Cloud-Resolving Simulations of Deep Convection over a Heated Mountain

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ABSTRACT

Cloud-resolving numerical simulations of airflow over a diurnally heated mountain ridge are conducted to explore the mechanisms and sensitivities of convective initiation under high pressure conditions. The simulations are based on a well-observed convection event from the Convective and Orographically Induced Precipitation Study (COPS) during summer 2007, where an isolated afternoon thunderstorm developed over the Black Forest mountains of central Europe, but they are idealized to facilitate understanding and reduce computational expense.

In the conditionally unstable but strongly inhibited flow under consideration, sharp horizontal convergence over the mountain acts to locally weaken the inhibition and moisten the dry midtroposphere through shallow cumulus detrainment. The onset of deep convection occurs not through the deep ascent of a single updraft but rather through a rapid succession of thermals that are vented through the mountain convergence zone into the deepening cloud mass. Emerging thermals rise through the saturated wakes of their predecessors, which diminishes the suppressive effects of entrainment and allows for rapid glaciation above the freezing level as supercooled cloud drops rime onto preexisting ice particles. These effects strongly enhance the midlevel cloud buoyancy and enable rapid ascent to the tropopause. The existence and vigor of the convection is highly sensitive to small changes in background wind speed $U_0$, which controls the strength of the mountain convergence and the ability of midlevel moisture to accumulate above the mountain. Whereas vigorous deep convection develops for $U_0 = 0$ m s$^{-1}$, deep convection is completely eliminated for $U_0 = 3$ m s$^{-1}$. Although deep convection is able to develop under intermediate winds ($U_0 = 1.5$ m s$^{-1}$), its formation is highly sensitive to small-amplitude perturbations in the initial flow.

1. Introduction

The mechanisms responsible for the initiation of deep moist convection have been increasingly studied through explicit numerical simulation. Such studies often consider horizontally homogenous environments in which an ensemble of convective plumes forms in response to the diurnal cycle (e.g., Derbyshire et al. 2004; Khairoutdinov and Randall 2006; Wu et al. 2009). Some consistent themes have emerged, notably that the traditional parcel analysis of convective available potential energy (CAPE) and convective inhibition (CIN) is unable to explain the transition from shallow to deep convection, which depends critically on mixing between clouds and their environment. This transition only occurs when parcels are able to maintain buoyancy by minimizing and/or overcoming the dilution and evaporative cooling induced by clear-air entrainment. Deep convection is favored not only by stronger moist instability but also by moister midlevel flow, which weakens the suppressive impacts of entrainment on ascending cumuli, and wider updrafts, which are less diluted by a given amount of environmental mixing.

Although horizontally homogeneous flows are a useful idealization, real flows commonly possess localized regions of uplift that concentrate deep convection. One prominent uplift mechanism is thermally induced mountain circulations, which are driven by horizontal buoyancy gradients generated by elevated solar heating. These circulations draw boundary layer air up the mountain slope and vent it into the free troposphere within sharp convergence zones (e.g., Banta 1990). A full understanding of these circulations remains elusive because of their nonlinearity and complex lower boundaries, which present formidable analytical challenges, but some basic physical insight is available from linear theory (e.g., Crook and Tucker 2005). This theory generally agrees with observations and numerical simulations in that the thermal circulations are...
enhanced by stronger sensible heating rates, more extensive mountains, and weaker background winds. Background winds suppress convection by ventilating heat away from the mountain, which weakens the local buoyancy gradients. They also force stronger mechanical descent in the lee, which competes against the thermally driven updrafts.

What distinguishes the evolution of mountain convection from that in horizontally homogenous environments? Valuable insight was provided by Zehnder et al. (2009), who presented observations of two convective events over the Santa Catalina Mountains in southern Arizona during the Cumulus Photogrammetric, In Situ, and Doppler Observations (CuPIDO) experiment in 2006. In one case the transition from shallow to deep convection was initiated by gravity wave activity associated with a nearby cumulonimbus cloud; in the other it occurred because of rapid midlevel moistening associated with the detrainment of shallow cumuli. The latter mechanism is consistent with the transition over flat terrain except that the cumulus preconditioning is narrowly focused over the mountain and highly sensitive to the ambient winds (e.g., Orville 1968; Banta 1990; Damiani et al. 2008). Under weak winds, midlevel moisture may build up for several hours, leading to stronger preconditioning and a greater likelihood for deep convection. However, even modest winds of 5–10 m s\(^{-1}\) may strongly diminish the preconditioning by transporting moistened air away from the high terrain.

Another characteristic of mountain convection is that cumuli may form in such rapid succession that new thermals pass through the saturated wakes of their predecessors before they have time to detrain. The majority of previous studies into this process have focused on its role in precipitation initiation (e.g., Mason and Jonas 1974; Roesner et al. 1990; Blyth and Latham 1993; Ovtchinnikov et al. 2000), but few have focused on its role in facilitating the transition to deep convection. In this regard, successive thermals may be important for diminishing the suppressive effects of entrainment by shielding fresh updrafts from the dry ambient flow. Moreover, when this wake contains frozen water particles, sedimentation or entrainment of these particles into fresh updrafts may rapidly accelerate cloud glaciation through riming and/or vapor diffusion. This enhances the midlevel latent heat release and may help the cumuli to survive ascent through a marginally unstable and dry atmosphere. On the other hand, preceding clouds may have a suppressive effect on subsequent convection by releasing conditional instability and generating outflow boundaries that terminate the sequence of clouds.

To improve the understanding of mountain convection, recent numerical studies of observed events have been carried out at \(O(1\ \text{km})\) resolution (e.g., Demko and Geerts 2010; Barthlott et al. 2011). Although such experiments are often successful in reproducing gross features of the observed convection, their inability to adequately resolve boundary layer eddies and subcloud-scale drafts may introduce significant errors in the timing, structure, and evolution of the convective cells. Along with a mountain-scale toroidal circulation (called the convective core), boundary layer eddies are a fundamental component of the flow response to mountain heating (e.g., Raymond and Wilkening 1980). The absence of these features leads to artificially coherent circulations that lack turbulent variability. In addition, the absence of subcloud-scale drafts that regulate mixing between cumuli and their surroundings leads to quasi-laminar clouds with artificially low entrainment rates. This may result in overly rapid cumulus growth and inflated precipitation amounts (e.g., Bryan et al. 2003). Spatial resolutions of at least \(O(100\ \text{m})\), which are commonly referred to as “cloud-resolving” resolutions, are needed to close the scale gap between the resolved, buoyancy-driven eddies and the inertial subrange.

In this study, we conduct cloud-resolving simulations of the diurnal cycle over a mountain ridge to provide a detailed analysis of the mechanisms by which deep convection develops in thermally driven mountain flow. These simulations are idealized from intensive observational period (IOP) 8b from the Convective and Orographically Induced Precipitation Study (COPS), which took place over the Vosges and Black Forest mountains in central Europe (Wulfmeyer et al. 2008). In this event, a deep cumulonimbus developed within a band of shallow cumuli over the Black Forest and then rapidly dissipated as it detached from the mountain region (Kottmeier et al. 2008). The fleeting nature of this storm reflected the fact that the mountain forcing was barely adequate to overcome the environmental resistance to deep convection, which included strong convective inhibition and a dry midlevel flow (Barthlott et al. 2011). The hostility of this flow to deep convection makes it a useful case for closely examining the role of mountain forcing in preconditioning the atmosphere for deep convection.

2. Numerical setup

We use the Bryan cloud model version 14 (Bryan and Fritsch 2002), which is a fully nonlinear, compressible, and nonhydrostatic model designed for the explicit simulation of moist convection. To minimize computational expense and allow for many high-resolution experiments, we adopt a 2D configuration, which has provided acceptable representations of convective clouds in many previous studies (e.g., Redelsperger et al. 2000; Wu et al. 2009; Robinson et al. 2011). However, some significant errors have also been reported, including an overly rapid
transition from shallow to deep convection owing to reduced cloud entrainment rates (Grabowski et al. 2006; Petch et al. 2008). Although these simulations must thus be interpreted with caution, they are still a useful exercise that will help to motivate and guide the design of more computationally expensive 3D studies.

Forward time integration of the model equations uses a third-order Runge–Kutta scheme with time-splitting to maintain stability of acoustic waves. Sixth-order centered horizontal advection is coupled with sixth-order horizontal diffusion to limit spurious grid-scale noise. Fifth-order advection with implicit diffusion is used in the vertical. Positive-definite advection is applied to moisture variables to enforce water conservation. Subgrid-scale turbulence is parameterized using a 1.5-order turbulent kinetic energy (TKE) scheme, and cloud microphysics are parameterized using the Morrison two-moment scheme. The cloud droplet concentration is set to a uniform value of 1000 cm$^{-3}$ to represent polluted air over central Europe. No parameterizations are used for the atmospheric boundary layer (ABL) or moist convection, which are explicitly resolved. Surface drag is represented using a bulk-transfer relationship with a drag coefficient of $C_d = 0.01$. Surface sensible ($H$) and latent (LE) heat fluxes are specified by the following sinusoidal functions:

\begin{align}
H &= SFF_\text{tot} \sin(\Omega t) \quad \text{and} \\
\text{LE} &= (1 - SF)F_\text{tot} \sin(\Omega t),
\end{align}

where $F_\text{tot}$ is the total surface heat flux, SF is the fraction of the total flux that goes into sensible heating, $\Omega$ is the angular speed of the earth, and $t$ is the time from the start of the simulation. Based on COPS surface energy balance observations and real-case simulations of this event (see Barthlott et al. 2011), we estimate $F_\text{tot} = 420$ W m$^{-2}$ and use a horizontally varying SF that increases with terrain height $h$ [SF = 0.25(1 + $h$/$h_m$)], where $h_m$ is the mountain crest height. The Coriolis effect is applied to perturbations from the base state using an $f$-plane approximation with $f = 10^{-4}$ s$^{-1}$.

The domain has a length of $L_x = 200$ km and a depth of $L_z = 17$ km. Rayleigh damping is used over the uppermost 5 km to minimize spurious reflections of gravity waves. The lateral boundaries are periodic to allow for natural ABL development without boundary interference. Although periodic boundaries are unable to radiate gravity wave energy out of the domain, the absence of gravity wave instabilities or a ducting mechanism renders the wave amplitude negligible at these boundaries in the flows considered here. The grid spacing is variable in both directions, with a minimum horizontal spacing of $\Delta x = 100$ m over the central 100 km, which increases to 200 m at the lateral boundaries, giving $N_x = 1500$ grid points. The nominal resolution of the terrain-following vertical grid is $\Delta z = 50$ m from the surface up to 5 km, which smoothly increases from 50 to 200 m between 5 and 10 km and remains fixed at 200 m above, giving $N_z = 175$.

The mountain profile is a symmetric Gaussian function with $h_m = 1.5$ km and horizontal half-width $a_m = 20$ km, which is roughly representative of the southern Black Forest where the observed thunderstorm developed. The initial flow is based on a COPS sounding that was launched at 0800 UTC (1000 local time) from Burnhaupt Le Bas, which is located at the southern end of the Rhine Valley (Fig. 1). This location was chosen over other sounding sites in the region (some of which were closer to the observed storm location) because it lies upstream of the southern Black Forest in the generally southerly flow during the event. The sounding is characterized by a shallow mixed layer near the surface, a stable layer to 1 km, an elevated mixed layer up to 700 mb, a nearly pseudoadiabatic layer up to 220 mb, and a stable stratosphere. The winds are weak and easterly near the surface and then turn southwesterly and increase to 10–15 m s$^{-1}$ in the mid and upper troposphere. This sounding shares key features with previous diurnally forced mountain convection events (e.g., Maddox et al. 1978; Tripoli and Cotton 1989; Zehnder et al. 2006, 2009): large convective inhibition ($\sim 300$ J kg$^{-1}$), steep low- to midlevel lapse rates, weak low-level winds, and a dry midtroposphere. Although the

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig1}
\caption{Skew-T profile of COPS sounding from Burnhaupt Le Bas in the southern Rhine valley at 0800 UTC 15 Jul 2007 (black lines). The idealized thermodynamic sounding used for the simulations is overlaid in gray.}
\end{figure}
CAPE is initially very small, diurnal heating causes it to gradually increase to about 1000–1200 J kg$^{-1}$ by the midafternoon.

To capture the basic characteristics of this sounding and remove extraneous details that might complicate our interpretations, we idealize it using a four-layer temperature profile and a three-layer humidity profile. The former is characterized by an isothermal layer with $T = 296$ K from the surface up to 1 km, a fixed Brunt–Väisälä frequency of $N = 0.002$ s$^{-1}$ up to 3 km, a reversible pseudoadiabatic profile to 11.5 km, and a stratosphere with $N = 0.02$ s$^{-1}$. Following Bryan and Fritsch (2004), the saturated adiabat $T$ and the evolution of properties following the flow. The parcels are initialized at $t = 0$ h at every horizontal gridpoint over the lowest 20 vertical levels (i.e., 1 km), and their evolution is computed online at each large time step of the simulation (and output to file every minute).

Note that the simulations conducted here are not intended to reproduce the observed convection in COPS IOP8b—only 3D simulations with accurate large-scale flow evolution and mountain forcing are appropriate for that. As such, no direct comparisons will be provided between these simulations and the observations from the event. For a more detailed analysis of the predictability of this event in real-case simulations, we refer interested readers to Barthlott et al. (2011).

### 3. Results

#### a. Moist convection

An example of the simulated transition from shallow-to-deep convection is shown by the total hydrometeor field $q_h$ ($= q_l + q_r + q_i + q_s$), where $q_l$, $q_r$, $q_i$, and $q_s$ are the liquid cloud, rain, ice cloud, snow, and graupel water contents, respectively) at eight different times during the U1.5–2 simulation in Fig. 2. In the simulation naming, the first four digits represent the wind speed (“U1.5” for $U_0 = 1.5$ m s$^{-1}$), and the final digit is the ensemble member (ranging from 1 to 8). The ABL top, defined as the lowest vertical gridpoint with $N > 0.01$ s$^{-1}$, is also overlaid for reference.

Along with a remnant cloud at $x = 120$ km, a series of shallow cumuli with tops below 6 km develop while $t = 510–555$ min over the lee slope (Figs. 2a–d). Over $t = 555–585$ min, the updraft at $x = 112$ km strengthens, rises to 6.5 km, and then dissipates (Figs. 2d–f). This is evident from the five 30-min forward parcel trajectories initialized at $t = 570$ min in Fig. 2e. Although these parcels are initially rising and buoyant, they quickly dissipate and spread laterally away from the cloud core. By $t = 585$ min, two vigorous updrafts characterized by relatively large $q_h$ (labeled C1 and C2) rapidly ascend through the cloud mass and merge into a larger updraft that reaches 12 km by $t = 600$ min (Figs. 2f.g). Forward trajectories initialized within C1 indicate direct ascent up to 10–12 km. This cell members are insufficient for a thorough predictability study, it provides some indication of the robustness of deep convection within a given large-scale flow.
reaches maturity by $t = 615$ min (Fig. 2h) and then slowly dissipates as it drifts downwind. Note that the 30-min backward trajectories in Figs. 2e,f illustrate that the pulses of energy that give rise to the cumulonimbus originate within the ABL.

Differences in the moist convection between the three ensembles are illustrated by ensemble-averaged time series of convective mass flux at lower levels ($z = 3.5$ km, denoted $M_{c,3.5}$) and upper levels ($z = 10$ km, $M_{c,10}$) and the averaged precipitation accumulation over the central mountain region over $x = 80–120$ km ($P_{avg}$). The $M_c$ profiles are obtained by horizontally integrating $\rho w$, where $\rho$ is the density and $w$ is the vertical velocity, over all cloudy, ascending grid points at each height. Shallow convection develops the earliest and becomes the strongest in the U0.0 case, then diminishes under increasing

![Fig. 2. Plots of $q_h$ in the U1.5-2 simulation at 15-min intervals during $t = 510–615$ min. The boundary layer top, as defined in the text, is shown by the thick black line. A period of shallow convection during $t = 510–570$ min is followed by the initiation of deep convection during $t = 585–615$ min. Forward (black solid lines) and backward (black dashed lines) 30-min parcel trajectories are overlaid on (e) and (f) to show the origin and evolution of air parcels within the updrafts. The C1 and C2 labels on (f) correspond to two active updrafts that later merge to initiate deep convection.](image-url)
background winds (Fig. 3a). In the U0.0 case, a sharp decrease in $M_{c3.5}$ occurs over $t = 5–6$ h because of convective precipitation, which generates ABL cold pools that temporarily shut down the mountain convergence. This effect is also apparent in the U1.5 ensemble to a lesser degree over $t = 7–9$ h. In all cases, $M_{c10}$ generally decreases late in the day as a consequence of diminished surface heating and weakened mountain convergence. Deep convection, as revealed by $M_{c10}$, is also the most vigorous in the U0.0 case and tails off sharply as the winds are increased, with $M_{c10} = 0$ Mg m$^{-1}$ s$^{-1}$ for all of the U3.0 simulations (Fig. 3b). Precipitation amounts, which increase rapidly during periods of deep convection, are the largest for the U0.0 runs and fall to nearly zero for the U3.0 cases (Fig. 3c).

b. Ensemble variability of deep convection

To illustrate the variability in deep convection across a given ensemble, Fig. 4 shows the ensemble spread of $h_q$, which is defined as the maximum height at which $q_h$ exceeds 0.1 g kg$^{-1}$. This threshold is applied to emphasize active convective clouds with large water contents rather than the thin cirrus anvils that may persist for hours after the decay of deep convection. In the U0.0 case, cumuli first develop at $t = 3$ h and gradually deepen until 5 h. Large ensemble variability occurs between 5 and 8 h because of differences in the timing of convective initiation between the members. Thereafter, the ensemble tightly clusters around $z = 12$ km as a consequence of deep convection in all of the members. By contrast, shallow clouds in the U1.5 ensemble form later in the morning and deepen more slowly over $t = 4–6$ h, and then exhibit persistently large ensemble variability after $t = 6$ h. This large spread reflects broad differences in cloud depth between the different members. Shallow cumuli fail to deepen substantially in the U3.0 case, and the ensemble $h_q$ clusters at 3–6 km. Note that the trends shown in Fig. 4 are insensitive to the $q_h$ threshold over the range from 0.01 to 1 g kg$^{-1}$.

These results are consistent with those of Petch (2004), who simulated diurnal convection in horizontally homogeneous 2D flow, in that the evolution of moist convection may strongly depend on small-amplitude and random perturbations added to the initial ABL flow. Because these perturbations seed turbulent eddies in the ABL, which in turn initiate cumuli, they can nontrivially impact the evolution of moist convection. This is particularly noticeable in the U1.5 ensemble, in which the very existence of deep convection is sensitive to the random seed of the initial perturbations. As a method to distinguish vigorous deep convection from weaker or shallower convective events,
we use a threshold value of $M_c |_{10} = 10^4 \text{ kg m}^{-1}$ (which roughly corresponds to an updraft of 1-km width ascending at $w = 10 \text{ m s}^{-1}$). Using this threshold, only five of the eight U1.5 ensemble members qualify as deep, as opposed to all of the U0.0 members and none of the U3.0 members. A physical explanation for the periods of large spread within the U0.0 and U1.5 ensembles is provided in section 4d.

c. Boundary layer convergence

As a step toward deciphering the differences between the three ensembles, we examine the sensitivity of the thermally induced updrafts to $U_0$ through three “dry” simulations (U0.0-dry, U1.5-dry, and U3.0-dry) that are identical to the U0.0-1, U1.5-1, and U3.0-1 simulations except that $q_v$ is initially set to zero to remove any latent heating effects. Figure 5 shows a running average of vertical velocity $\bar{w}$ and potential temperature $\bar{\theta}$ over a 60-min window centered at $t = 6 \text{ h}$, which removes most of the high-frequency variability (the boundary layer eddies) and exposes just the mean convective core. The updraft is the strongest in the U0.0-dry case, where it develops directly over the mountain top (Fig. 5a). As $U_0$ increases, the updraft weakens and forms farther down the lee slope, and the flows become increasingly asymmetric about the mountain center point, as the very stable near-surface layer is carried farther up the windward slope and down the lee slope (Figs. 5b,c).

Although not shown, simple linear solutions based on Crook and Tucker (2005) strongly underpredict the strength and sharpness of the simulated mountain updrafts in Fig. 5, which reflects that these circulations are strongly nonlinear. Nonetheless, the theory does capture the basic tendency for stronger background flows to weaken the convective core and shift it downwind. This suggests that the same physical mechanism—the ventilation of heat away from the mountain by the mean wind—is responsible for the strong sensitivity to $U_0$ in both linear theory and the simulations.

d. Adiabatic parcel analysis

One might suspect that the more intense convection in the weaker-wind cases is the result of the stronger convective cores more effectively eroding CIN and allowing the moist updrafts to ascend uninhibited. To examine this hypothesis, we calculate CAPE, CIN, and the level of neutral buoyancy (LNB) at each $x$ grid point by lifting surface-based parcels through the vertical column. Conservation of reversible ice-liquid water potential temperature $\theta_{ir}$, as defined by Eq. (23) of Bryan and Fritsch (2004), is used to derive the parcel properties at each level. As in the model governing equations, buoyancy $b$ is defined as

$$b = g \frac{\theta - \theta_\rho}{\theta_{\rho 0}},$$  \hspace{1cm} (3)$$

where

$$\theta_{\rho} = \theta + \frac{R_d q_v}{1 + q_v},$$ \hspace{1cm} (4)$$

$R_d$ and $R_v$ are the gas constants for dry air and water vapor, and the subscript 0 denotes the initial state. Vertical integration of $b$ from the surface to the lowest level at which $b > 0$, or the level of free convection (LFC), gives CIN, and integration of $b$ from the LFC to the highest level at which the parcel $b > 0$, or the LNB, gives CAPE.

Figure 6 presents time series of ensemble-averaged CAPE, CIN, and LNB, averaged horizontally over the central mountain region ($x = 80–120 \text{ km}$), for the three ensembles. CAPE gradually increases over $t = 0–4 \text{ h}$ and then remains roughly level between 1000 and 2000 J kg$^{-1}$ (Fig. 6a). The strong decrease in CAPE in the U0.0 case over $t = 7–10 \text{ h}$ corresponds to the release of moist instability by deep, precipitating convection. CIN generally decreases over the first 4 h to 10–20 J kg$^{-1}$ then remains

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**FIG. 5.** The strength of the thermally induced mountain updrafts, as revealed by $\bar{w}$ over a 60-min window centered at $t = 6 \text{ h}$ in the dry versions of the (a) U0.0-dry, (b) U1.5-dry, and (c) U3.0-dry simulations. The overlaid black contours are $\bar{\theta}$. 

**FIG. 6.** Time series of ensemble-averaged CAPE, CIN, and LNB, averaged horizontally over the central mountain region ($x = 80–120 \text{ km}$), for the three ensembles. CAPE gradually increases over $t = 0–4 \text{ h}$ and then remains roughly level between 1000 and 2000 J kg$^{-1}$ (Fig. 6a). The strong decrease in CAPE in the U0.0 case over $t = 7–10 \text{ h}$ corresponds to the release of moist instability by deep, precipitating convection. CIN generally decreases over the first 4 h to 10–20 J kg$^{-1}$ then remains...
roughly level, except in the U0.0 case where cold-pool formation and ABL stabilization cause it to increase over $t = 7–10$ h (Fig. 6b). Time series of LNB generally mirror CAPE, with a rapid increase to 10 km after which the profiles remain roughly level (Fig. 6b). Again, the notable exception is the U0.0 case, where the LNB decreases over $t = 7–10$ h because of the release of moist instability.

The small differences between the three profiles in Fig. 6 do not reflect the strong sensitivity to $U_0$ that was seen in Fig. 3. Coincident periods of very low CIN ($<10$ J kg$^{-1}$) and high CAPE ($>1000$ J kg$^{-1}$) suggest a strong likelihood for deep convection in all three cases. Using simple scaling arguments such as in Tian and Parker (2003), one can show that boundary layer vertical velocities of 4–5 m s$^{-1}$ should be sufficient to overcome the CIN and allow parcels to reach their LFC. Such updraft velocities are typical of the turbulent eddies over the high terrain (not shown), even in the U3.0 simulations where deep convection fails to develop. Thus, adiabatic parcel analysis is unable to distinguish between the varying outcomes of the three ensembles.

de. Midtropospheric humidity

The limitations of adiabatic parcel analysis result in part from the role of midlevel humidity on suppressing the growth of entraining thermals. The evolution of midlevel humidity and its sensitivity to $U_0$ is illustrated by the midtropospheric moisture budget over the central mountain region, the control volume for which is shown in Fig. 7. This volume is defined by edges of $x = 80$ and 120 km and $z = 3.5$ and 10 km. The fluxes of moisture through the left $F_{ql}$, right $F_{qr}$, bottom $F_{qb}$, and top $F_{qt}$ boundaries are computed by

$$F_{qs} = \int_{s_1}^{s_2} \rho_d q_s \cdot n \, ds,$$

where $s$ parallels the boundary with endpoints of $s_1$ and $s_2$, $n$ is the unit vector normal to the boundary (directed into the control volume), $\rho_d$ is the dry-air density, and $u$ is the wind vector. Conservation of moisture implies that the sum of these boundary fluxes and the precipitation exiting the control volume should equal the time tendency of the storage term $S_q$, which is defined as

$$S_q = \int_{t_1}^{t_2} \int_{x_1}^{x_2} \int_{z_1}^{z_2} \rho_d q_s \, dx \, dz.$$

Time series of the budget terms in Fig. 8 show a much closer connection with convective vigor than the preceding parcel analysis. First consider the U0.0 case, where the boundary fluxes are generally weak in the horizontal (Figs. 8a,b) but large in the vertical (Figs. 8c,d). Vertical transport of ABL moisture through the convective core causes $F_{qb}$ to increase rapidly over $t = 3–5$ h, which then decreases as outflow from precipitating clouds stabilizes the ABL from 4.5 to 6 h and temporarily reduces the mountain convergence (Fig. 8c). Few of these early cumulus towers are able to survive their ascent through the dry troposphere, which limits the moisture flux across the 10-km level over this period (Fig. 8d). Continued surface heating again destabilizes the boundary layer over 6–7 h, causing $F_{qb}$ to recover rapidly and then remain roughly constant until 8 h. A round of vigorous deep convection
over 8–10 h, which generates strongly negative $F_{qt}$ and heavy precipitation (Figs. 8d and 3c), again stabilizes the ABL, and shuts down the mountain convergence for the remainder of the simulation. The moisture storage generally increases during the peak heating period ($t = 3–9$ h) except during the deep convective episodes where precipitation removes large quantities of moisture and suppresses shallow convection (Fig. 8e).

Increased background winds strongly change the budget terms, beginning with an increase (decrease) in $F_{ql}$ ($F_{qr}$) as more ambient moisture flows across the lateral control-volume boundaries (Figs. 8a,b). The progressively weaker convective cores (see Fig. 5) also generate less moisture transport through the bottom boundary (Fig. 8c), which impedes the growth of $S_q$ over 3–5 h. These weakened midlevel moisture anomalies are further depleted by advection across the right boundary, which is shown by the strongly negative $F_{qr}$ in the U1.5 and U3.0 cases after 5 h (Fig. 8b). A sharp decline in deep convection is clearly apparent in Fig. 8d, where $F_{qt}$ is substantially reduced in the U1.5 case and then eliminated altogether in the U3.0 case. The midlevel storage increases over $t = 4–7$ h because of strengthening $F_{qb}$ (this effect is almost imperceptible in the U3.0 case) and then returns back toward its initial values as the outward flux across the right boundary exceeds the upward flux across the bottom boundary over the final 5 h (Fig. 8e).

Note that the large enhancement in $S_q$ prior to the strong burst of convection in the U0.0 cases at $t \approx 8$ h suggests that substantial midlevel preconditioning is needed for intense convection to develop. This one- to twofold enhancement in $S_q$ in Fig. 8e corresponds to an increase in midlevel RH from $\sim 20\%$ to $\sim 50\%$. Such intense preconditioning does not arise in the cases with nonzero $U_0$ because the midlevel moisture is transported away from the mountain by the mean flow.

At first glance, the elimination of simulated convection for $U_0 = 3$ m s$^{-1}$ may seem inconsistent with COPS IOP8b, where deep convection developed in the presence of stronger winds aloft (Fig. 1). However, the southwesterly winds at the time of convection initiation were oriented roughly parallel to the long axis of the Black Forest massif rather than across it, which limited the ventilation of heat and moisture away from the ridge. Upstream radiosonde ascents just prior to convection initiation (not shown) reveal a cross-barrier wind component of less than 2 m s$^{-1}$ averaged over the troposphere. As a result, heat and moisture were still able to accumulate over the ridge axis, which permitted the ventilation of heat and moisture away from the ridge. Deep convection is clearly apparent in Fig. 8d, where $F_{qt}$ is substantially reduced in the U0.0 cases at $t \approx 8$ h.

4. Interpretation and sensitivities

a. The successive thermal mechanism

Inspection of the full suite of simulations reveals a characteristic pattern of convective initiation that is broadly similar to that of the U1.5-2 simulation in Fig. 2. In the midafternoon, after the mountain flow has been preconditioned by the removal of CIN and shallow-cumulus moistening, a sequence of deepening thermals culminates in a vigorous updraft that ascends through the wake of its predecessors and reaches the tropopause. As a result, heat and moisture were still able to accumulate over the ridge axis, which permitted the ventilation of heat and moisture away from the ridge. Deep convection is clearly apparent in Fig. 8d, where $F_{qt}$ is substantially reduced in the U0.0 cases at $t \approx 8$ h.
seen in Fig. 2. An active thermal, here termed the “pre-cursor thermal,” first increases $h_c$ over $t = 565–580$ min. As this cloud begins to dissipate at midlevels, $h_c$ collapses back to the boundary layer top ($t = 580–585$ min). Shortly thereafter, a vigorous new thermal (the “successor thermal”) emerges from the boundary layer, which increases $h_c$ again ($t = 590–600$ min). This updraft, which ultimately reaches the tropopause at $z \approx 12$ km, marks the transition to deep convection.

The above process, which we term the “successive thermal mechanism,” is the dominant convection initiation mechanism in all of the deep-convective simulations. This is illustrated by Figs. 9b–d, which repeat the above analysis for three other simulations from the U1.5 ensemble (U1.5-1, U1.5-4, and U1.5-7). In each case, $t_i$ is defined as the first time when $M_c|_{10}$ exceeds $10^4$ kg m$^{-1}$ s$^{-1}$, and $x_i$ is the center of mass of $M_c$ at $z = 10$ km at $t_i$. All three cases show one or more strong precursor pulses followed by the emergence of a successor pulse 20–40 min later that ascends through the troposphere. Figures 9e–h repeat this analysis again for four members of the U0.0 ensemble (U0.0-1, U0.0-2, U0.0-7, and U0.0-8), showing a similar general pattern to that in the U1.5 simulations. Note that because shallow cumuli are more widespread in the U0.0 ensembles, and because the clouds do not translate horizontally over their lifetimes, $R_i$ is reduced to 5 km in these cases to focus only on the cloud evolution in the vicinity of the deep cell.

b. Parcel trajectories

Detailed insight into the role of saturated wakes for promoting deep convection is provided by parcel analysis of developing clouds in the U1.5-2 simulation. Consider the shallow cumulus at $t = 555$ min, which briefly intensifies but then rapidly decays over the following 30 min (as seen in Figs. 2d–f). Forward trajectories of four parcels initiated at the base of this cloud at $t_p = 552$ min, $x_p = 110$ km, and $z_p = 3$ km (where the subscript $p$ denotes the
parcel initialization) are presented in Figs. 10 and 11. The parcels are selected by first identifying all of the saturated parcels with positive $b$ and $w$ in a small box surrounding $x_p$ ($x_p \pm 2.5$ km) and $z_p$ ($z_p \pm 0.25$ km) and retaining only the four parcels with the largest vertical velocities at $t_p$.

The positions of these parcels over the 45 min after $t_p$ are shown in Fig. 10, and the parcel properties are plotted as functions of height in Fig. 11. These properties include $b$, equivalent potential temperature $\theta_e$ as defined in Emanuel (1994), $q_t$, and the frozen water mixing ratio $q_f (= q_i + q_s + q_g)$. Although three of the parcels (A, B, and C) fail to reach the 6-km level, one parcel (D) does ascend to 10 km (the evolution of this parcel is discussed more below). The buoyancy of all four parcels rapidly decreases between 3 and 5 km, which strongly differs from the larger buoyancy that would arise from adiabatic ascent (Fig. 11a). The $\theta_e$ and $q_t$ profiles also strongly decrease toward their ambient values within this layer (Figs. 11c,d). Because these two quantities are conserved for adiabatic ascent in the absence of precipitation fallout and glaciation (both of which are minimal at this time), entrainment of ambient air is the principal cause for their sharp decrease. The frozen water mixing ratio is generally zero except for a gradual increase over 5–8 km within parcel D (Fig. 11d).

Although parcel D ultimately reaches the tropopause, it first oscillates about the 5-km level before joining the main convective plume (Fig. 10). Its initial ascent terminates at this level after it loses its buoyancy through severe entrainment (Fig. 11a). As it begins to detrain, the associated evaporative cooling induces negative buoyancy and sinking motion (Fig. 10b). Only when this parcel is reentrained into a subsequent updraft at $t \approx 582$ min does it begin to ascend deeper. Note that although the parcel’s $\theta_e$ and $q_t$ are reduced by entrainment over 3–5 km, they still remain significantly larger than the ambient value (Figs. 11c,d). Hence, compared to mixing with the undisturbed ambient flow, the recycling of such parcels into subsequent updrafts reduces the dilution and evaporation associated with entrainment.

As the shallow cumulus in Figs. 10a–c dissipates, successor updrafts ascend through its wake. Forward trajectories of four parcels initialized at $t = 585$ min in Figs. 12 and 13, which are selected identically to those in Figs. 10 and 11 except for a slight downwind shift ($x_p \approx 115$ km), reveal a markedly different evolution than their predecessors. In particular, parcels C and D rise freely to the tropopause with little evidence of mixing, which contrasts with the heavily entrained parcels in Fig. 11.
Below $z \approx 5$ km, the buoyancy within these parcels lies very close to the adiabatic profile (Fig. 13b), and both $\theta_e$ and $q_t$ remain nearly constant, reflecting nearly undilute ascent (Figs. 13b,c). Over 5–6 km, the latent heating associated with a sharp increase in $q_f$ (as the cloud temperature falls below $-5^\circ$C) increases $\theta_e$ and causes $b$ to exceed the adiabatic profile, which assumes a more gradual transition to ice (Figs. 13a,d). Ice formation accelerates precipitation production, which leads to a sharp decrease in $q_t$ above 6 km and further enhances $b$ through a reduction in condensate loading (Figs. 13a,c).

The above analysis provides a typical example of how a rapid succession of thermals, ejected through the quasi-stationary convective core, may lead to a deep updraft that ascends through the saturated wake of its predecessors. The successor updrafts are aided by two key effects: 1) they ascend through moisture-laden air with similar properties, which shields them from the severe dilution and evaporative cooling resulting from entrainment of dry ambient air, and 2) they entrain or ascend through ice particles that formed within the decaying precursor clouds, which leads to rapid glaciation that enhances the latent heating. Analysis of process conversion rates from the microphysics scheme (not shown) indicates that this rapid glaciation is associated primarily with accretion of supercooled liquid drops by these ice particles, not by the freezing of liquid drops or the Hallet–Mossop mechanism.
c. The importance of saturated wakes

To reinforce the importance of saturated wakes for promoting deep convection, we conduct a sensitivity simulation (RMHYD) that is identical to the U1.5–2 case (CTRL) until \( t_r = 579 \) min, just before updrafts C1 and C2 emerge from the boundary layer and initiate deep convection (Figs. 2f,g). At \( t = t_r \), all hydrometeors are removed above \( z = 4 \) km so that the updrafts ascend through an unsaturated (but still strongly moistened) atmosphere. The influence of this cloud removal on the buoyancy of the incipient updrafts is illustrated in Fig. 14, which compares the evolution of \( u_r \) between the RMHYD and the CTRL simulation.

At \( t = 585 \) min, the emerging updrafts C1 and C2 possess similar structures and buoyancies in the two cases, except that the weakly buoyant upper portion of C2 is eliminated in the RMHYD case (Figs. 14a,b). In addition, the ice particles within C1 in the CTRL case, which arise because of sedimentation from the existing midlevel cloud, are completely absent in the RMHYD case. Large differences develop by \( t = 600 \) min, where the CTRL case forms a vigorous, deep, and buoyant cloud but the RMHYD case develops only a weakly buoyant and disorganized cloud mass (Figs. 14c,d). These differences are magnified at \( t = 615 \) min, where the mature storm in the CTRL case spreads laterally at the tropopause while the cloud in the RMHYD case continues to decay (Figs. 14e,f). Evidently, the absence of midlevel saturation renders the cloud in the RMHYD unable to maintain its buoyancy as it ascends through midlevels, which leads to a much shallower and weaker convective storm.

d. Precipitation and ensemble spread

By what mechanisms do small-amplitude initial perturbations produce such large differences in cloud depth among the U0.0 (over \( t = 5–8 \) h) and U1.5 (over \( t = 7–12 \) h) ensembles in Figs. 4a and 4b? Some insight into this question is provided by time series from four members of the U1.5 ensemble in Fig. 15, two of which (U1.5-1 and U1.5-2) develop vigorous deep convection and two of which (U1.5-5 and U1.5-6) do not. Plotted are the lee-side surface precipitation \( P_{\text{lee}} \) and surface buoyancy \( b_{\text{lee}} \), both of which are averaged over \( x = 100–120 \) km where the moist convection occurs. Also plotted are the previously defined \( M_{e|3.5} \), \( S_{\theta_r} \), and \( M_{e|10} \). As seen in Fig. 15a, the convective showers at \( t \approx 7 \) h differ slightly between the realizations as a result of differences in the details of the ABL eddies that trigger the clouds. Heavier
precipitation in the U1.5-5 and U1.5-6 cases leads to increased evaporative cooling and noticeably stronger cold pools (Fig. 15b). Because these cold pools temporarily halt the mountain convergence and suppress the formation of shallow cumuli, \( S_q \) decreases as the existing midlevel moisture anomaly above the mountain is transported downwind (Figs. 15c,d). By the time that the cold pools in the U1.5-5 and U1.5-6 cases are eliminated (\( t = 9 \) h in Fig. 15b), the surface heating is too weak, and the midlevel flow too dry, to support vigorous moist convection. By contrast, rapid erosion of the weaker cold pools in the U1.5-1 and U1.5-2 cases allows shallow cumuli to quickly regenerate (Fig. 15c). The attendant midlevel moistening fosters the development of vigorous deep convection at \( t = 9 \) h (10 h) in the U1.5-1 (U1.5-2) simulations (Fig. 15e).

A similar mechanism is responsible for the large ensemble spread in the U0.0 ensemble over \( t = 5-8 \) h, except that the lack of ambient wind allows the midlevel moisture to remain directly over the mountain region. As a consequence, deep convection readily develops in all of the members once the mountain convergence is reestablished.

### 5. Summary and conclusions

Cloud-resolving, 2D simulations of thermally driven mountain flow have been conducted to study the mechanisms of mountain convective initiation under high pressure conditions. The simulations are based on a well-observed event during the Convective and Orographically Induced Precipitation Study (COPS) field program where an isolated cumulonimbus developed over the Black Forest mountains in southwestern Germany. The background flow in this event was conditionally unstable but strongly inhibited and very dry at midlevels, which presented a hostile environment for cumuli that formed in response to thermally driven mountain convergence. The simulations...
are idealized to reduce computational expense, facilitate conceptual understanding, and explore the sensitivities of the convection to small changes in the background winds ($U_0 = 0, 1.5,$ and $3 \text{ m s}^{-1}$). For each value of $U_0$, an ensemble of eight simulations with slightly different random, low-amplitude initial perturbations is conducted to evaluate the robustness and predictability of the convective response.

In all cases, a well-defined horizontal convergence region (or “convective core”) forms over the high terrain during the morning, which locally weakens the convective inhibition and gives rise to shallow cumuli. Despite large CAPE and small CIN over this region, deep convection does not immediately develop because the cumuli rapidly lose buoyancy through mixing with the dry midlevel flow. A period of shallow convection is required to moisten the midlevel flow in order to reduce the suppressive impact of entrainment on subsequent clouds.

After the mountain environment has been preconditioned by the removal of CIN and shallow-cumulus moistening, the transition to deep convection occurs not through a single isolated updraft but rather through a rapid succession of updrafts ejected by the convective core. Precursor updrafts ascend through the moistened low-level flow and dissipate at mid- to upper levels, leaving saturated wakes behind that provide a favorable

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**Fig. 14.** Evolution of in-cloud $\theta'_p$ (filled gray contours), $q_r$ (solid lines showing the 0.001 g kg$^{-1}$ contour), and $q_f$ (dashed lines showing the 0.001 g kg$^{-1}$ contour) between the CTRL and RMHYD simulations, over $t = 585$–$615 \text{ min}$. As in Fig. 2, the boundary layer top is shown by the thick black line. For clarity, only the $\theta'_p$ contours in regions of $q_h > 0.001 \text{ g kg}^{-1}$ are shown.
environment for subsequent updrafts. Successor updrafts emerging from the convective core rapidly ascend through these wakes and rise to the tropopause, marking the initiation of deep convection. Rather than entraining dry ambient flow, these updrafts recycle saturated air from the existing cloud mass, which shields them from the suppressive effects of entrainment and allows them to maintain their positive buoyancy as they ascend through the midtroposphere. In addition, preexisting ice particles from decaying updrafts seed glaciation through the rapid accretion of supercooled drops, which enhances the latent heating and boosts the cloud buoyancy just above the freezing level (5 km) at temperatures of around $-5^\circC$. As a consequence, the buoyancy of parcels within these updrafts meets or even exceeds that predicted by adiabatic ascent (assuming a gradual transition from liquid to ice). Buoyancy is enhanced further by rapid precipitation fallout at upper levels, which diminishes the condensate loading.

The simulated convection exhibits strong sensitivity to the cross-barrier winds due to two principal effects. First, ambient winds weaken the thermally induced updrafts through ventilation of heat away from the mountain, which reduces the vertical fluxes of mass and moisture at low levels. Second, by transporting moistened midlevel air away from the mountain region, these winds limit the ability of shallow convection to effectively precondition the flow directly above the mountain. As a consequence, deep convection is strongly suppressed for $U_0 = 1.5$ m s$^{-1}$ and completely eliminated for $U_0 = 3$ m s$^{-1}$. Note that the ability of such modest cross-barrier winds to eliminate deep convection should not be interpreted as a universal feature of orographic convection; rather, it is a potentially strong sensitivity in cases with marginal instability and dry midtropospheric flow.

Significant variability in the depth and intensity of the convection is found within the ensembles, which reveal a strong sensitivity to the initial perturbation fields. A particularly large and persistent ensemble spread is found in the $U_0 = 1.5$ m s$^{-1}$ case, where variability in the strength of the ABL cold pools that form in response to midday showers leads to profoundly different outcomes. Heavier showers in some members create stronger and more persistent ABL cold pools, which strongly suppress the mountain convergence and the re-emergence of shallow cumuli. Advection of moisture away from the mountain during this time reduces the midlevel humidity, which inhibits deep convection for the remainder of the simulation. By contrast, the weaker cold pools in the other members are quickly eliminated, which leads to the rapid regeneration of shallow convection and, ultimately, a transition to vigorous deep convection.

Because a number of potentially important physical processes are either absent or inadequately represented in these simulations, continued investigation is necessary to confirm the importance of the mechanisms cited here. In particular, the 2D restriction precludes a faithful treatment of turbulent motions and scale interactions, which demands that the study be extended to 3D. The use of a single thermodynamic sounding profile limits the generality of the results and demands a wider range of parameter exploration. The very simplistic Gaussian mountain shape neglects the strong localized forcing that can arise from flow channeling through valleys and enhanced convergence over isolated peaks. The use of prescribed surface fluxes also neglects the impact of the background winds on the ABL heat uptake and the interaction between the surface forcing and thermally driven circulations.

Work is currently underway to address a number of the above issues by extending the numerical experiments to 3D and broadening the parameter space under investigation. Although the dimension of parameter space is very large for this problem, we focus on certain parameters (e.g., the conditional instability and midlevel humidity)
that intuitively should be expected to strongly influence the solution. An additional focus is the performance of various entraining cloud models (e.g., Raymond and Blyth 1986; Kain and Fritsch 1990; Wu et al. 2009) in predicting the transition from shallow to deep convection, and the determination of whether such models perform acceptably over complex terrain.

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