A Modeling Study of a Trapped Lee-Wave Event over the Pyrénées

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ABSTRACT

A trapped lee-wave mountain event in the southern part of the Pyrénées area is analyzed using the Weather Research and Forecasting (WRF) Model. Model experiments are designed to address the WRF predictability of such an event and to explore the influence of the model parameters in resolving the mountain waves. The results show that the model is able to capture a trapped lee-wave event using the 1-km horizontal grid model outputs. Different initial conditions, the vertical grid resolution, and the resolved topography lead to changes in the wave field distribution and the wave amplitude meaning that an ensemble of different model settings may be able to quantify the uncertainty of the numerical solutions. However, the model experiments do not significantly change the wavelength of the generated mountain waves, which is shorter in the three-dimensional real simulations than the one derived from satellite imagery. Comparison with observational data from the surface stations and a wind profiler upstream of the mountain range shows that the model underestimates the horizontal wind speed and this can be the reason for the underestimation of the wavelength. In addition, the valley circulations and the formation of a rotor near the surface are explored. The formation of a low-level rotor in the model is intermittent and brief, and it interacts with other flows coming from multiple directions. The first strong wave updraft is located over the valley aligned with the highest mountain peaks and strong vorticity is captured from the surface up to the first wave crest.

1. Introduction

Mountain waves are topographically generated gravity waves or buoyancy oscillations produced in a stably stratified flow by the disturbance of an air current that encounters an obstacle. Mountain waves impact the atmosphere circulation from the surface to the stratosphere and even to the mesosphere, across a wide range of different scales. They can be accompanied by clear-air turbulence (e.g., Clark et al. 2000), lee-wave rotor formation (e.g., Mobbs et al. 2005; Darby and Poulos 2006; Vosper et al. 2006; Grubišić and Billings 2007; Sheridan et al. 2007; Cohn et al. 2011), downslope winds (e.g., Klemp and Lilly 1975; Mobbs et al. 2005; Sheridan and Vosper 2012), and windstorms (e.g., Lilly 1978), also modifying the surface wind and precipitation intensity.

Linear theory has been applied for the understanding of mountain wave dynamics (Smith 1979). Analytical solutions for a given topography can distinguish between the trapped lee waves and the vertically propagating waves. Trapped lee waves are frequently shown as lee-wave clouds in satellite imagery and they occur when the Scorer parameter ($l^2$) decreases with height (Scorer 1949) or when $l^2$ changes abruptly with height because of a potential temperature inversion. Although linear theory is useful for predicting the wave processes, nonlinearity becomes important when the Froude number ($Fr = U/Nh$) is closer to or less than 1. Then, mountain waves are accompanied by flow reversal near the ground or by mountain wave breaking. In these situations, when the linear theory cannot be applied, data obtained from observational field campaigns and from numerical weather prediction (NWP) modeling becomes fundamental for the understanding of the dynamics of the mountain waves and the associated phenomena (Grubišić et al. 2008; Fritts et al. 2015).
The Pyrénées mountain range has an important influence on the atmospheric flows crossing it, as it is a nearly two-dimensional ridge oriented west–east, so the northern (southern) flows become disturbed along their south (north) side when they pass over the mountain barrier. These disturbances (gravity waves, mountain waves) are able to transport momentum and energy farther downstream or farther up in the vertical and they may also be a hazard for commercial flight routes, which are very frequent in southeast Europe. In addition, they are of interest for glider pilots, who sometimes take advantage of them in order to fly long distances. Mountain waves can also influence the circulation patterns near the surface, within the boundary layer (BL). In particular, trapped lee waves can introduce momentum and energy at its top, they can be absorbed by the BL (e.g., Jiang et al. 2006; Smith et al. 2006; Smith 2007), or they can lead to rotor circulations within the BL and may break into turbulence (e.g., Doyle et al. 2009; Cohn et al. 2011).

Despite their significance, the Pyrénées have not been a focus of study for some time. The most remarkable studies in this area were the Alpine Experiment (ALPEX) field program in the 1980s and the Momentum Budget over the Pyrénées (PYREX) experiment in the 1990s. In 1982, ALPEX addressed the problem of airflow over and around mountains in Europe, and how the resultant features of such airflow affect the global, regional, and local weather. From that campaign, Hoinka (1984) and Cox (1986) analyzed a mountain wave event over the Pyrénées with a synoptic northern flow. Hoinka (1984) derived a vertical profile of the momentum flux from observations and concluded that the mountain wave contributed to the global momentum balance during its brief lifetime. Cox (1986) analyzed the same event, comparing it to Hoinka’s and confirmed the presence of a vertically propagating wave with partially extended trapped lee waves. In 1990, PYREX (Bougeault et al. 1990, 1997) campaign was carried out over the French (northern) of the Pyrénées with the objective of establishing the influence of the large mountains on the mesoscale flow and the effects on the momentum budget. They found that mesoscale models performed reasonably well in resolving the flow and some wave features when they passed over the mountainous terrain (Broad 1996; Masson and Bougeault 1996; Olafsson and Bougeault 1996; Bougeault et al. 1997; Georgelin and Lott 2001). However, they did not deeply study the mesoscale models’ sensitivity to physics parameterizations or configuration in reproducing mountain waves, nor did they investigate the effects of these waves within the BL and the possible formation of rotors, which needs a sufficient refined resolution.

In the last decade, other relevant field experiments have been carried out in order to improve our understanding of mountain wave generation and propagation: the Fronts and Atlantic Storm Track Experiment (FASTEX) over Greenland (Doyle et al. 2005) and the Mesoscale Alpine Programme (MAP) over the Alps (Bougeault et al. 2001; Smith et al. 2007), more recently the Sierra Rotor Project (SRP) and the Terrain-Induced Rotor Experiment (T-REX; Grubišić et al. 2008) over the Sierra Nevada in the United States, and the most recent Deep Propagating Gravity Wave Experiment (DEEPWAVE; Fritts et al. 2015) in New Zealand.

Using two-dimensional (2D) ideal flow modeling, Smith (1989) classified different flow regimes, Vosper (2004) investigated the upstream inversion effects on the lee waves and wave breaking, and Grubišić and Stiperski (2009) and Stiperski and Grubišić (2011) explored the interference over double bell-shaped obstacles. Doyle et al. (2000, 2011) intercompared mesoscale models for a wave breaking in a Boulder, Colorado, windstorm and for the T-REX mountain wave simulations, respectively. In Doyle et al. (2011) they revealed that differences among models were due to numerical techniques, suggesting the need for using an ensemble approach in three-dimensional (3D) simulations. Since mountain waves are often accompanied by low-level turbulent zones, other studies have dealt with the rotor dynamics and BL interaction. Doyle and Durran (2002) analyzed the boundary layer separation and the low-level rotor formation induced by the trapped mountain lee waves using 2D simulations. They concluded that the strength of the rotor is a function of the lee-wave amplitude which decreases as the roughness length increases. In addition, Doyle and Durran (2007) analyzed the internal structure of rotors through 2D and 3D large-eddy simulations (LES), revealing differences in the detailed subrotor structures along the vortex sheet lifted by the lee waves. Later, Jiang et al. (2006), Smith et al. (2006), and Smith (2007) explored the absorption and interaction between waves and the BL and highlighted the dependence of the BL thickness and evolution on the wave features (wavelength, amplitude, and phase). Therefore, for reproducing the lower turbulent zones below the trapped lee waves and the real BL structure, a correct representation of the lee-wave characteristics is crucial.

Using 3D modeling, Doyle and Smith (2003) found a trapped train of lee waves ducted from vertically propagating waves, a consequence of diabatic processes associated with precipitation upstream of the Hohe Tauern crest. Doyle et al. (2005) showed a good representation of a wave breaking event with the COAMPS model and Smith and Skyringstad (2009) showed the influence of different upstream surface heating sources.
in the generation of strong winds and rotors downstream. Specifically using the WRF (Skamarock et al. 2008), other numerical studies of mountain waves in real atmospheric conditions have been done: combining the satellite water vapor imagery (Otkin and Greenwald 2008; Feltz et al. 2009), using radar observations over Scandinavia (Kirkwood et al. 2010), in the Andes Cordillera region (Spiga et al. 2008), over the Basen nunatak in Antarctica (Valkonen et al. 2010), in the upper troposphere–lower stratosphere (Plougonven et al. 2008; Mahalov et al. 2011), and looking at the formation of reversed lee flow during bora (Telišman Prtenjak and Belušič 2009).

As Doyle et al. (2011) pointed out, model dynamical cores are an important component of diversity when designing mesoscale ensemble systems for topographically forced flows. Therefore, studying a model’s performance in reproducing trapped lee waves can be a relevant task that can be done as an ensemble approach. In this paper, we use the WRF Model exploring how the model responds to different model settings such as initial conditions, physics options, vertical grid resolution, and resolved topography to study the predictability of a trapped lee-wave case study and the associated turbulent phenomena near the surface. Since mountain waves are present in many areas, mesoscale models need to correctly represent the flow after crossing a mountain barrier, to adequately resolve the atmospheric circulations. Specifically, in the Pyrénéés area, mountain wave events are repeated during the winter season; therefore, there is a necessity to resolving the phenomena accurately. Furthermore, as the lee-wave structure determines BL circulations and the rotor formation within the BL, a correct representation of the wavelength and the wave amplitude is needed for capturing the lower-troposphere turbulent zones.

For all these reasons, we simulate a trapped lee-wave event over the southern side of the Pyrénéés in order to explore the capability of the model to reproduce it. Mainly, the study aims to determine the influence of different parameters in the WRF Model: the initial and boundary conditions, the vertical and horizontal grid resolution, the different WRF Model physics options, and the underlying topography in resolving the mountain wave event spatial characteristics. Additionally, we look into their effects within the BL, and at its top, exploring the valley circulation patterns and the formation of a rotor near the surface.

The paper is organized as follows. First, the site characteristics are detailed in section 2, along with an analysis of the mountain wave occurrence and model setup. Next, the case study description with the available observations and the model verification is given in section 3. Section 4 focuses on mountain wave characteristics simulated by the model in its different configurations, the wave distribution, wavelength, and the amplitude wave analysis, along with the valley circulations and some rotor signals seen near the surface. Finally, a summary and conclusions are given in section 5.

2. Site description and model setup

a. Site characteristics and mountain wave occurrence

The Pyrénéés mountain range in southwest Europe stands at the border between France and Spain. The mountain range is about 400 km long (west–east) and 100 km wide (south–north) (Bougeault et al. 1997) (Fig. 1a). The average elevation gradually increases from the west to the central part where the highest summits are found, with the highest point (3404 m) in Aneto, Spain. In the eastern part, where we focus our study, the mean elevation is remarkably uniform with a crestline around 3000 m MSL or around 2000 m above the valley floor (Fig. 1b).

In this study, we will consider the mountain wave appearance at the south part of the Oriental Pyrénéés, when northerly flows influence the downwind side in the area around Catalonia, Spain. Ideally, for a northern flow perpendicular to the mountain range, the given topography elevation ($h_m = 2000$ m) within a stable environment ($N = 0.01$ s$^{-1}$) yields a Froude number (Fr) from 0.5 (for $U = 10$ m s$^{-1}$) to 1 (for $U = 20$ m s$^{-1}$). Within this range of values the flow is affected by non-linearity; thus, flow splitting and wave breaking can occur downstream of the mountain range. According to Smith (1989) and Vosper (2004), mountain waves and lee waves will be produced with this characteristic values, with chances to be accompanied by wave breaking and flow splitting.

To identify and characterize the presence of mountain waves in the Pyrénéés, we have used atmospheric soundings and high-resolution satellite imagery from the Moderate Resolution Imaging Spectroradiometer (MODIS) at 1-km resolution. We have analyzed data from 2008 to 2014 taken from atmospheric soundings launched in Barcelona, Spain (see location in Fig. 1a), twice a day. Despite being relatively far away from and downwind of the mountain range, we hypothesize that the conditions upstream may resemble the conditions downstream above the crestline of the mountain barrier. To select the days with possible mountain wave appearances, specifically trapped lee waves, we have imposed the following conditions: (i) a temperature inversion ($\Delta T/\Delta z > 0$) in any layer between 1 and 6 km, (ii) northerly winds between 315$^\circ$ and 45$^\circ$, and (iii) wind
speeds greater than 10 m s⁻¹ in the selected layers. The first condition ensures the presence of a temperature inversion, which delimits a region above where the air is statically stable and waves can develop. The second condition imposes a northern wind component, perpendicular to the orientation of the Pyrénées. The wind speed conditions ensure that part of the flow will be able to cross the mountain barrier.

The sounding analyses (Table 1) reveal that winter is the most likely season for mountain-trapped wave appearances. On average, 52 days per year are found during the seven analyzed years (2008–14), which represents 15% of the days per year for mountain wave occurrence. During the winter season (December–February), 25% of days have good conditions for wave development and March is also a probable month, with a large number of days of occurrence, depending on the year. Although there exist methodologies for verifying the presence of gravity waves from soundings (Zink and Vincent 2001), we have used the satellite imagery to detect the wave occurrence with altocumulus (Ac) lenticular cloud formations. A review of the MODIS satellite imagery during the selected days during the winter among the 7 yr revealed a variety of situations, in some cases with the presence of Ac but others with clear-sky conditions or low-level clouds. If bands of Ac clouds are formed, the presence of a lee wave train is confirmed. However, waves can be also developed without cloud formation when there is not enough moisture available in the atmosphere. In addition, many days with Ac present in the MODIS imagery occurred during the late spring and early autumn, probably because there is

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<td>17.3</td>
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enough moisture availability to reach the condensation level, which leads to cloud formation. From the several days that were selected from the sounding analysis and matching with the presence of Ac, in this study we analyze the specific event that occurred on 13 September 2012, when the MODIS and Meteosat satellites revealed a Ac cloud train formation over the southern part of the Pyrénées (Fig. 2).

b. Model setup and model experiments

In this study, we use version 3.4.1 of the numerical WRF Model that is a state-of-the-art mesoscale model developed by the National Center for Atmospheric Research (NCAR). WRF is a fully compressible non-hydrostatic model, with a terrain-following vertical coordinate and with horizontal and vertical grid staggering of Arakawa C grid type. We use four one-way nested domains of 9-, 3-, 1-, and 0.5-km horizontal resolutions (Δx = Δy) and 200 × 200 grid points in the three first domains (Fig. 1a), centered at 42.48°N, 1.867°E, near the Das (DP) station (Fig. 1b), and 141 × 181 grid points for the fourth inner domain. The control simulation uses 31 nonuniform vertical sigma pressure levels from the surface up to 100-hPa with the first level around 25 m, increasing every 100 m near the surface and increasing 300 m at 2-km height. A damping layer is included in the upper boundary, with an implicit gravity wave 2-km deep damping layer from the model top; results for two additional simulations with a deeper sponge layer of 4 km and higher model top showed very similar results (not shown). The WRF simulations are run from 0000 UTC 12 September to 0000 UTC 14 September 2012, allowing the first 24 h of spinup. The initialization of the simulations is done with the ERA-Interim from the European Centre for Medium-Range Weather Forecasts (ECMWF), which was interpolated at 0.125° in the horizontal and has 36 levels up to 1 hPa in the vertical. The control simulation (ctrl) uses the Yonsei University (YSU) planetary boundary layer (PBL) scheme, a nonlocal scheme that performs well during daytime in unstable conditions (Shin and Hong 2011; Udina et al. 2013). The surface layer for the YSU scheme uses the Monin–Obukhov similarity theory (Zhang and Anthes 1982). Other physics parameterizations used in this study include the Rapid Radiative Transfer Model (RRTM) scheme for longwave radiation (Mlawer et al. 1997), the Dudhia scheme for shortwave radiation (Dudhia 1989), the new Thompson microphysics scheme (Thompson et al. 2004), the Noah land surface scheme (Chen and Dudhia 2001), and the Kain–Fritsch scheme for cumulus (Kain 2004); this last approach is only activated for 9- and 3-km grid spacing domains.

A set of simulations (Table 2) is performed to study the sensitivity of the variables and the mountain wave features to the initial and boundary conditions, model grid resolution, physics options, and model topography. We have studied the resolution effects by evaluating the model outputs obtained from the horizontal grids of the D2 (ctrlD2) and D3 (ctrl) domains at 3- and 1-km horizontal resolutions, respectively. The sensitivity to the initial and boundary conditions has also been investigated: (i) by initializing the simulation 12 h later than the control case, thus starting the simulation at 1200 UTC 12 September 2012 (experiment named i12); (ii) using a coarse domain on a 27-km horizontal grid as a first domain, which leads to having five domains in total (experiment named i5dom); and (iii) changing from the ECMWF global model to the National Centers for Environmental Prediction Final (FNL) Operational Model Global Tropospheric Analyses dataset (NCAR 2013) at 1° horizontal resolution and with 52 vertical levels (experiment named INCEP). Another model sensitivity test has been performed by increasing the vertical resolution from 31 to 45 levels (experiment named z45) and up to 60 vertical levels (experiment named z60). Additionally, different boundary layer physics are explored using the Mellor–Yamada–Janjic scheme (Janjic 1990, 1996; Janjic 2002) (experiment named MYJ), which is a local scheme, and the Mellor–Yamada–Nakanishi–Niino scheme (experiment named MYNN; Nakanishi and Niino2006), which is a nonlocal scheme similar to YSU, to compare with the control case that uses the YSU scheme. Additional tests are done using the surface scheme for complex topography following the work of Jiménez and Dudhia (2012) that corrects the positive wind bias near the plains and valleys and the negative wind speed biases in the mountains and hills (experiment named twind). On the other hand, the 3-km grid-resolved terrain is introduced in the 1-km simulation to show the influence of the unresolved topography (experiment named topo3km).

3. Case study description and model validation

In this study, 13 September 2012 is chosen as a representative day of trapped lee-wave formation that summarizes the main characteristics of the phenomenon and the model performance. The synoptic situation was determined by the Azores high (1025 hPa) over the Atlantic Ocean and a shallow low pressure system over Italy and the Baltic Sea. Both lead to a northerly flow over the Pyrénées.

The visible satellite image from MODIS valid at 1130 UTC 13 September 2012 indicates a broad region of clouds upstream of the Pyrénées and bands of consecutive Ac lenticular clouds downstream (Fig. 2a), over
the Catalan area, which probably formed at the wave crests after the air parcels were displaced from their equilibrium level. These types of clouds reveal a stable layer in the atmosphere where waves are trapped and can travel a long distance from the mountain range (i.e., trapped lee waves). They can also be distinguished in the Meteosat satellite infrared brightness temperature (BT) image from channel 9 at 1100 UTC (Fig. 2b), although
the 3-km grid resolution is not sufficient to observe the details as in the MODIS product.

To evaluate the model performance in the following subsections we use the measurements that are available for this case study: the atmospheric soundings, a wind profiler, and the surface stations, comparing them with the corresponding WRF Model outputs.

a. Atmospheric soundings

Figure 3 shows vertical profiles obtained from the atmospheric soundings launched at 1200 UTC 13 September 2012 (solid line), and vertical profiles from the WRF control simulation at the same time (dashed line) corresponding to the 36-h forecast time. Two different locations are presented: downwind of the Pyrénées, at Barcelona (Bcn; Fig. 3, top), and upwind, at Nimes (Nim; Fig. 3, bottom), which is shifted slightly eastward from the area of interest (see location in Fig. 1a). Air and potential temperature vertical profiles reveal a three-layer structure divided by an inversion layer around 2 km (Figs. 3a,b and 3g,h) and around 5 km, where high values of Brunt–Väisälä frequency ($N^2$) are seen at Bcn (Fig. 3f). At the upwind site, the layer between the surface and 2 km is weakly stably stratified and it becomes strongly stratified above 2 km (Fig. 3g). At first glance, we can see that the model simulation tends to smooth the temperature inversion although the layered structure can be distinguished.

Even increasing the number of vertical levels of the model to 45 or even 60 does not help much with discerning the details of the temperature profiles (the inversions) and discontinuities observed from radiosoundings (not shown). This fact can be a limitation for the model in correctly capturing the mountain wave structure (Agüestsson et al. 2014). In addition, a drying near the 2-km discontinuity is present at both sites (Figs. 3d,j), but a moist layer is maintained aloft at the downwind site, which may be favorable for the lenticular cloud formation. The wind direction is maintained from the north above 1000 m for both locations (not shown), perpendicular to the mountain range. The wind speed increases with height in both locations until the 2-km layer (Fig. 3i). Other upwind locations near the Pyrénées (Bordeaux sounding) show a stronger shear with wind speeds increasing with height (from 15 m s$^{-1}$ at 2-km height to around 30 m s$^{-1}$ at 4 km), which reveals favorable conditions for trapped lee-wave formation (Ray 1986). The simplified Scorer parameter at the upwind site, defined as $\hat{P} = N^2/U^2$, where $U$ is the wind speed, slightly decreases with height above 4 km (Fig. 3k), which is also consistent with the possibility of trapped lee-wave formation (Scorer 1949).

b. Upstream wind profiler

To evaluate the model performance of the three wind components, we use the observations from the ultra-high frequency (UHF) wind profiler radar located in Perpignan (42.73°N, 2.87°E) (see location in Fig. 1b), which are available as part of the Hydrological Cycle in the Mediterranean Experiment (HyMeX) dataset collected during the studied period. Wind profiler measurements are a very useful tool for analyzing the wind structure and for monitoring the properties of mountain waves (Cohn et al. 2011). Although Perpignan is displaced east of the Pyrénées, it gives us an idea of the wind speed profile and evolution in the upstream part of the flow. The vertical wind component and the horizontal wind speed profiles up to 5 km are compared with the simulation through time–height plots from...
0000 UTC 13 September until 2300 UTC 13 September 2012 (Fig. 4), which allow us to validate the upstream conditions in the simulation. Figure 4a shows vertical updrafts before 0900 and until 1200 UTC that are captured by the simulation although they are not continuously maintained and are less strong in intensity (Fig. 4b). The visible MODIS image at 1130 UTC shows lenticular clouds upwind of the Pyrénées (Fig. 2), probably generated by another upwind mountain range, such as the Massif Central. At that time, the UHF radar reveals positive vertical velocities of $\approx 2 \text{m s}^{-1}$, enough for the vertical displacement of the parcels and the cloud formation at the wave crests. Although mountain waves are stationary, a time oscillating behavior is seen in the measured wind profiler vertical velocity at Perpignan during the following hours. Indeed, lee waves can exhibit nonstationarity (Nance and Durran 1997; Ralph et al. 1997; Caccia et al. 1997) with time variation of the
lee waves due to transitions in the background flow. Regarding the horizontal wind, a strong wind current is established around 1100 UTC at heights between 1 and 3 km (Fig. 4c), when the lenticular clouds are also seen in the satellite images. The changes in horizontal wind speed are generally well reproduced by the model but the magnitude is, overall, underestimated, at around 3 or 4 m s\(^{-1}\) (Fig. 4d).

c. Surface stations

The simulated 24-h outputs of 2-m temperature and 10-m wind speed are validated with surface station measurements for the different model configurations (Table 2) and results are summarized with statistics (Table 3). The upwind surface stations data are obtained from the surface synoptic observation (SYNOP) measurements and are compared with the D2 (3-km horizontal grid) domain model outputs, as D3 does not include enough upwind area (see Fig. 1). The results from downwind surface stations are taken from the Catalan Meteorological Service and compared with those from the 1-km domain.

As shown in Table 3, upwind of the Pyrénées, all 3-km grid model simulations have a negative bias in temperature, with better correlation metrics and biases for the z60, iNCEP, and i12 cases. However, the poor 2-m temperature correlation indicates that 3-km resolution may not be fine enough. On the downwind side, using a 1-km horizontal grid, the best correlation for temperature is for the ctrl case and the smallest bias and root-mean-square error (RMSE) is obtained with the simulation initialized at 1200 UTC, i12. In all cases, the best metrics are obtained using the YSU PBL scheme, which may represent the convective unstable conditions in a better way.

The 10-m wind speed is generally overestimated by the model on the downwind side, as a positive bias is
Table 3. Statistics for the observed and simulated 2-m surface temperature and 10-m wind speed for the 24-h period from 0000 to 2300 UTC 13 Sep 2013, for the performed model experiments described in Table 2. The calculated statistics correspond to the correlation coefficient (CC), the mean bias (MB), and the RMSE between the upwind surface stations from the SYNOP database and the 3-km simulation experiments and the downwind surface stations from the Catalan Meteorological Service (meteo.cat) and the 1-km simulation experiments. The values highlighted in boldface are the best statistical values for each variable and section.

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<tr>
<td>MYNN</td>
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</tr>
<tr>
<td>twind</td>
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<td>−1.582</td>
</tr>
<tr>
<td>topo3km</td>
<td>0.695</td>
<td>−1.448</td>
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<table>
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<th>10-m wind speed</th>
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<td></td>
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<td>MB</td>
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</tr>
<tr>
<td>topo3km</td>
<td>0.927</td>
<td>−1.041</td>
</tr>
</tbody>
</table>

The obtained model results show an underestimation of the upstream horizontal wind speed both in the wind profiler and in the surface stations verification, which can be a relevant factor for the wave characteristics produced downstream. Biases in the model surface wind speed performance are not new (Jiménez and Dudhia 2012). Among other factors, a better initialization, with finer horizontal and vertical resolution, less smoothing in the model fields, and an accurate land-use representation would improve the wind representation in the numerical models. In our simulated case, the MYJ local PBL scheme with the associated Eta surface layer (Janjic’ 1996, 2002) seems to perform better when representing the 10-m wind speed upstream surface stations. However, when the flow has passed the mountain range, the best 10-m wind representation is found for the YSU nonlocal PBL scheme using the surface parameterization from Jiménez et al. (2012).

4. Simulation of mountain waves

a. Mountain wave characteristics

The relative humidity and vertical wind horizontal plane views (Fig. 5) from the WRF indicate high-amplitude wave train formation over northern Catalonia, after crossing the Pyrénéen mountain range, at 1200 UTC. Vertical velocities exceed 6 m s$^{-1}$ at around 4 km MSL (Fig. 5b) with ascent–descent couplets after crossing the first crestline and farther downstream, where a second crestline is also present. According to the model, the mountain waves remain stationary for a few hours and are extended even farther downstream, with the maximum intensity between 1000 and 1600 UTC. The general pattern and the main magnitudes resemble the observations from the north side of the Pyrénéen, which have been obtained by several studies during the PYREX campaign (Broad 1996; Masson and Bougeault 1996; Georgelin and Lott 2001).

Vertical cross section through the middle of the domain following the solid line in Fig. 5b reveal the vertical extension of the wave field, which is strongly developed from 2- to 6-km height and smoothed above (Fig. 6). The horizontal wind is perturbed after crossing the crestline with a phase opposite that of the potential temperature (Fig. 6a, top), which follows the linear theory (Gossard and Hooke 1975; Scorer 1949) and the buoyancy-generated gravity waves structure, similar to Vosper (2004), Smith et al. (2007), and Mahalov et al. (2011) above the PBL, as well as to Sun et al. (2015) within the PBL. Flow deceleration is seen below the wave crest down to the ground, in the valley area (more in section 4c). Since the wind speed increases with height (Scorer 1949),

obtained in all model setups. In contrast, it is generally underestimated on the upwind side of the Pyrénéen for all the experiments using 3-km model outputs, compared with the surface station data obtained from the SYNOP measurements. The standard deviation of the WRF is larger than that from the observed wind speed on the downwind side, meaning that the model tends to exaggerate the wind variability (not shown). The best metrics in wind speed, with significant reductions in the positive bias and the RMSE, are obtained by the twind simulation over the 1-km domain, which uses the YSU scheme for the PBL and has a surface parameterization proposed by Jiménez et al. (2012) that leads to a better representation of the horizontal wind speed near the surface in this complex terrain area.
the Scorer parameter decreases with height; thus, the vertical structure of the atmosphere is favorable for a trapped lee wave train formation (Scorer 1949; Vosper et al. 2006). In addition, the amplitude of the waves is evanescent with height above 6 km. In the horizontal, the lee waves are extended downstream and attenuated by the BL (Jiang et al. 2006) or dissipated farther downstream. According to the model, the isolines of water vapor mixing ratio are in phase with the isentropes and few bands of water clouds are formed around the altitude of the crestline below the wave crests (not shown).

The wavelet function applied to the model output spatial transects allows us to investigate the wavelength and the spatial location of the generated waves. A Morlet function is used as a mother wavelet. Wavelet plots show the power spectrum of the magnitude in color, with the latitude along the x axis and the number of grid points (ngrid) along the y axis. The corresponding wavelength ($\lambda$) is calculated as the product of the number of grid points and the horizontal resolution ($\lambda = n\text{grid} \times \Delta x$); thus, $\lambda = n\text{grid}$ in the 1-km grid simulations. The power spectrum is calculated in each vertical level, then averaged between levels from 3 to 6 km, in the region where the waves are fully developed.

The power spectrum from the horizontal wind speed (Fig. 6, bottom), potential temperature (not shown), and the vertical velocity (Fig. 8a, bottom) transects shows a strong signal around $\lambda = 12$ km in a broad region leeward of the mountain, after crossing the second crestline. In theory, the simulated wavelength from the potential temperature fields can be compared with the observed wavelength derived from the lenticular clouds from the satellite images, because a cloud is likely to be formed at each wave crest. The separation distance between clouds is also calculated using the wavelet power spectrum (Fig. 7), as obtained from the wavelet function applied to the BT field from the Meteosat satellite along a fixed longitude (red line in Fig. 2b). As the Meteosat satellite has 3-km grid resolution, the observed signal given by the power spectrum is between $n\text{grid} = 6–7$ (around 42°N), which corresponds to a wavelength $\lambda = 18–21$ km (see white mark in Fig. 7). Looking at the MODIS image (Fig. 2), we can approximate a similar distance interval between the lenticular clouds. Therefore, the calculated wavelength of $\lambda = 18–21$ km obtained from the satellite imagery is larger than the $\lambda = 12$ km found in the model. Our first hypothesis is that these differences between the wavelength derived from the satellite images and the wavelength obtained from the simulation can be a result of the imposed upstream conditions. As we have seen from the measurements in section 3, the model is generally underestimating the horizontal wind speed, according to the wind profiler and the surface stations located on the northern side of the Pyrénées. Following the lee-wave linear theory for an idealized 2D flow where the wind and the stability increases with height, an intensification.
of the incident wind (stronger wind) would lead to a longer wavelength (Smith 1989; Wurtele et al. 1996; Vosper 2004). Therefore, we think that the underestimation of the wind speed intensity upstream of the mountain range can be the reason for the model underestimation of the wave wavelength. On the other hand, we also hypothesize that in the real atmosphere the lenticular clouds may not be present in all the wave crests and so the estimated wavelength from the satellite lenticular clouds may be larger than the one of the wave itself. As we have no other measurements than satellite images to verify the characteristics of the mountain waves, we cannot verify this hypothesis. In addition, the different vertical structures of the Scorer parameter between the observed and modeled profiles (Figs. 3e,k) may be also the reason for the wavelength discrepancy (Scorer 1949; Wurtele et al. 1996).

To explore the influence of the model configuration parameters in the resolved wave field and wave characteristics, the next subsection shows the results of the mountain wave patterns using different model options.

b. Sensitivity to model options

Using the set of simulations detailed in section 2, we explore the sensitivity of the wave characteristics in model outputs to different model configurations such as initial conditions, physics options, vertical resolution, resolved topography, and horizontal resolution. The analysis is done with the vertical velocity ($w$) and potential temperature ($\theta$) fields (Fig. 8) through the same cross section as in Fig. 6. In addition, in Fig. 9 we present the vertical velocity standard deviation for four groups of simulation experiments in order to compare the dispersion among them and the sensitivity to the model.
settings, similar to what was done in Doyle et al. (2011). Here, the comparison between experiments is shown for a group of different initial conditions (ctl, i12, i5dom, iNCEP; see Fig. 9a), a group of different PBL and surface physics options (ctl, MYJ, MYNN, and twind; see Fig. 9b), using the three vertical levels experiments (ctl, z45, and z60; see Fig. 9c), and changing the underlying topography (ctl and topo3km; see Fig. 9d).

In Table 4 the main parameters of the waves for each experiment are summarized. The wavelength is determined by the wavelet analysis for the vertical velocity transect between 3 and 6 km (included at the bottom of each simulation experiment). The wave amplitude in the vertical dimension is calculated as the maximum value of the half-altitude difference between the first wave through and crest along the transect for each simulation. The horizontal wind speed corresponds to the averaged value between 2000 and 4000 m MSL at an upstream point (42.61°E, 1.92°N).

In all the 1-km resolution simulation experiments (Figs. 8a–f) the vertical wind cross section shows ascent–descent couplets in phase with potential temperature isentropes after crossing the Pyrénées, forming a wave train that is extended more than 100 km downstream. The first updraft at around 42.5°N matches with the maximum of the vertical wind component. Regarding the time evolution, gravity waves are stationary during several hours, and, as we should expect from linear theory, \( \theta \) is out of phase 90° with \( w \). The wave train vertical extension starts around 2–3 km MSL, where the higher mountain peaks are located and smooths out above 6 km.

All 1-km experiments (ctl, i12, i5dom, iNCEP, z45, z60, MYJ, MYNN, and topo3km) produce similar wavelengths, between 12 and 14 km (Table 4); results that are still much smaller than the one extracted from the satellite images. Although the differences are very small, the experiments with larger wavelengths match with stronger upstream horizontal wind speeds, as the wavelength tends to increase as the incident wind speed increases, consistent with a reduced Scorer parameter. The amplitude of the reproduced waves varies between 646 and 889 m among the experiments (Table 4), with the lower values corresponding to the strongest upstream horizontal wind speed cases.

In the control simulation (Fig. 8a), the dominating wavelength is \( \lambda = 12 \) km, consistent with the wavelet analysis from the \( \theta \) transects in Fig. 6. The maximum wave amplitude over the transect is found around 4-km altitude with values around 727 m with vertical velocities around 6 m s\(^{-1}\). These magnitudes are similar to results for events observed previously in the area, such as by Hoinka (1984) and Georgelin and Lott (2001).

Changing to different initial and boundary conditions (Fig. 9a) leads to a moderate variability in the amplitude of the resolved waves, mostly after crossing the second crestline. In particular, the iNCEP case (Fig. 8b, top) slightly reduces the wave amplitude and therefore the vertical wind intensity, which could be due to the better vertical resolution of the initial data. In contrast, the i5dom experiment enhances the wave amplitude (see Table 4). The wavelengths in all cases are similar to or slightly larger than in the ctrl case (Fig. 8b, bottom, and Table 4).

The smallest standard deviation is found within the physics group of experiments (Fig. 9b), meaning that changes in PBL schemes and surface layer parameterization do not affect the wave’s wavelength or amplitude, which seems reasonable, as the waves are generated above these layers.

Another sensitivity test is performed by increasing the vertical resolution from 31 (ctl) to 45 levels (z45) (Fig. 8c, top) and up to 60 vertical levels (z60) (Fig. 8d, top). Although the wavelengths remain similar between the experiments (Fig. 8c, bottom, and Fig. 8d, bottom), the amplitude of the generated waves shows large dispersion over the plain after crossing the mountain barrier (Fig. 9c), so that the amplitude increases when increasing the vertical levels. In addition, the refinement in the vertical levels seems to propagate the wave train farther downstream.

To examine the influence of the resolved topography, the experiment topo3km uses the terrain height of the 3-km resolution (domain D2), meaning that the terrain elevation is smoothed and less detailed than in the control case (Figs. 8e and 9d). We can see that the less well resolved terrain does not significantly affect the generated wave pattern near the first crestline but it does.

![Fig. 7. Wavelet power spectrum of the Meteosat BT from an infrared channel 9 image at 1100 UTC along a north–south transect at the fixed longitude of 2.2°E, corresponding to the red line in Fig. 2b. The white circle indicates the region over the Pyrénées. The hatched area corresponds to the cone of influence, the region of the wavelet spectrum in which edge effects become important.](image-url)
Fig. 8. Cross sections of the vertical wind component (m s$^{-1}$, shaded), potential temperature (K, gray contour lines), and the underlying terrain (black shaded) at 1200 UTC 13 Sep 2012 following the line in Fig. 5 for (a) ctrl, (b) NCEP, (c) z45, (d) z60, (e) topo3km, and (f) ctrlD2 simulation experiments, with its corresponding averaged power spectrum at the bottom part, as in Fig. 6.
far downstream of the mountain range, above the plain (Fig. 9d). In addition, the wavelength is slightly larger for the topo3km case than the ctrl case ($\lambda = 14$ km versus $\lambda = 12$ km). Indeed, the second crestline is smoothed in topo3km, and it is not interfering with the wave generated after the first crestline (Stiperski and Grubišić 2011), which could be the reason for the large dispersion and the larger wavelength.

Table 4. Wave parameters among model experiments. The wavelength corresponds to the value where the signal is stronger according to the wavelet analysis from Figs. 8. The amplitude is calculated as the maximum value of half-altitude difference between the first wave through and the crest along the transect. The horizontal wind speed corresponds to the averaged value between 2000 and 4000 m MSL at an upstream point (42.61°N, 1.92°E).

<table>
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<tr>
<th>Simulation</th>
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<th>Amplitude (m)</th>
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<td>—</td>
<td>—</td>
</tr>
<tr>
<td>ctrl</td>
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<td>topo3km</td>
<td>14</td>
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<td>23.69</td>
</tr>
</tbody>
</table>
The final experiment shown is the ctrlD2 case corresponding to the coarser-resolution 3-km outputs (Fig. 8f). In this case, the vertical velocity cross section reveals smoother updrafts and downdrafts, which are less intense in magnitude, as well as smaller in amplitude but also in their extension downstream, similar to results reported by Smith et al. (2007). An important updraft after the first range is captured. The wavelet signal is seen between 4 and 7 grid spacings and so between $\lambda = 12-21$ km ($\lambda = n_{grid} \times 3$), thus generating a larger wavelength than the other experiments (Fig. 8f, bottom). Although the updraft after the first crestline is reproduced, no clear mountain waves are reproduced in this case, as the 3-km horizontal resolution may be too coarse to resolve the mountain waves in this case.

From these results, we conclude that different model configurations lead to a wide range of solutions in terms of mountain wave locations and amplitudes. Besides the need for fine horizontal resolution (1 km in our case), changes in the initial conditions, the number of the configured vertical levels and the resolved details in the topography shape influence the wave field distribution and its amplitude, but ultimately they do not vary the wavelength much. An ensemble of model simulations with high horizontal resolution including a variety of initial and boundary conditions and different vertical levels could be a good approach for representing the uncertainty of the numerical solutions.

c. Valley circulations and rotor signals

Trapped lee waves can be associated with rotors and turbulent zones that usually develop under the wave crests (Doyle and Durran 2002; Vosper 2004; Hertenstein and Kuettner 2005). Here, we investigate the presence of these rotor structure in La Cerdanya valley, a 10-km wide valley oriented southwest to northeast, whose base is around 1100 m MSL and with surrounding peaks that reach 3000 m (Fig. 10a). Looking in detail at the first wave crest (Fig. 10b), we can see two maximum updrafts with a maximum overpassing vertical velocities of around 8 m s$^{-1}$ located over the northern downslope in the La Cerdanya valley, after the higher mountain crestline that the northern flow encounters. One is located aligned with the Carlit mountain peak, which is one of the highest in the French area, at an elevation of 2909 m, and the other maximum is aligned with the Puigpedrós peak of 2915 m. A similar location for these maximum updrafts of the first wave crest was also seen in other simulated mountain wave episodes (not shown). This is in agreement with the linear theory of mountain waves, as the vertical velocity increases with the mountain height. Next, we explore the possible formation of a wave rotor system in the area below these maximum updrafts, similar to the one from the T-REX experiment reported upon by Cohn et al. (2011).

A valley cross section (Fig. 11) is plotted to illustrate the circulation patterns over the valley after the strong updraft aligned with the Carlit mountain peak, for three different PBL experiments: control simulation (using YSU), MYJ, and MYNN. The cross section corresponds to the black line in Fig. 10b. As it is a narrow valley, we use the 500-m grid domain (D4) to capture the topography details and to be able to resolve the flow circulations in a better way while still using the mesoscale approach. However, a turbulence-resolving model, such as a large-eddy simulation model, would be necessary to fully resolve the rotor structure.

From north (left) to south (right) after the downdraft over the slope, a strong updraft occurs before the wave crest followed by a downdraft over the center of the valley for all PBL experiments (Figs. 11a–c). The maximum vertical wind is located between 3 and 4 km MSL but a strong updraft starts near the surface. We can also see that the PBL height (green dots in Figs. 11a–c) is highly influenced by the wave shape, following the potential temperature isolines with the wave updrafts and downdrafts over the valley, mostly for the YSU and MYNN experiments. In the MYJ case it is more irregular and does not follow the wave shape as clearly as in YSU and MYNN above the downstream plane. The YSU and MYNN schemes are nonlocal schemes that calculate the PBL height in convective situations using the Richardson number criteria, based on the potential temperature and wind speed gradients; then, the PBL top follows the isentropes of the wave train. Instead, the height of the PBL in the local MYJ scheme is determined as the lower level where the turbulent kinetic energy approximates a minimum value of the length scale, and therefore, is less dependent on the potential temperature.

The horizontal wind speed is weak and variable over the valley, below the first wave crest, while it is very strong above it (Figs. 11d–f); therefore, a strong shear is seen at the intersection (as in Grubišić and Billings 2007). In all experiments, the $y$-wind vectors suggest a recirculation zone from the valley surface up to 1 or 1.5 km above it (Figs. 11d–f). In addition, similar to the experiments performed by Doyle and Durran (2007), the $x$ component of the horizontal vorticity ($\eta = \partial w/\partial y - \partial v/\partial z$) has large positive values along the upstream edge of the lee wave (Figs. 11g–i). From the MYJ and MYNN experiments we can see different locations of the turbulent kinetic energy (TKE) maxima. In both cases these high values of TKE are located along the upstream edge of the first wave, where the maximum
Vorticity was also seen, as in Doyle and Durran (2002) and Doyle and Durran (2007). In MYJ there is also a maximum TKE at the center of the valley, close to the surface (Fig. 11h), while in MYNN the TKE maxima occur below the two small crests at elevated heights (Fig. 11i). We can see that different physics parameterizations lead to a variety of solutions for the location of the turbulent zones, PBL top, and wind circulations; therefore, an ensemble of different model settings could be a good strategy for resolving the uncertainty of the model solutions.

In all cross sections, the wind vectors near the plain and below the wave crest suggest a recirculating wind zone. Despite the signals of wind reversal found near the surface, the wind is weak and variable in these valley areas within the first hundreds of meters above ground (Fig. 12a). Thus, there are some signals, but no clear wave rotor system (as in Cohn et al. 2011) is seen in our 3D real simulation experiments because the wind is flowing from other directions near the surface rather than from the north to the south, which is the dominant pattern. Similar to Kühnlein et al. (2013), we conclude that the formation of a low-level rotor is transient and intermittent and interacts with the along-valley flow and other flows from other directions. We can identify the turbulent area below the first wave crest where the rotor is likely to be formed. However, the 2D ideal classical rotor structure is replaced by a more complex turbulent flow of multiple scales and directions (Kühnlein et al. 2013). Moreover, the small subrotors along the ascending branch of the mountain lee wave seen by Doyle et al. (2009) and Cohn et al. (2011) are not reproduced in our model simulations, mainly because finer horizontal and vertical resolution is needed. In addition, different PBL scheme parameterizations lead to similar patterns near the surface in terms of the wind and temperature structure, although parameterized turbulent kinetic energy is differently distributed.

From this case study and after analyzing several northern wind events, we can hypothesize a detailed circulation pattern over the Cerdanya valley when it is influenced by northern flows and mountain waves. The strongest wave updrafts occur along the lee side of the highest mountain peaks of Carlit and Puigpedrós.
FIG. 11. Valley cross section (see black line in Fig. 10) at 1300 UTC 13 Sep 2012 for (a),(d),(g) the ctrl case and the (b),(e),(h) MYJ and (c),(f),(i) MYNN configurations of (a)–(c) the vertical wind component (m s$^{-1}$, shaded), the potential temperature (K, contour lines), and PBL height (green dots); (d)–(f) the horizontal wind speed (m s$^{-1}$, shaded), potential temperature (K, contour lines), and y wind component (m s$^{-1}$, vector); and (g)–(i) the x component of the horizontal vorticity (s$^{-1}$, shaded) and potential temperature (K, contour lines). For the (h) MYJ and (i) MYNN experiments, the TKE prognostic variable is also plotted in purple lines.
Aligned with them, over the valley area, winds are weak and variable, which indicates possible areas where rotors may be formed (Fig. 12b). In any case, to explore this conceptual model about the valley’s circulation patterns, measurements of vertical profiles of the main magnitudes would be needed around the area. In addition, small-scale simulations would help to interpret the turbulence structure and location of low-turbulent zones and rotors near the surface.

5. Summary and conclusions

In this paper, we have studied the mountain wave phenomena over the Pyrénées, a mountainous region that had not been examined for many years. We have documented a trapped lee mountain wave event in this area through mesoscale simulations using the WRF Model and we have explored the predictability of such an event. Model results show trapped lee waves, illustrated by wind, potential temperature, and humidity fields after crossing the mountain range, that are extended several hundreds of meters farther downstream. The mountain wave vertical extension reaches from the mountaintop (around 3 km) up to the middle and upper troposphere, although the amplitude of the waves is strong up to 6 km and smoothed above.

The specific event that occurred on 13 September 2012 is described in detail, although other selected episodes were also simulated and analyzed. In this case, the satellite images from MODIS and the Meteosat brightness temperature field revealed lenticular cloud formation over the southern part of the Pyrénées. The ability of the WRF Model to reproduce this trapped lee-wave event has been analyzed through a series of simulations that varied the basic model configuration. We have seen that the model can generate mountain waves downwind of the Pyrénées in model outputs using a 1-km horizontal grid and that coarser horizontal resolutions (3 km) are not adequate to obtain the wave fields. From the different model experiments we conclude that the initial and boundary conditions, the number of vertical levels, and the details in the resolved topography influence the wave field distribution and their amplitude but do not have much of an effect on the wavelength. Changes in physics options, the PBL, and surface layer parameterizations have very low impact in the mountain wave solutions but lead to different results in the generated PBL height top and the three-dimensional wind circulation patterns with associated turbulence near the surface. We conclude that an ensemble of different model settings may be able to quantify the uncertainty of the numerical solutions.

A 24-h model experiment verification is also included. The surface stations’ validation revealed that the 2-m temperature is underestimated on the up- and downwind sides of the Pyrénées. The best model configurations for the temperature forecasts are for the experiments using the YSU PBL parameterization, which would simulate the
convective, unstable flow in a better way. The 10-m wind speed model verification reveals an underestimation on the upwind side using the surface stations, which is confirmed also by the UHF wind profiler data, and an overestimation on the downwind side. Poor wind correlation is found among all of the model experiments although the bias is improved in the 1-km horizontal grid twind simulation, using the adjustment of Jiménez and Dudhia (2012).

The mountain waves obtained with 3D real simulations have a shorter wavelength than those derived from the satellite images. We hypothesize that the reason for the different wavelengths could be in the underestimation of the wind speed profiles upstream of the mountain range, as it is shown by the model evaluation. Furthermore, we suggest that the lenticular clouds may not be formed in all wave crests, which would also explain the wavelength discrepancy, although we do not have enough measurements to prove this statement.

The presence of a rotor has also been investigated over La Cerdanya valley, where the first wave updraft is usually located above a region of weak and variable wind near the surface, where turbulence may be generated. Results suggest that flows over the plain are highly turbulent, influenced by the presence of the updraft aloft, but no clear ideal 2D rotor structure is captured at 500-m grid resolution, which can be due to the complexity of the flow, including multiple scales and flow interactions, as pointed out in Kühnlein et al. (2013).

Hence, further investigation is still needed to understand the turbulent zones within the rotors associated with mountain waves and their interaction with the boundary layer. For this purpose, in addition to vertical atmospheric measurements, increasing the horizontal and vertical resolution in numerical modeling seems to be necessary. Thus, a turbulence-resolving model, like a large-eddy simulation model, should be adequate for this future research as these models are designed to directly simulate turbulent eddies that control the boundary layer dynamics with more fidelity than models with parameterized turbulence closures. It is known that a portion of the energy-carrying turbulent eddies is resolved in the model, whereas most other options, such as mesoscale simulations, completely parameterize turbulent fluxes (Smith and Skyringstad 2009).

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