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

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ABSTRACT

Currently, major efforts are under way to refine the horizontal resolution of weather and climate models to kilometer-scale grid spacing ($\Delta x$). 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 $\Delta x$ 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...
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; White 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 Baalen 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
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 6-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
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 re-analysis 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 $1153 \times 1082$ km$^2$ 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 $\Delta x$ 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 $\Delta x$. 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 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 $\Delta x$, 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
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. 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 \( \Delta x \) 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.
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 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 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 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 $\Delta x$ from (left) 4.4 to (right) 1.1 km.

In Figs. 3d and 3e, the diurnal cycles of precipitation in RAW1 and SM1 are compared against the observed precipitation from the CombiPrecip dataset. 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 precipitation...
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. 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 $\Delta x$ from (left) 4.4 to (right) 1.1 km. Panel (g) shows the observed daytime precipitation in the CombiPrecip dataset.

**FIG. 6.** As in Fig. 5, but for accumulated nighttime precipitation (between 2000 and 0800 UTC).
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 convec-
tive 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 character-
istic 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 2–3 m s\(^{-1}\) are twice as frequent in RAW,
whereas updrafts around > 10 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 \(\Delta x\). The spectral energy peak at wave lengths around 6 km is more pronounced in RAW1 than in SM1,
whereas wave lengths > 10 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
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 ($>0.01$ 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. 8](image-url) 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.

![Fig. 9](image-url) 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 $\Delta x$ is 4.4 (black), 2.2 (blue), and 1.1 (red) km.
orders of $O(100-1000)$ 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 ($\rho q_y v$) 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 ($\rho v$) and not by the moisture content ($q_y$). 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 ($pq_vx$) 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.
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 north-easterly background flow, this signal is partially advected over the southern Alpine flanks and the Po Valley by 2000 UTC.

Figure 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 $q_v = 8$ g kg$^{-1}$ to highlight locations of high $q_v$. The thick purple contour lines indicate cloudy patches (specific cloud liquid water content >1 g kg$^{-1}$).
FIG. 12. Latitude–altitude cross sections of the meridional water vapor flux \( \rho u v \) \( (g m^{-2}s^{-1}) \). 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).
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 \( (pq_y) \) between 1600 and 1800 UTC. The northward branch of the plain-to-mountain circulation (red banners at \( y = 250-350 \) 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 \( y = 350-400 \) 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
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 $\Delta x = 4.4, 2.2, \text{and} 1.1 \text{ km}$: one with fixed low-resolution orography (SM), and one where the representation of orography is refined together with $\Delta x$ (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. 1989, 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.

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