Numerical Simulation of Episodes of Organized Convection in Tropical
Northern Africa

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ABSTRACT

A large-domain convection-permitting numerical model is used to simulate episodes of deep convection, which are generated during the day over the Ethiopian Highlands and then propagate westward over the eastern and central Sahel region (5°–20°N) of northern Africa. The simulation comprises 12.5 days within the African Monsoon Multidisciplinary Analysis (AMMA) field campaign in 2006. During this period, long-lived precipitation episodes that survived beyond a single diurnal cycle occurred in the lee of the Ethiopian Highlands only every 2–3 days in both the simulation and observations. This contrasts with some other latitudinal corridors in the lee of major topography, such as the central United States, where long-lived heavy precipitation episodes frequently occur on successive nights.

The intermittency of long-lived events for the current case occurs despite regular daily triggering of convection along the upstream orography, and is linked to strong lower-tropospheric stabilization and reduction of daytime surface sensible heat flux due to residual cloudiness in the wake of long-lived precipitation events during the previous diurnal cycle. The vertical shear that helps organize deep convection is also weakened in the wake of the long-lived events by temporary disruptions of the midtropospheric African easterly jet.

The environments of mesoscale convection are presented for the eastern Sahel, a region where most Sahelian convection originates, but about which little is known at the mesoscale. The study describes the potential for early identification of long-lived convection episodes that are likely to have high impact on the central Sahel and West Africa.

1. Introduction

Recent studies have established a relationship between daytime initiation of deep convection over major topography and its movement overnight through adjacent lowlands in midlatitude North America (Carbone et al. 2002; Carbone and Tuttle 2008), East Asia (Wang et al. 2004; He and Zhang 2010), and South America (Pereira Filho et al. 2010). These studies have emphasized the coherence of the propagating rainfall “episodes” often lasting beyond the life cycle of individual mesoscale convective systems (MCSs). Similar relationships exist in parts of the continental tropics, with the most notable example occurring in the Sahel region (5°–20°N) of Africa (Laing et al. 2008, hereafter L08).

Most of the rainfall in the Sahel occurs in MCSs (e.g., Mathon et al. 2002a) and the longer-lasting rainfall episodes are supported by significant convective available potential energy (CAPE) and vertical shear (e.g., Hodges and Thornicroft 1997; Mohr and Thornicroft 2006; L08). Similar characteristics are found in the central United States, where the vertical shear is influenced by the southerly low-level jet. In the Sahel the vertical shear is strongly influenced by the midtropospheric African easterly jet (AEJ; e.g., Parker et al. 2005) and the southwesterly monsoon. Sahelian convection, most of which originates in the central and eastern Sahel, appears to play a role in initiating African easterly waves (AEWs), which intensify as they move westward (e.g., Mekonnen et al. 2006, hereafter M06; Thornicroft et al. 2008; Mekonnen and Rossow 2011) and are sometimes associated with tropical cyclone formation in the eastern Atlantic (Avila and Pasch 1992; Berry and Thornicroft 2005; Lin et al. 2005). Northern tropical Africa is different from the other
continental regions in having greater zonal dimension (∼6000 km) and larger number of major topographic features that can influence the life cycle of precipitation episodes.

Though several studies have documented the intensity, life cycle, and spatial distribution of convection in tropical northern Africa, the existing data network, especially in the eastern Sahel, is insufficient to resolve convectively induced modifications to the mesoscale environment, which may influence precipitation episodes. Organized, propagating convection is triggered along mountain ranges of various sizes in northern tropical Africa (Hodges and Thorncroft 1997; Mohr and Thorncroft 2006; Fink et al. 2006; Thorncroft et al. 2008; L08) and general circulation models do not resolve those features sufficiently well, which contribute to their limited ability to simulate mesoscale convection. Moreover, general circulation models are exposed to errors resulting from their reliance on subgrid parameterizations used to represent deep convection. Herein, we mitigate these shortcomings by using a convection-permitting model over a horizontal domain that spans the majority of tropical northern Africa (Fig. 1) to study 12.5 days of active convection (12–24 August 2006) during the African Monsoon Multidisciplinary Analysis (AMMA) field campaign (Redelsperger et al. 2006). While this approach does not enable analysis of seasonal statistics, it allows novel analysis of possible mechanisms influencing the longevity of successive cycles of topographically generated Sahelian convection.

The simulation design and model configuration are described in section 2. Validation of the simulation is presented in section 3. Also in section 3 are discussions of the environments of episodes of organized convection and composite environments of short-lived and long-lived events. Finally, the results are summarized in section 4.

2. Numerical model and experiment design

Our simulation is a 12.5-day integration of the Advanced Research Weather Research and Forecasting Model (ARW; Skamarock and Klemp 2008) using a single 1200 × 416 horizontal domain (Fig. 1) with 4-km horizontal grid spacing. This horizontal resolution, while not ideal for resolving individual convective cells (Bryan et al. 2003), captures salient mesoscale aspects of convection obviating the need for cumulus parameterization. The vertical grid contains 35 levels and a model top near 0.5 hPa, with enhanced resolution in the planetary boundary layer (PBL), where Δz < 100 m near the surface. Physical parameterizations include the Yonsei University (YSU) PBL (Noh et al. 2003), the Thompson et al. (2008) bulk microphysics, and the Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) and Dudhia (1989) shortwave radiation schemes. ARW is coupled with the Noah land surface model (Ek et al. 2003). The atmospheric initial and lateral boundary conditions are obtained from 6-hourly National Centers for Environmental Prediction (NCEP) Global Forecast System (GFS) analyses on a 1° × 1° horizontal grid.

Though many different mountain ranges can trigger long-lived precipitation episodes in northern Africa, we focus our most detailed analysis in the lee of the Ethiopian Highlands (Fig. 1). Because it is the easternmost topographic feature and initiation point for westward-propagating convection, this region is relatively uncomplicated by the effects of upstream convection, and the
diurnal forcing from the topography is highly regular. Hence, understanding of the initiation of long-lived westward-moving heavy rainfall events may be most achievable in this region.

3. Results

a. Validation and statistics of the simulation

A Hovmöller rain-rate diagram from the simulation (Fig. 2a) indicates afternoon convection beginning regularly in the Ethiopian Highlands (40°E), indicated here by the gray shading, and reaching the MCS maturing region (MR; Fig. 1) as coherent downward-sloping rain streaks overnight. Observed rainfall episodes (Fig. 2b) are derived using the Climate Prediction Center morphing method (CMORPH), a combined passive satellite microwave and geosynchronous infrared (IR) precipitation estimate, available every 30 min with a nominal grid size of 8 km (Joyce et al. 2004). The simulation also produces a few episodes along the slopes of the Darfur Mountains that remained beyond 1 day of their initiation, similar to the observations.

The simulation is diagnosed in terms of the statistical properties of the mesoscale convection (e.g., the duration and frequency of precipitation), rather than a deterministic prediction of individual convective systems. Similarity in the statistics of modeled and observed precipitation (e.g., Fig. 3) lends confidence for using the simulation to diagnose the environments of organized, long-lived convection.

Rain streaks from both the observations and ARW are defined objectively, as in Carbone et al. (2002), by using a 2D autocorrelation function centered at a given longitude–time coordinate that is rotated at 1° intervals through different angles until the linear correlation between the function and precipitation is maximized. These objectively defined rain streaks, depicted as dark gray lines in Figs. 2c,d, each have correlations of 0.35 or greater. In some cases, the episodes are continuous along a streak with varying intensity; while for others, periods of deep convection are intermittent but aligned with a constant phase line.

Simulated (Fig. 2c) and observed (Fig. 2d) midtropospheric winds have a northerly component in the Ethiopian Highlands. However, oscillations to weak southerly flow with a ~4-day period occur farther west. The rain streaks are not clearly related to these meridional wind oscillations, though southerlies consistently follow the longest-lived simulated rain streaks by 12–24 h. Those precipitation episodes are in the northerly phase of the easterly wave, ahead of the trough. The tendency for convection in easterly waves to be ahead of the trough was described by several studies (Carlson 1969; Reed et al. 1977; Diedhiou et al. 1999; Kiladis et al. 2006). Other studies observed enhanced convective precipitation in the south- 11erlies north of 12.5°N (Burpee 1974; Duvel 1990; Mathon et al. 2002b; Fink and Reiner 2003).

The phase speeds of the simulated and observed precipitation episodes are similar (Figs. 3a,b) with respective means of 9.45 and 8.77 m s⁻¹, which are a few meters per second slower than for Sahelian convection averaged over entire seasons (L08). The average longevity and westward distance traveled during our 12.5-day simulation are somewhat greater for CMORPH (14 h and 450 km) than for ARW (11 h and 380 km). ARW (CMORPH) had 10 (12) episodes with a zonal span >1000 km. The span and duration are highly correlated for simulated and observed precipitation episodes, with correlation coefficients of 0.88 and 0.89, respectively (Figs. 3c,d); which is comparable to the correlation for all of northern, tropical Africa (L08). Duration versus distance plots for tropical Africa has a narrow spread compared with the broader spread for midlatitude regions where propagation of organized convection is influenced by fronts or upper-level waves (Carbone et al. 2002).

Time–longitude frequency diagrams of precipitation are computed to determine the zonal progression of rainfall during the diurnal cycle (Fig. 4). Maxima indicate the regular occurrence of precipitation at a particular longitude at the same time of day. The simulated mean diurnal cycle has a similar pattern to the observed diurnal cycle of precipitation. A domain-wide, diurnal oscillation has a maximum centered between 1200 and 0000 UTC, associated with local diurnal heating and propagating convection. The maximum precipitation frequencies (Fig. 4) are in the lee of the Ethiopian Highlands (E) and Darfur Mountains (D), with weaker maxima associated with the Jos Plateau (J), Air Mountains (A), and the mountains of Cameroon (C). The simulation precipitation frequency (Fig. 4a) is greatest over the Ethiopian Highlands, while the frequency in observed precipitation (Fig. 4b) and cold clouds is slightly greater over the Darfur Mountains than the Ethiopian Highlands (L08). In both the simulation and observations, propagating convective systems modulate the diurnal cycle so that precipitation maxima occur as axes across the domain. The result is nighttime and early morning rainfall maxima downstream of high terrain, which is similar to other continental regions (Laing and Fritsch 1997; Carbone et al. 2002; Mapes et al. 2003; Wang et al. 2004).

b. Rainfall intermittency

Similar attributes of the ARW and CMORPH precipitation include the large daily variations in both the longevity of the rain streaks that move from Ethiopian
Highlands into the MR (Figs. 2a,b) and their total rainfall over the MR (Fig. 5). A 2–3-day spacing between the heaviest rain events in the MR is a common feature of both the simulation and observations (Fig. 5) and contrasts with the dominant diurnal periodicity over the Ethiopian Highlands (Figs. 2a,b). These lower-frequency rainfall variations over the MR are consistent with M06 who found a peak variance in satellite-derived rainfall of 2–6 days downstream of the Ethiopian Highlands.

We illustrate the cause of rainfall intermittency over the MR (Fig. 5) with an example from $t = 228$ to 252 h in Fig. 6. On the afternoon preceding a strong system, high equivalent potential temperature $\theta_e$ exists over the MR as convection initiates farther east within the Ethiopian Highlands into the MR (Figs. 2a,b) and their total rainfall over the MR (Fig. 5). A 2–3-day spacing between the heaviest rain events in the MR is a common feature of both the simulation and observations (Fig. 5) and contrasts with the dominant diurnal periodicity over the Ethiopian Highlands (Figs. 2a,b). These lower-frequency rainfall variations over the MR are consistent with M06 who found a peak variance in satellite-derived rainfall of 2–6 days downstream of the Ethiopian Highlands.

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Highlands (Fig. 6a). Furthermore, the minimum negative buoyancy, \( B_{\text{min}} \), of the most unstable parcel (Trier et al. 2011, their Fig. 10) approaches zero (Fig. 6b), consistent with minimal convective inhibition (CIN). Here \( B_{\text{min}} \) is analyzed because, unlike CIN, it is a continuous field. The relationship between CIN and \( B_{\text{min}} \) is analogous to the relationship between the CAPE and the minimum lifted index for air parcels with positive buoyancy. Together, these thermodynamic conditions favor the growth and intensification of the convection into an intense MCS when it enters the MR. However, when the MCS leaves the MR the following afternoon, lower \( u_e \) (Fig. 6c) and smaller (i.e., more negative) \( B_{\text{min}} \) (Fig. 6d) in its wake are deleterious to convection the next night. The precipitation episode on the second day is short lived compared with the longer-lived previous precipitation episode.

The quasi-2-day periodicity in local occurrences of deep convection is consistent with that attributed to similar effects over the equatorial Pacific (Chen and Houze 1997) and offers a possible explanation for the higher-frequency portion of the nondiurnal precipitation variations downstream of the Ethiopian Highlands in M06. A similar cycle may also apply in West Africa where Flamant et al. (2009) found that the influence of a cold pool generated from an MCS persisted as a coherent dynamic and thermodynamic feature over two days. The cold pool cooled the surface considerably and delayed the vertical development of the boundary layer on the day after the MCS.

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**FIG. 3.** Histogram of zonal phase speed (m s\(^{-1}\)) for (a) simulated and (b) CMORPH-derived observed rain episodes. Scatterplots of duration vs distance traveled for (c) simulated rain episodes and (d) CMORPH-derived observed rain episodes. In the bottom right of (c) and (d) are the regression equations for the best fit lines \( y \) and the correlation coefficients \( R^2 \).
Strong vertical shear from 0.5 to 3.5 km above ground level (AGL) is present over the MR in advance of the developing MCS (Fig. 6b) and remains ahead of the mature MCS after it exits the MR the next afternoon (Fig. 6d). This particular precipitation episode, beginning at $t = 228$ h, persists for at least 72 h through the remainder of the simulation (Fig. 2a), consistent with the organizing effects of the strong vertical shear (e.g., Rotunno et al. 1988; Nicholls et al. 1988). The shear vector in advance of the long-lived precipitation episode is directed to the southwest, the region of high $\theta_e$. In contrast, on the following afternoon (Fig. 6d), shear vectors at the leading edge of the MR are directed toward the northwest. Strong vertical shear is more localized over the MR and shear vectors point in various directions and are therefore less supportive of long-lived organized convection. A similar shift in the shear vector before and after a strong MCS was found by Barthe et al. (2010) in their study of MCSs and an AEW over West Africa. In their case, MCSs acted to reduce the monsoon flow and to generate southerlies at midlevels and contribute to the circulation in the AEW trough.

c. Composite environments

The previous example illustrates substantial differences in the thermodynamic and kinematic environments of a long-lived and a short-lived precipitation episode west of the Ethiopian Highlands, which we now extend by constructing composites (for the entire 12.5-day simulation) of these different environments. Long-lived cases are required to have average rain rates exceeding 0.45 mm h$^{-1}$ along a 1° longitudinal slice (33.5°–32.5°E) in the lee of the Ethiopian Highlands from 7° to 16°N and the associated precipitation is required to extend to ~30°E and westward before dissipating. Five episodes satisfied these criteria with three progressing well west...
of 30°E (Fig. 2a). Since afternoon convection triggered regularly in the Ethiopian Highlands during the simulation, the seven days where long-lived criteria are not met were classified as short lived.

The average westward speed of the long-lived events was 10.4 m s\(^{-1}\) compared to 6.9 m s\(^{-1}\) for the short-lived events. The tendency for faster-moving convective systems to be longer-lived agrees with observations of fast- and slow-moving convective lines over the tropical eastern Atlantic (Barnes and Sieckman 1984).

The 1200 UTC (1400 LT) composite environment in advance of long-lived cases has significant CAPE for the most unstable parcel (MUCAPE), with \(B_{\min} > -0.5\) K (Fig. 7b) and strong easterly vertical shear of \(\sim 12\) m s\(^{-1}\) (Figs. 7a,b). This is not the case for the environmental composite of the short-lived cases (Figs. 7c,d), where both the vertical shear and MUCAPE are smaller and \(B_{\min} < -1.0\) K over a wide area. These differences between the two composites are quantified in Figs. 7e,f. While these mean environments do not necessarily resemble individual cases, they capture key differences between the long- and short-lived cases, particularly in the minimum negative buoyancy. For instance, the cumulative distributions of gridpoint values of \(B_{\min}\) over a 400 \times 400 km\(^2\) region within the composite domain (rectangles in Fig. 7) differ markedly between the long- and short-lived cases (Fig. 8). Here the 75th percentile value of the midafternoon \(B_{\min}\) from the five long-lived cases is \(\sim 0.5\) K, which supports local convection initiation (CI). In contrast, the 75th percentile value from the seven short-lived cases is \(\sim 1.8\) K, which presents a much more significant energy barrier to CI. For these short-lived cases only \(\sim 35\%\) of the grid points within the 400 \times 400 km\(^2\) region have midafternoon \(B_{\min} \geq -0.5\) K (Fig. 8).

Composite afternoon environmental soundings are averaged along the red and blue lines in Fig. 7. In constructing the mean profiles, our intent was to identify a particular latitude that best emphasized differences between wake effects from past convection and...
FIG. 7. 1200 UTC (1400 LT) composite simulated environmental fields for (a),(b) long-lived precipitation events, (c),(d) short-lived precipitation events, and (e),(f) the difference between the two. The shading in each is a smoothed rendition of the model terrain height in 0.6-km contour intervals starting with light gray shading of 0.6–1.2 km above mean sea level (MSL). The contoured field on the left is the CAPE of the most unstable parcel (MUCAPE) in 200 J kg$^{-1}$ contour intervals. The contoured field on the right is the minimum negative buoyancy beneath the level of free convection for the most unstable parcel ($B_{\text{min}}$, in temperature units) in 0.4-K contour intervals. The vectors in each plot are the 0.5–3.5-km AGL vertical shear with reference vector at bottom right. The location of the composite domain is indicated by the dashed rectangle in Fig. 1. The inset rectangles indicate the regions over which gridpoint frequencies (e.g., Fig. 8) are calculated. The mean soundings in Fig. 9 are calculated along the red and blue lines.
undisturbed environments.\textsuperscript{1} The profiles (Fig. 9) show pronounced differences between the long- and short-lived cases. The PBL is relatively deep and well mixed in the environment of the long-lived cases and the AEJ near 600 hPa (\textasciitilde 3.5 km AGL) contributes substantially to lower-tropospheric vertical shear. The associated vertical velocity profile is characteristic of undisturbed conditions with weak mean ascent within the PBL and weak subsidence above (Fig. 9b, red). In contrast, the composite environment for the short-lived cases is characteristic of disturbed environments often found in the wake of large MCSs, which themselves occurred the previous night prior to a large majority of the short-lived cases. Here, the vertical velocity profile has lower-tropospheric descent with mid- to upper-tropospheric ascent (Fig. 9b, blue) and is consistent with the sounding (Fig. 9a, blue), which shows a shallow PBL topped by a relatively dry layer near the level of maximum subsidence (\textasciitilde 900 hPa) and a deep moist layer above. Note that the AEJ is disrupted in this environment prior to short-lived events (Fig. 9a, blue), leading to weaker vertical shear.

The mean area-averaged diurnal cycles of both precipitation and thermodynamic and kinematic variables are substantially different over the MR (Fig. 1a) for the long- and short-lived cases (Fig. 10). For the long-lived cases, precipitation increases at \textasciitilde 2000 UTC (2200 LT) and continues on average until after sunrise the next morning, whereas it decreases overnight for the short-lived cases (Fig. 10a). The less favorable vertical shear and thermodynamic conditions during the afternoon for the short-lived cases illustrated in Figs. 7 and 9 persist for most of the diurnal cycle (Figs. 10a–d). The short-lived averages approach the long-lived averages late in the cycle, when the long-lived convection is weakening (Figs. 10a–d).

Applying the Wilcoxin–Mann–Whitney rank-sum test (e.g., Wilks 1995, section 5.3.1),\textsuperscript{2} we find that differences in the area-averaged $B_{\text{min}}$ over the MR between the independent long- and short-lived populations are significant.

\textsuperscript{1} The length of the averaging was motivated by the need to have similar surface elevation at composite locations in order not to introduce artifacts in the lower troposphere. Thus, the lines in Fig. 7 traverse the plains between the Ethiopian Highlands and the Darfur Mountains.

\textsuperscript{2} This method of hypothesis testing was chosen because of potential significant departures from Gaussian distributions in the small long- and short-lived samples.
at the 95% confidence level from 1300 to 1900 UTC (Fig. 11), which immediately precedes the heavy rain over the MR in long-lived cases (cf. Fig. 10a). Though the means for MUCAPE (Fig. 10d) and zonal vertical shear (Fig. 10b) are also substantially different over the MR for the long- and short-lived cases, we are unable to establish statistical significance over sustained 3-h intervals (Fig. 11) due to the relatively broad distributions (compared to those of $B_{\text{min}}$) within these small samples. The distribution of zonal shear over the MR among cases (not shown) is particularly broad, and is likely influenced by synoptic variability during the 300-h simulation in addition to wake effects from organized convection.

The less favorable MUCAPE (Fig. 10d) and $B_{\text{min}}$ (Fig. 10c) found in the mean environment of short-lived cases are associated with colder mean near-surface potential temperatures (Fig. 10e). These lower temperatures are in turn influenced by the greater stabilization in the wake of heavier rain events that typically occurred the preceding night and also by less surface warming during the subsequent diurnal cycle due to residual cloudiness as found by Chen and Houze (1997). This latter effect is confirmed by the smaller daytime-averaged downwelling shortwave radiation at the ground and outgoing longwave radiation in the MR for short-lived cases, which are, respectively, only 87% and 88% of that for long-lived cases.

**Fig. 10.** Diurnally averaged time series of (a) precipitation rate, (b) 0.5–3.5-km AGL difference of the U-component wind, (c) minimum negative buoyancy beneath the level of free convection of the most unstable air parcel, (d) convective available potential energy of the most unstable air parcel, (e) 500-m AGL potential temperature, and (f) 500-m AGL water vapor mixing ratio, each spatially averaged over the MR in Fig. 1, for the composites of long-lived (solid) and short-lived (dashed) simulated precipitation episodes indicated individually in Fig. 2a.

**Fig. 11.** The $p$ values from rank-sum tests (see text) for differences in the area averages over the MR (solid rectangle in Fig. 1) between the populations of long- and short-lived cases. Values underneath the horizontal line indicate statistical significance at the 95% confidence level.
cases (Table 1). The smaller shortwave radiative forcing results in a total daytime-averaged surface flux that is only 88% of that of the long-lived cases. Here, the percentage decrease in sensible heat flux (15%) is greater than for latent flux (10%) and contributes significantly to the lower near-surface temperatures for the environments of short-lived cases.

4. Summary and discussion

In this study we use a 12.5-day, convection-permitting simulation to examine the diurnal cycle of convection and its environment in the lee of the Ethiopian Highlands. This topography marks the initiation zone for many of the long-lived precipitation episodes that traverse the more extensive Sahel region of tropical northern Africa. The simulation produces coherent, westward-moving precipitation episodes that are similar to those of satellite-derived precipitation and previous studies of cold cloud patterns (L08). Statistical attributes of the simulated and satellite-derived rainfall, including the westward phase speed, longevity, and zonal span of precipitation episodes are similar, though average values of the latter two attributes are about ~15%–20% greater in the observations. Maxima in the simulated and satellite-derived precipitation are associated with local diurnal heating combined with orographic forcing and westward-moving convection.

Long-lived, westward-moving rainfall episodes in both the observations and the simulation are separated by 2–3 days. Analysis of an example case together with composite thermodynamic quantities for the entire simulation revealed that this periodicity was largely the result of lower-tropospheric stabilization in the wake of heavy rain events and subsequent residual cloudiness reducing surface sensible heat flux during the following diurnal cycle. These mechanisms are similar to those proposed by Chen and Houze (1997) to explain a 2-day periodicity in local occurrences of heavy rainfall over the warm pool of the equatorial Pacific Ocean. In the current case, the AEJ, which influences the vertical shear to help organize convection, is also disrupted in the wake of heavy rainfall events. Understanding what determines the shear is an important issue that remains to be explored.

The 2–3-day separation between heavy rain events in the lee of the Ethiopian Highlands contrasts with the situation observed in some other continental locations. For example, previous studies (e.g., Fritsch et al. 1986; Tuttle and Davis 2006) have noted that heaviest rainfall events over the central United States in the lee of the Rocky Mountains often occur on successive nights. Trier et al. (2010) demonstrated with idealized simulations that moisture convergence stemming from the interaction of a particularly strong nocturnal low-level jet (LLJ) with a lower-tropospheric baroclinic zone provided critical thermodynamic destabilization that operates on diurnal time scales. The relative weakness of the southwesternly LLJ in the West African monsoon could be contributing to the comparatively longer recovery time for conditions favoring organized convection.

Finally, it is important to recognize that, while potentially illuminating, the current 12.5-day simulation comprises only a single monsoon weather regime. The study simulated the statistical properties and environments of mesoscale convection within a period in which synoptic conditions favor westward-propagating episodes of organized convection—a common occurrence in the climatology (L08). Additional modeling studies are needed to explore convective responses within a broader range of large-scale variability.

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