This is an indication that the shallow convective regime is enhanced over higher resolved orography at the cost of the deep convective regime.

For the domain-averaged variables, the difference between RAW1 and SM1 is larger than the difference between SM1 and SM4. Thus, the effect of the topographic resolution on the simulated domain-averaged convective activity is more important than the effect of the model grid spacing. Looking at individual convective cells, small-scale details in the orography impose finer scales on the prognostic fields, but at these spatial scales, the effects caused by orography are smaller than the ones caused by the present increments in model grid spacing. The radiation scheme used in this study does not account for topographic heterogeneity. Observational studies (e.g., Whiteman et al. 1989b,a) and modeling studies (Hoch et al. 2011) show that the latter can largely influence the thermal structure and associated circulations of mountain valleys. Using a radiation scheme that considers topographic effects (such as sky-view factors) would likely further pronounce the differences between RAW and SM.

The differences due to changed orography increase with model resolution for all analyzed variables and are thus systematic. Given this and the robustness of the signal across the nine simulated days, we are confident that the small sample considered is large enough to be representative of the specific weather situation considered. It can, however, be expected that a different local forcing (e.g., soil moisture) or large-scale forcing (e.g., different subsidence strength) could lead to different

results, even for days with a similar synoptic situation. In addition, in synoptic situations with a stronger background flow, additional orographic forcing through mesoscale lifting is expected and will change the triggering of deep-convective cells (see, e.g., Kirshbaum et al. 2018). Finally, the closeness of the Po Valley to the Ligurian Sea and the fact that it is separated by a coastal mountain range (the Ligurian Alps) raises the question of how much the moisture flux from the Ligurian Sea into the Po Valley is affected by the sea breeze which itself may be modulated by the representation of the coastal orography.

It would be valuable to investigate how the changes in the valley geometry caused by the higher topographic resolution affect the local along- and cross-valley circulations. The cross-valley circulation has been found to intensify within steeper and narrower valleys (Wagner et al. 2015a,b) up to the formation of two stacked circulation cells (Serafin and Zardi 2010; Schmidli 2013; Wagner et al. 2014), which could well be the case in our RAW1 simulation. According to the discussion by Serafin and Zardi (2010, 2011) and Schmidli (2013), which is nicely summarized in Rotach et al. (2015), one would expect a net cooling effect from a stronger cross-valley circulation in steeper valleys. However, we find an enhanced daytime temperature anomaly in the RAW simulations. This could result from the stronger daytime convergence of the mesoscale flow. Besides that, the valley volume effect may play a role in enhancing the daytime temperature anomaly in RAW. However, the mountain volume in RAW and SM is identical and thus also the heated air volume. It could nevertheless be that the deeper and narrower RAW valleys at low levels experience a valley volume effect compared to the relatively “flat” Alpine topography in SM. Contradicting this argumentation as well as our finding of increased daytime temperature in RAW, Wagner et al. (2014) found increased temperature in a smoothed valley rather than in a well resolved deep valley. They attributed their finding to the smaller air volume that needs to be heated during the day in the smoothed valley. Another explanation for the temperature difference between RAW and SM is a different surface sensible heat flux, as pointed out by de Wekker et al. (1998). The valley and slope winds are stronger in RAW which could thus enhance the warming during the day. Further investigations pointing in these directions would be of interest, since they may link the higher resolved orography to the stronger Alpine pumping.

Concluding, this study showed that a more detailed representation of the orography in kilometer-scale simulations substantially affects the simulation of thermally driven orographic moist convection. In real-world simulations with a similar setup, the changes found here

can be expected to occur in a similar way. This should thus be kept in mind when increasing the resolution in a numerical weather prediction or climate model.

Acknowledgments. We want to thank two anonymous reviewers for their careful and constructive feedback that helped a lot improving the manuscript. Partial funding for this study has been provided by the Swiss National Science Foundation through Sinergia Grant CRSII2 154486/1 crCLIM. We also acknowledge PRACE for awarding us access to Piz Daint at CSCS, Switzerland. The authors declare no conflict of interest.

REFERENCES

- Adler, B., and N. Kalthoff, 2014: Multi-scale transport processes observed in mountain layer over a mountainous island. *Bound.-Layer Meteor.*, **153**, 515–537, <https://doi.org/10.1007/s10546-014-9957-8>.
- Baldauf, M., and G. Zängl, 2012: Horizontal nonlinear Smagorinsky diffusion. *COSMO Newsletter*, No. 2, Consortium for Small-Scale Modeling, Offenbach, Germany, 3–7, <http://cosmo-model.cscs.ch/content/model/documentation/newsLetters/newsLetter12/1-baldauf.pdf>.
- , A. Seifert, J. Förstner, D. Majewski, M. Raschendorfer, and T. Reinhardt, 2011: Operational convective-scale numerical weather prediction with the COSMO model: Description and sensitivities. *Mon. Wea. Rev.*, **139**, 3887–3905, <https://doi.org/10.1175/MWR-D-10-05013.1>.
- Ban, N., J. Schmidli, and C. Schär, 2014: Evaluation of the convection-resolving regional climate modeling approach in decade-long simulations. *J. Geophys. Res. Atmos.*, **119**, 7889–7907, <https://doi.org/10.1002/2014JD021478>.
- Barthlott, C., B. Adler, N. Kalthoff, J. Handwerker, M. Kohler, and A. Wieser, 2016: The role of Corsica in initiating nocturnal offshore convection. *Quart. J. Roy. Meteor. Soc.*, **142**, 222–237, <https://doi.org/10.1002/qj.2415>.
- Bott, A., 1989: A positive definite advection scheme obtained by nonlinear renormalization of the advective fluxes. *Mon. Wea. Rev.*, **117**, 1006–1016, [https://doi.org/10.1175/1520-0493\(1989\)117<1006:APDASO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1989)117<1006:APDASO>2.0.CO;2).
- Bretherton, C. S., and M. F. Khairoutdinov, 2015: Convective self-aggregation feedbacks in near-global cloud-resolving simulations of an aquaplanet. *J. Adv. Model. Earth Syst.*, **7**, 1765–1787, <https://doi.org/10.1002/2015MS000499>.
- Cosma, S., E. Richard, and F. Miniscloux, 2002: The role of small-scale orographic features in the spatial distribution of precipitation. *Quart. J. Roy. Meteor. Soc.*, **128**, 75–92, <https://doi.org/10.1256/00359000260498798>.
- Crook, N. A., 1996: Sensitivity of moist convection forced by boundary layer processes to low-level thermodynamic fields. *Mon. Wea. Rev.*, **124**, 1767–1785, [https://doi.org/10.1175/1520-0493\(1996\)124<1767:SOMCFB>2.0.CO;2](https://doi.org/10.1175/1520-0493(1996)124<1767:SOMCFB>2.0.CO;2).
- de Wekker, S. F. J., S. Zhong, J. D. Fast, and C. D. Whiteman, 1998: A numerical study of the thermally driven plain-to-basin wind over idealized basin topographies. *J. Appl. Meteor.*, **37**, 606–622, [https://doi.org/10.1175/1520-0450\(1998\)037<0606:ANSOTT>2.0.CO;2](https://doi.org/10.1175/1520-0450(1998)037<0606:ANSOTT>2.0.CO;2).
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Demko, J. C., and B. Geerts, 2010: A numerical study of the evolving convective boundary layer and orographic circulation around the Santa Catalina Mountains in Arizona. Part II: Interaction with deep convection. *Mon. Wea. Rev.*, **138**, 3603–3622, <https://doi.org/10.1175/2010MWR3318.1>.
- Egger, J., 1990: Thermally forced flows: Theory. *Atmospheric Processes over Complex Terrain*, W. Blumen, Ed., Amer. Meteor. Soc., 43–58.
- Frei, C., and C. Schär, 1998: A precipitation climatology of the Alps from high-resolution rain-gauge observations. *Int. J. Climatol.*, **18**, 873–900, [https://doi.org/10.1002/\(SICI\)1097-0088\(19980630\)18:8<873::AID-JOC255>3.0.CO;2-9](https://doi.org/10.1002/(SICI)1097-0088(19980630)18:8<873::AID-JOC255>3.0.CO;2-9).
- Fuhrer, O., and C. Schär, 2007: Dynamics of orographically triggered banded convection in sheared moist orographic flows. *J. Atmos. Sci.*, **64**, 3542–3561, <https://doi.org/10.1175/JAS4024.1>.
- , and Coauthors, 2018: Near-global climate simulation at 1 km resolution: Establishing a performance baseline on 4888 GPUs with COSMO 5.0. *Geosci. Model Dev.*, **11**, 1665–1681, <https://doi.org/10.5194/gmd-11-1665-2018>.
- Hastings, D. A., and Coauthors, 1999: The Global Land One-kilometer Base Elevation (GLOBE) Digital Elevation Model, version 1.0. National Oceanic and Atmospheric Administration, 147 pp., <https://www.ngdc.noaa.gov/mgg/topo/report/globedocumentationmanual.pdf>.
- Heise, E., M. Lange, B. Ritter, and R. Schrodin, 2003: Improvement and validation of the multi-layer soil model. *COSMO Newsletter*, No. 3, Consortium for Small-Scale Modeling, Offenbach, Germany, 198–203, <http://www.cosmo-model.org/content/model/documentation/newsLetters/newsLetter03/cnl3-chp9-12.pdf>.
- Hentgen, L., N. Ban, N. Kröner, D. Leutwyler, and C. Schär, 2019: Clouds in convection-resolving climate simulations over Europe. *J. Geophys. Res. Atmos.*, **124**, 3849–3870, <https://doi.org/10.1029/2018JD030150>.
- Hoch, S. W., C. D. Whiteman, and B. Mayer, 2011: A systematic study of longwave radiative heating and cooling within valleys and basins using a three-dimensional radiative transfer model. *J. Appl. Meteor. Climatol.*, **50**, 2473–2489, <https://doi.org/10.1175/JAMC-D-11-083.1>.
- Hogan, R. J., P. R. Field, A. J. Illingworth, R. J. Cotton, and T. W. Choullarton, 2002: Properties of embedded convection in warm-frontal mixed-phase cloud from aircraft and polarimetric radar. *Quart. J. Roy. Meteor. Soc.*, **128**, 451–476, <https://doi.org/10.1256/003590002321042054>.
- Hohenegger, C., and C. Schär, 2007: Atmospheric predictability at synoptic versus cloud-resolving scales. *Bull. Amer. Meteor. Soc.*, **88**, 1783–1794, <https://doi.org/10.1175/BAMS-88-11-1783>.
- , P. Brockhaus, and C. Schär, 2008: Towards climate simulations at cloud-resolving scales. *Meteor. Z.*, **17**, 383–394, <https://doi.org/10.1127/0941-2948/2008/0303>.
- , —, C. S. Bretherton, and C. Schär, 2009: The soil moisture–precipitation feedback in simulations with explicit and parameterized convection. *J. Climate*, **22**, 5003–5020, <https://doi.org/10.1175/2009JCLI2604.1>.
- Imamovic, A., L. Schlemmer, and C. Schär, 2019: Mountain volume control on deep-convective rain amount during episodes of weak synoptic forcing. *J. Atmos. Sci.*, **76**, 605–626, <https://doi.org/10.1175/JAS-D-18-0217.1>.
- Jaeger, E. B., I. Anders, D. Lüthi, B. Rockel, C. Schär, and S. I. Senéviratne, 2008: Analysis of ERA40-driven CLM simulations

- for Europe. *Meteor. Z.*, **17**, 349–367, <https://doi.org/10.1127/0941-2948/2008/0301>.
- Juckes, M., and R. K. Smith, 2000: Convective destabilization by upper-level troughs. *Quart. J. Roy. Meteor. Soc.*, **126**, 111–123, <https://doi.org/10.1002/qj.49712656206>.
- Kendon, E. J., N. M. Roberts, H. J. Fowler, M. J. Roberts, S. C. Chan, and C. A. Senior, 2014: Heavier summer downpours with climate change revealed by weather forecast resolution model. *Nat. Climate Change*, **4**, 570–576, <https://doi.org/10.1038/nclimate2258>.
- Kirshbaum, D. J., 2011: Cloud-resolving simulations of deep convection over a heated mountain. *J. Atmos. Sci.*, **68**, 361–378, <https://doi.org/10.1175/2010JAS3642.1>.
- , and A. L. M. Grant, 2012: Invigoration of cumulus cloud fields by mesoscale ascent. *Quart. J. Roy. Meteor. Soc.*, **138**, 2136–2150, <https://doi.org/10.1002/qj.1954>.
- , G. H. Bryan, R. Rotunno, and D. R. Durran, 2007: The triggering of orographic rainbands by small-scale topography. *J. Atmos. Sci.*, **64**, 1530–1549, <https://doi.org/10.1175/JAS3924.1>.
- , B. Adler, N. Kalthoff, C. Barthlott, and S. Serafin, 2018: Moist orographic convection: Physical mechanisms and links to surface-exchange processes. *Atmosphere*, **9**, 80, <https://doi.org/10.3390/atmos9030080>.
- Klemp, J. B., and R. B. Wilhelmson, 1978: The simulation of three-dimensional convective storm dynamics. *J. Atmos. Sci.*, **35**, 1070–1096, [https://doi.org/10.1175/1520-0469\(1978\)035<1070:TSOTDC>2.0.CO;2](https://doi.org/10.1175/1520-0469(1978)035<1070:TSOTDC>2.0.CO;2).
- Kuo, J.-T., and H. D. Orville, 1973: A radar climatology of summertime convective clouds in the Black Hills. *J. Appl. Meteor.*, **12**, 359–368, [https://doi.org/10.1175/1520-0450\(1973\)012<0359:ARCOSC>2.0.CO;2](https://doi.org/10.1175/1520-0450(1973)012<0359:ARCOSC>2.0.CO;2).
- Lang, M. N., A. Gohm, and J. S. Wagner, 2015: The impact of embedded valleys on daytime pollution transport over a mountain range. *Atmos. Chem. Phys.*, **15**, 11 981–11 998, <https://doi.org/10.5194/acp-15-11981-2015>.
- Langhans, W., J. Schmidli, and C. Schär, 2012: Bulk convergence of cloud-resolving simulations of moist convection over complex terrain. *J. Atmos. Sci.*, **69**, 2207–2228, <https://doi.org/10.1175/JAS-D-11-0252.1>.
- , —, O. Fuhrer, S. Bieri, and C. Schär, 2013: Long-term simulations of thermally driven flows and orographic convection at convection-parameterizing and cloud-resolving resolutions. *J. Appl. Meteor. Climatol.*, **52**, 1490–1510, <https://doi.org/10.1175/JAMC-D-12-0167.1>.
- Lean, H. W., P. A. Clark, M. Dixon, N. M. Roberts, A. Fitch, R. Forbes, and C. Halliwell, 2008: Characteristics of high-resolution versions of the Met Office Unified Model for forecasting convection over the United Kingdom. *Mon. Wea. Rev.*, **136**, 3408–3424, <https://doi.org/10.1175/2008MWR2332.1>.
- Leutwyler, D., O. Fuhrer, X. Lapillonne, D. Lüthi, and C. Schär, 2016: Towards European-scale convection-resolving climate simulations with GPUs: A study with COSMO 4.19. *Geosci. Model Dev.*, **9**, 3393–3412, <https://doi.org/10.5194/gmd-9-3393-2016>.
- , D. Lüthi, N. Ban, O. Fuhrer, and C. Schär, 2017: Evaluation of the convection-resolving climate modeling approach on continental scales. *J. Geophys. Res. Atmos.*, **122**, 5237–5258, <https://doi.org/10.1002/2016JD026013>.
- Lugauer, M., and P. Winkler, 2005: Thermal circulation in South Bavaria climatology and synoptic aspects. *Meteor. Z.*, **14**, 15–30, <https://doi.org/10.1127/0941-2948/2005/0014-0015>.
- Mazon, J., and D. Pino, 2013: Nocturnal offshore precipitation near the Mediterranean coast of the Iberian Peninsula. *Meteor. Atmos. Phys.*, **120**, 11–28, <https://doi.org/10.1007/s00703-012-0229-1>.
- Miyamoto, Y., Y. Kajikawa, R. Yoshida, T. Yamaura, H. Yashiro, and H. Tomita, 2013: Deep moist atmospheric convection in a subkilometer global simulation. *Geophys. Res. Lett.*, **40**, 4922–4926, <https://doi.org/10.1002/grl.50944>.
- Neumann, P., and Coauthors, 2019: Assessing the scales in numerical weather and climate predictions: Will exascale be the rescue? *Philos. Trans. Roy. Soc.*, **377A**, 20180148, <https://doi.org/10.1098/rsta.2018.0148>.
- Panosetti, D., S. Böing, L. Schlemmer, and J. Schmidli, 2016: Idealized large-eddy and convection-resolving simulations of moist convection over mountainous terrain. *J. Atmos. Sci.*, **73**, 4021–4041, <https://doi.org/10.1175/JAS-D-15-0341.1>.
- , L. Schlemmer, and C. Schär, 2019: Bulk and structural convergence at convection-resolving scales in real-case simulations of summertime moist convection over land. *Quart. J. Roy. Meteor. Soc.*, **145**, 1427–1443, <https://doi.org/10.1002/qj.3502>.
- Picard, L., and C. Mass, 2017: The sensitivity of orographic precipitation to flow direction: An idealized modeling approach. *J. Hydrometeorol.*, **18**, 1673–1688, <https://doi.org/10.1175/JHM-D-16-0209.1>.
- Prein, A. F., A. Gobiet, M. Suklitsch, H. Truhetz, N. K. Awan, K. Keuler, and G. Georgievski, 2013: Added value of convection permitting seasonal simulations. *Climate Dyn.*, **41**, 2655–2677, <https://doi.org/10.1007/s00382-013-1744-6>.
- , and Coauthors, 2015: A review on regional convection-permitting climate modeling: Demonstrations, prospects, and challenges. *Rev. Geophys.*, **53**, 323–361, <https://doi.org/10.1002/2014RG000475>.
- , R. M. Rasmussen, K. Ikeda, C. Liu, M. P. Clark, and G. J. Holland, 2017: The future intensification of hourly precipitation extremes. *Nat. Climate Change*, **7**, 48–52, <https://doi.org/10.1038/nclimate3168>.
- Raschendorfer, M., 2001: The new turbulence parameterization of LM. *COSMO Newsletter*, No. 1, Consortium for Small-Scale Modeling, Offenbach, Germany, 89–97, http://www.cosmo-model.org/content/model/documentation/newsLetters/newsLetter01/newsLetter_01.pdf.
- Rasp, S., T. Selz, and G. C. Craig, 2016: Convective and slantwise trajectory ascent in convection-permitting simulations of midlatitude cyclones. *Mon. Wea. Rev.*, **144**, 3961–3976, <https://doi.org/10.1175/MWR-D-16-0112.1>.
- Raymond, W. H., 1988: High-order low-pass implicit tangent filters for use in finite area calculations. *Mon. Wea. Rev.*, **116**, 2132–2141, [https://doi.org/10.1175/1520-0493\(1988\)116<2132:HOLPIT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1988)116<2132:HOLPIT>2.0.CO;2).
- Reinhardt, T., and A. Seifert, 2006: A three-category ice scheme for LMK. *COSMO Newsletter*, No. 6, Consortium for Small-Scale Modeling, Offenbach, Germany, 115–120, http://www.cosmo-model.org/content/model/documentation/newsLetters/newsLetter06/cnl6_reinhardt.pdf.
- Ritter, B., and J.-F. Geleyn, 1992: A comprehensive radiation scheme for numerical weather prediction models with potential applications in climate simulations. *Mon. Wea. Rev.*, **120**, 303–325, [https://doi.org/10.1175/1520-0493\(1992\)120<0303:ACRSFN>2.0.CO;2](https://doi.org/10.1175/1520-0493(1992)120<0303:ACRSFN>2.0.CO;2).
- Rockel, B., A. Will, and A. Hense, 2008: The regional climate model COSMO-CLM (CCLM). *Meteor. Z.*, **17**, 347–348, <https://doi.org/10.1127/0941-2948/2008/0309>.
- Rotach, M. W., A. Gohm, M. N. Lang, D. Leukauf, I. Stiperski, and J. S. Wagner, 2015: On the vertical exchange of heat, mass, and

- momentum over complex, mountainous terrain. *Front. Earth Sci.*, **3**, 1–14, <https://doi.org/10.3389/FEART.2015.00076>.
- Sato, T., and F. Kimura, 2005: Diurnal cycle of convective instability around the central mountains in Japan during the warm season. *J. Atmos. Sci.*, **62**, 1626–1636, <https://doi.org/10.1175/JAS3423.1>.
- Schlemmer, L., and C. Hohenegger, 2014: The formation of wider and deeper clouds as a result of cold-pool dynamics. *J. Atmos. Sci.*, **71**, 2842–2858, <https://doi.org/10.1175/JAS-D-13-0170.1>.
- , and —, 2016: Modifications of the atmospheric moisture field as a result of cold-pool dynamics. *Quart. J. Roy. Meteor. Soc.*, **142**, 30–42, <https://doi.org/10.1002/qj.2625>.
- Schmidli, J., 2013: Daytime heat transfer processes over mountainous terrain. *J. Atmos. Sci.*, **70**, 4041–4066, <https://doi.org/10.1175/JAS-D-13-083.1>.
- , S. Böing, and O. Fuhrer, 2018: Accuracy of simulated diurnal valley winds in the Swiss Alps: Influence of grid resolution, topography filtering, and land surface datasets. *Atmosphere*, **9**, 196, <https://doi.org/10.3390/atmos9050196>.
- Schumacher, J., and O. Pauluis, 2010: Buoyancy statistics in moist turbulent Rayleigh–Bénard convection. *J. Fluid Mech.*, **648**, 509–519, <https://doi.org/10.1017/S0022112010000030>.
- Schumacher, R. S., D. M. Schultz, and J. A. Knox, 2015: Influence of terrain resolution on banded convection in the lee of the Rocky Mountains. *Mon. Wea. Rev.*, **143**, 1399–1416, <https://doi.org/10.1175/MWR-D-14-00255.1>.
- Serafin, S., and D. Zardi, 2010: Daytime heat transfer processes related to slope flows and turbulent convection in an idealized mountain valley. *J. Atmos. Sci.*, **67**, 3739–3756, <https://doi.org/10.1175/2010JAS3428.1>.
- , and —, 2011: Daytime development of the boundary layer over a plain and in a valley under fair weather conditions: A comparison by means of idealized numerical simulations. *J. Atmos. Sci.*, **68**, 2128–2141, <https://doi.org/10.1175/2011JAS3610.1>.
- Sideris, I. V., M. Gabella, R. Erdin, and U. Germann, 2014: Real-time radar-rain-gauge merging using spatio-temporal co-kriging with external drift in the alpine terrain of Switzerland. *Quart. J. Roy. Meteor. Soc.*, **140**, 1097–1111, <https://doi.org/10.1002/qj.2188>.
- Skamarock, W. C., 2004: Evaluating mesoscale NWP models using kinetic energy spectra. *Mon. Wea. Rev.*, **132**, 3019–3032, <https://doi.org/10.1175/MWR2830.1>.
- Steinacker, R., 1984: Area–height distribution of a valley and its relation to the valley wind. *Contrib. Atmos. Phys.*, **57**, 64–71.
- Trier, S. B., F. Chen, and K. W. Manning, 2004: A study of convection initiation in a mesoscale model using high-resolution land surface initial conditions. *Mon. Wea. Rev.*, **132**, 2954–2976, <https://doi.org/10.1175/MWR2839.1>.
- Tselioudis, G., C. Douvis, and C. Zerefos, 2012: Does dynamical downscaling introduce novel information in climate model simulations of precipitation change over a complex topography region? *Int. J. Climatol.*, **32**, 1572–1578, <https://doi.org/10.1002/joc.2360>.
- Van Baelen, J., M. Reverdy, F. Tridon, L. Labbouz, G. Dick, M. Bender, and M. Hagen, 2011: On the relationship between water vapour field evolution and the life cycle of precipitation systems. *Quart. J. Roy. Meteor. Soc.*, **137**, 204–223, <https://doi.org/10.1002/qj.785>.
- Wagner, A., 1932: Neue Theorie des Berg- und Talwindes (A new theory of mountain and valley winds). *Meteor. Z.*, **49**, 329–341.
- Wagner, J. S., A. Gohm, and M. W. Rotach, 2014: The impact of horizontal model grid resolution on the boundary layer structure over an idealized valley. *Mon. Wea. Rev.*, **142**, 3446–3465, <https://doi.org/10.1175/MWR-D-14-00002.1>.
- , —, and —, 2015a: Influence of along-valley terrain heterogeneity on exchange processes over idealized valleys. *Atmos. Chem. Phys.*, **15**, 6589–6603, <https://doi.org/10.5194/acp-15-6589-2015>.
- , —, and —, 2015b: The impact of valley geometry on daytime thermally driven flows and vertical transport processes. *Quart. J. Roy. Meteor. Soc.*, **141**, 1780–1794, <https://doi.org/10.1002/qj.2481>.
- Weckwerth, T. M., J. W. Wilson, M. Hagen, T. J. Emerson, J. O. Pinto, D. L. Rife, and L. Grebe, 2011: Radar climatology of the COPS region. *Quart. J. Roy. Meteor. Soc.*, **137**, 31–41, <https://doi.org/10.1002/qj.747>.
- Weusthoff, T., F. Ament, M. Arpagaus, and M. W. Rotach, 2010: Assessing the benefits of convection-permitting models by neighborhood verification: Examples from MAP D-PHASE. *Mon. Wea. Rev.*, **138**, 3418–3433, <https://doi.org/10.1175/2010MWR3380.1>.
- Whiteman, C. D., 1990: Observations of thermally developed wind systems in mountainous terrain. *Atmospheric Processes over Complex Terrain*, Meteor. Monogr., No. 23, Amer. Meteor. Soc., 5–42.
- , K. J. Allwine, L. J. Fritschen, M. M. Orgill, and J. R. Simpson, 1989a: Deep valley radiation and surface energy budget microclimates. Part I: Radiation. *J. Appl. Meteor.*, **28**, 414–426, [https://doi.org/10.1175/1520-0450\(1989\)028<0414:DVRASE>2.0.CO;2](https://doi.org/10.1175/1520-0450(1989)028<0414:DVRASE>2.0.CO;2).
- , —, —, and —, 1989b: Deep valley radiation and surface energy budget microclimates. Part II: Energy budget. *J. Appl. Meteor.*, **28**, 427–437, [https://doi.org/10.1175/1520-0450\(1989\)028<0427:DVRASE>2.0.CO;2](https://doi.org/10.1175/1520-0450(1989)028<0427:DVRASE>2.0.CO;2).
- Wicker, L. J., and W. C. Skamarock, 2002: Time-splitting methods for elastic models using forward time schemes. *Mon. Wea. Rev.*, **130**, 2088–2097, [https://doi.org/10.1175/1520-0493\(2002\)130<2088:TSMFEM>2.0.CO;2](https://doi.org/10.1175/1520-0493(2002)130<2088:TSMFEM>2.0.CO;2).
- Yu, C.-K., and B. J.-D. Jou, 2005: Radar observations of the diurnally forced offshore convective lines along the southeastern coast of Taiwan. *Mon. Wea. Rev.*, **133**, 1613–1636, <https://doi.org/10.1175/MWR2937.1>.