The Diurnal Cycle of Warm Season Rainfall over West Africa. Part I: Observational Analysis

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(Manuscript received 9 December 2015, in final form 1 August 2016)

ABSTRACT

This study provides an improved understanding of the diurnal cycle of warm season (June–September) rainfall over West Africa, including its underlying physical processes. Rainfall from the Tropical Rainfall Measuring Mission and atmospheric dynamics fields from reanalyses are used to evaluate the 1998–2013 climatology and a case study for 2006.

In both the climatology and the 2006 case study, most regions of West Africa are shown to have a single diurnal peak of rainfall either in the afternoon or at night. Averaging over West Africa produces a diurnal cycle with two peaks, but this type of diurnal cycle is quite atypical on smaller space scales. Rainfall systems are usually generated in the afternoon and propagate westward, lasting into the night. Afternoon rainfall peaks are associated with an unstable lower troposphere. They occur either over topography or in regions undisturbed by nocturnal systems, allowing locally generated instability to dominate. Nocturnal rainfall peaks are associated with the westward propagation of rainfall systems and not generally with local instability. Nocturnal rainfall peaks occur most frequently about 3°–10° of longitude downstream of regions with afternoon rainfall peaks. The diurnal cycle of rainfall is closely associated with the timing of extreme rainfall events.

1. Introduction

Warm season rainfall in the tropics varies greatly on wide-ranging time scales and remains a challenging issue for weather and climate prediction. Here we focus on the diurnal cycle of West African rainfall. A physical understanding of how the diurnal cycle is controlled is crucial for simulating and predicting changes in both mean precipitation and extreme rainfall events. For example, to confidently project how rainfall will change in the future under global warming, climate models are expected to realistically capture the diurnal cycle of rainfall in the contemporary climate. However, previous studies (e.g., Cook and Vizy 2006; Dai 2006; Xue et al. 2010) show that the diurnal cycle of tropical rainfall is poorly simulated by the current generation of general circulation models (GCMs). Even when models produce realistic average rainfall amounts on seasonal time scales, they fail to correctly simulate the diurnal cycle of rainfall (e.g., Dai 2006). This deficiency indicates that GCMs do not correctly represent all of the physical processes that control the diurnal cycle of tropical rainfall.

The purpose of this study is to better understand the diurnal cycle of warm season rainfall over West Africa, including its underlying physical processes. We aim to provide a more general characterization of the diurnal cycle of rainfall at localized spatial scales and to better understand the geographical distribution of the diurnal cycle of rainfall over a broad West African domain. Observed precipitation is used to characterize the diurnal cycle, and atmospheric reanalyses are used to explore the physical processes that control it.

The current understanding of the diurnal cycle of rainfall over West Africa is reviewed in section 2. Observational and reanalysis datasets used in this study are described in section 3. Results are presented in section 4, and conclusions are summarized in section 5. To further advance our understanding of how the diurnal cycle of rainfall is represented in regional climate models...
without using cumulus convection parameterization, convection-permitting simulations over West Africa are analyzed in Zhang et al. (2016, hereinafter Part II).

2. Background

The diurnal cycle of rainfall over West Africa is discussed in several studies that have a global perspective. Dai (2001) examines 3-hourly reports from weather stations around the globe to find that the Sahel region (defined as 10°W–20°E and 10°–15°N) exhibits a nocturnal peak for non-drizzle rainfall and both afternoon and nocturnal peaks for showery rainfall. Dai (2001) also shows that the afternoon rainfall peak is related to local instability as indicated by the presence of large values of convective available potential energy (CAPE). Using satellite brightness temperature as a proxy for rainfall, Yang and Slingo (2001) show that afternoon and evening rainfall peaks are dominant over West Africa and also explore connections with atmospheric instability and the life cycle of mesoscale convective systems (MCSs).

NASA Tropical Rainfall Measuring Mission (TRMM; Huffman et al. 2007) satellite observations have enabled a number of studies of the diurnal cycle over West Africa. Nesbitt and Zipser (2003) examined TRMM Precipitation Radar and Microwave Imager data for 1998–2001 and suggest that, over tropical land, rainfall associated with MCSs has a late evening peak while the intensity of MCSs is greatest in the late afternoon. The peak in non-MCS rainfall is in the afternoon.

In a detailed analysis of sub-Saharan African rainfall using the TRMM Microwave Imager data for 1998–2001, Mohr (2004) finds that the diurnal cycle of rainfall has a single late afternoon–evening peak south of 10°N and two peaks—one in the evening and the other near midnight—north of 10°N. This study also suggests that the frequency and life cycle of MCSs are important influences on the diurnal cycle. Another study using Niger station rainfall observations suggests that the peak hour of rainfall has regional variations (Shinoda et al. 1999).

In a recent study, Pfeifroth et al. (2016) used West African rain gauge data to evaluate several satellite products and found that they capture the diurnal cycle of rainfall and its interannual variations. This study also reveals that, for a site with evening rainfall peak, most satellite products show a delay in the peak time of up to 2 h.

The diurnal cycle of rainfall over West Africa is known to be influenced by the life cycle of convective systems. West African warm season rainfall is linked with convective systems that range from localized, short-lived (1–3 h) systems to long-lasting (6–24+ h) MCSs (Laing and Fritsch 1993; Le Barbé and Lebel 1997; Hodges and Thornicroft 1997; Mohr 2004). Mathon et al. (2002) suggest that more than 90% of the warm season rainfall over the Sahel is contributed by organized convective systems (OCSs), which are a class of MCSs. Their analysis over Nigeria implies that the diurnal cycle of rainfall is associated with the propagation of OCSs that originate over upstream topography. By tracking the life cycle of West African MCSs in infrared satellite images, Mathon and Laurent (2001) show that, over the central Sahel, the frequency of summer MCSs peaks at 1900 local solar time (LST). The MCSs are classified into a four-phase life cycle: initiation, splits, dissipations, and mergers. The frequency of occurrence of MCSs in the initiation phase peaks at 1600 LST, while other phases peak in the late afternoon or early evening. In an analysis of station rainfall data in a wet year and a dry year, Shinoda et al. (1999) conclude that the diurnal cycle of rainfall over Niger is related to the westward propagation of MCSs initiated around the Air Mountains in the afternoon. Laing et al. (2008) also find that the diurnal cycle of West African rainfall is influenced by the zonal propagation of convective systems that are initiated over elevated terrain and propagate westward. Using the TRMM precipitation product, Janiga and Thornicroft (2014) conclude that Sahelian rainfall in the afternoon (early morning) is mainly generated by small (large) convective systems.

In summary, previous studies have examined the diurnal cycle of rainfall over specific West African regions or as an average over large portions of West Africa. The time periods analyzed have been limited to case studies of a single season or climatologies formed by averaging a relatively small number of years. In this paper we build on previous work to further advance our understanding of how and why the diurnal cycle varies spatially across West Africa, examining the geographical distribution of the diurnal cycle of rainfall over a broad West African domain and then focusing on localized diurnal cycles to understand the observed distribution. In Part II, we compare the observed geographical distribution of the diurnal cycle in simulations with a convection-permitting regional model to further advance our understanding of the physical processes that drive the diurnal cycle of precipitation over West Africa and our ability to capture these processes correctly in models.

3. Methodology

To examine the diurnal cycle of rainfall, we use the NASA 3-hourly Tropical Rainfall Measuring Mission precipitation 3B42V7 product (Huffman et al. 2007). TRMM coverage includes all longitudes from 50°S to 50°N with a spatial resolution of 0.25°. Compared with previous generations of remote sensing products for rainfall (e.g., rainfall estimates based on brightness
temperature), the TRMM rainfall product has a much finer spatial resolution with robust statistics, which allows advancements of the characterization of the diurnal cycle of rainfall to finer scales. In addition, the 3-hourly Precipitation Estimation from Remotely Sensed Information Using Artificial Neural Networks (PERSIANN) rainfall data (Sorooshian et al. 2000) at 0.25° resolution are used. One of the advantages of using these temporally continuous products is to avoid having to perform large-scale spatial averaging to remove temporal sampling bias from instruments on low-earth-orbiting satellites.

We analyze the diurnal climatology for June, July, August, and September (JJAS) from 1998 through 2013. JJAS is used for the warm season because the rainfall over West Africa usually starts in June (Hagos and Cook 2007) and ends in October (Zhang and Cook 2014). In addition, the warm season of 2006 is used as a case study since multiyear averaging can obscure the physics of synoptic-scale events. The year of 2006 is selected for our case study because this year coincides with the African monsoon multidisciplinary analysis (AMMA) special observing period (Lebel et al. 2010).

The 3-hourly Modern-Era Retrospective Analysis for Research and Applications (MERRA; Rienecker et al. 2011) is used for atmospheric dynamics fields. At the time of this study, MERRA was the only reanalysis product that supplied atmospheric dynamics fields on 3-hourly intervals over West Africa. To reduce the uncertainty of using reanalysis data, other global reanalyses are also compared with MERRA, including the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR; Saha et al. 2010), the ECMWF interim reanalysis (ERA-Interim; Dee et al. 2011), and the ECMWF reanalysis from their operational forecasts for the AMMA observational campaign with AMMA radiosonde data assimilated (ECMWF-OPERA; Agusti-Panareda et al. 2010). Note that CFSR, ERA-Interim, and ECMWF-OPERA are available at 6-hourly intervals for 0000, 0600, 1200, and 1800 UTC.

The mean diurnal cycle of rainfall rate $R$ is defined as the mean rainfall rate at each 3-hourly interval:

$$
R(h) = \frac{\sum_{d=1}^{N} P(d, h)}{N}, \tag{1}
$$

where $d$ and $h$ are indices indicating the day and the 3-hourly interval, respectively. The quantity $P(d, h)$ is the precipitation rate at $(d, h)$. The $N$ is the number of days in the sample.

The analysis domain (7°–17°N and 7°W–21°E; indicated by the outer bold box in Fig. 1) is chosen to avoid coastal rainfall effects over the West African monsoon region, which is the same as the model domain for the convection-permitting simulations in Part II. This is a large area, covering $3.38 \times 10^6$ km$^2$, or nearly half the surface area of the contiguous United States.

4. Results

a. Geographical distribution of the diurnal cycle of rainfall over West Africa

Figures 2a–h show the percentage of total daily rainfall delivered during each 3-hourly interval in the TRMM climatology (1998–2013). Note that the TRMM rainfall product is at 3-hourly intervals. Since the zonal range of our analysis region is 7°W–21°E, which is within the 3-h time zone centered on Greenwich meridian, UTC is used as the local time of the analysis region. The highest percentages, 25%–50%, generally occur in the late afternoon (i.e., 1800 UTC) or at night (i.e., 2100, 0000, and 0300 UTC). Rainfall percentages are minimal from 0900 until 1200 UTC, with values below 15% in most locations. Both maxima and minima exhibit pronounced spatial variations.

Figure 3 displays the distribution of the peak hour of rainfall in the climatological diurnal cycle. Major features include late afternoon peaks over southwestern Mali, Burkina Faso, Ghana, Togo, and Benin; nocturnal peaks (0000 and 0300 UTC) around the political boundaries between Burkina Faso, Niger, Benin, and Nigeria; late evening (2100 UTC) peaks in western Niger; and afternoon (1500 and 1800 UTC) peaks in the vicinity of Lake Chad. The spatial differences of rainfall peak hour are consistent with Shinoda et al. (1999).
This regionalization is apparent in Fig. 4, which shows the climatological diurnal cycle of rainfall area-averaged over 1° × 1° grid boxes from 7° to 17°N and 7°W to 21°E. A peak is defined as at least 1 mm day⁻¹ higher than its neighbor times. Of the grid boxes analyzed, 98% contain a single diurnal peak. The single peak is either in the afternoon or at night. The remaining 2% of the boxes exhibit double peaks, one in the afternoon and another at night.

Fig. 2. Diurnal distribution of percentage of total daily rainfall at each 3-hourly interval from June–September (JJAS) climatology (1998–2013). Dash–dotted lines indicate the country political boundaries in the area.
This result indicates that, climatologically speaking, most of West Africa exhibits a single diurnal peak of rainfall. This is different from some previous studies. For example, Mohr (2004) concludes that the rainfall cycle north of 10°N over the Sahel has two diurnal peaks, which is an artifact of the large-scale area averaging (10°–15°N, 18°W–32°E) needed when using temporally coarse data. As shown in Fig. 4, the rainfall in this region has a single diurnal peak but with substantially spatial variations of the peak time; two peaks in the climatological diurnal cycle only occur in the large-scale area average.

Different features in the geographical distribution of the diurnal cycle of rainfall can also arise because of time averaging. For this reason, a case study of the 2006 warm season is analyzed. This case study also allows us to more directly relate synoptic events to the diurnal cycle. Figure 5 shows peak hours of the diurnal cycle of rainfall for JJAS of 2006. A similar geographical distribution of the peak hours is found in the 3-hourly PERSIANN rainfall data (Sorooshian et al. 2000; results not shown). The large-scale features are similar to the climatology in Fig. 3, which suggests that the warm season (JJAS) in 2006 is reasonably representative of the climatology. The greater spatial variability is expected because of the shorter averaging period.

Figure 6 shows the JJAS 2006 mean diurnal cycle of rainfall in each 1° × 1° grid box within the region from 7° to 17°N and 7°W to 21°E. Distinct from the climatology, 72% of the small regions show a single peak, while 26% exhibit double peaks. This indicates that the very high percentage (98%) of single peaks seen in the climatology is, in part, a vestige of the multiyear time averaging. However, it is still the case that most regions experience a single diurnal rainfall maximum.

Three regions, indicated in Fig. 3, are defined for more in-depth analysis of the diurnal cycle. These regions cover areas with fairly consistent, uniform diurnal peaks of rainfall that are similar in the climatology (Figs. 3 and 4) and in 2006 (Fig. 5). One region has a late afternoon peak at 1800 UTC and is denoted AF (9°–11.5°N, 3°–0°W). Two domains have nocturnal rainfall peaks. The southern region with a nocturnal peak at 0000 UTC is referred to as SN (9°–11.5°N, 2°–5°E), and the northern region with a nocturnal peak at 0300 UTC is named NN (13.5°–16.5°N, 2°–5°E).

Figure 7 shows a Hovmöller diagram of the TRMM 3-hourly rainfall for JJAS 2006 averaged between 9° and 11.5°N, which includes the AF and SN domains.
Generally, rainfall systems form in the afternoon and propagate westward during the late afternoon and night. This westward propagation includes heavy rainfall episodes. Many of these rainfall events are associated with MCSs, and this westward propagation agrees with the studies of life cycle of MCSs (e.g., Mathon and Laurent 2001; Mathon et al. 2002).

Rainfall systems that propagate into the SN domain (2°–5°E) to create the nocturnal maximum are primarily the result of afternoon rainfall immediately to the east (about 5°–15°E). A few long-lasting rainfall episodes can be tracked back to about 20°E. One event begins in the afternoon of 20 July 2006 at 25°E and lasts until the early morning of 22 July 2006 at 2°E. Another event forms in the afternoon of 1 September 2006 at 27°E and remains strong through 5 September 2006 as it propagates westward across West Africa. Note that, although there is regular initiation of afternoon rainfall in the vicinity of the Ethiopian Highlands (35°–40°E), most of these rainfall systems do not remain intact past 10°E. A Hovmöller diagram averaged between 13.5° and 16.5°N (the latitudes of the NN domain; not shown) displays similar westward-propagating features but with less frequent events and lower rainfall rates compared with Fig. 7 because the overall rainfall is less intense at this latitude range.

b. Afternoon rainfall: The role of local instability

The relationship of local atmospheric instability to the afternoon rainfall peaks is explored by examining the vertical profile of the moist static energy (MSE). MSE is the sum of the sensible, latent, and geopotential heat contents of a parcel:

\[ \text{MSE} = c_p T + L q + g z, \]  

where \( c_p \) is the specific heat of air at constant pressure, \( T \) is the air temperature, \( L \) is the latent heat of water vaporization, \( q \) is the specific humidity, \( g \) is the gravity acceleration, and \( z \) is the geopotential height. MSE increasing with altitude indicates a stable atmosphere. Analysis of the MSE allows one to distinguish between the roles of temperature and moisture variations in changing atmospheric stability properties. MSE analysis is widely used in studying instability associated with rainfall (e.g., Pu and Cook 2012; Neupane and Cook 2013).

Figure 8a displays profiles of MSE anomalies (solid lines) for 0000 (green), 0300 (blue), and 1800 UTC (red) averaged over the AF domain. The anomaly is defined as the difference between the averaged MSE profiles at every 0000, 0300, 1800 UTC, respectively, and the mean of JJAS 2006. The moisture \( L q \) and temperature \( c_p T \) components of the MSE anomalies [see Eq. (2)] are

FIG. 5. Peak hour of rainfall from the JJAS 2006 mean. White contours show elevation (m). Boxes denote the three domains defined in Fig. 3.

FIG. 6. As in Fig. 4, but for JJAS of 2006.
Fig. 7. Hovmöller diagram of TRMM 3-hourly rainfall (mm day$^{-1}$) averaged between 9° and 11.5°N for (a) June and July and (b) August and September of 2006. Gray lines denote the boundaries of the AF (3° and 0°W) and SN (2° and 5°E) domains.
indicated in Fig. 8a by the dot–dashed and dashed lines, respectively.

The diurnal cycle of the MSE anomaly in the AF region is confined below 850 hPa. The afternoon rainfall peak is associated with an anomalous MSE profile (red solid line) that is unstable below 850 hPa and neutral above that level; it is near neutral at 0000 UTC and stable at 0300 UTC from the surface to 700 hPa. The

![Image of Fig. 8](image_url)

**FIG. 8.** (a) Profiles of MERRA total MSE (solid lines), $c_p T$ (dashed lines), and $L_q$ (dot–dashed lines) anomalies averaged over the AF domain at 0000 (green), 0300 (blue), and 1800 UTC (red). (b),(c) As in (a), but for the (b) SN and (c) NN domains. Units are $10^3 \text{m}^2 \text{s}^{-2}$.

![Image of Fig. 9](image_url)

**FIG. 9.** As in Fig. 8a, but for (a) CFSR, (b) ERA-Interim, and (c) ECMWF-OPERA at 0000 and 1800 UTC.
MSE anomalies are primarily associated with changes in both $Lq$ and $c_pT$, as the contribution by $g_z$ is negligible in the lower troposphere (not shown). The contribution from $c_pT$ reflects daytime warming of the land surface. Despite these diurnal variations of surface temperature, diurnal variations in $Lq$ are small at the lowest level (975 hPa) and greatest between 900 and 925 hPa, with positive anomalies at 1800 UTC and negative anomalies at 0000 and 0300 UTC.

The MSE anomaly profiles for the SN (Fig. 8b) and NN (Fig. 8c) regions have similar diurnal variations to the AF region, but in the NN region (Fig. 8c) the $Lq$ anomaly is negative near the surface at 1800 UTC. However, the total MSE profile remains unstable because of the sensible heating term. Although rainfall in these regions peaks at night, the MSE anomaly profiles are not favorable for the initiation of nocturnal convection. Note that the SN and NN domains still have afternoon rainfall, though not as peaks.

Figures 9a–c display the AF domain’s MSE anomaly profiles using the CFSR, ERA-Interim, and ECMWF-OPERA reanalyses. Only 1800 and 0000 UTC are shown because 0300 UTC is not available in these 6-hourly datasets. The vertical distribution of the 1800 UTC profiles of the MSE, temperature, and moisture terms are similar to those in Fig. 8a, although the magnitudes vary. This suggests that the physical processes evaluated here are not sensitive to the choice of reanalysis.

Over the AF region the difference in the sensible heating term between day and night can be explained by the fact that land surface heats (cools) the low-level atmosphere in the afternoon (at night). To understand the diurnal cycle of the latent heating term, Fig. 10a displays the diurnal cycle of low-level specific humidity averaged over the AF domain. The afternoon peaks of moisture at 900 and 925 hPa are associated with smaller vertical gradients, suggesting strong vertical mixing.

Figure 10b shows the diurnal cycle of the planetary boundary layer (PBL) height over AF. The nocturnal PBL is below 975 hPa, while the afternoon PBL is well-developed up to 900 hPa. Therefore, the nocturnal
atmospheric moisture from 975 to 875 hPa is more stratified, while in the afternoon the atmospheric moisture at 900 and 925 hPa is well mixed (Fig. 10a).

In general, the afternoon rainfall peaks over West Africa are associated with unstable atmospheric profiles in the lower troposphere. Based on Figs. 3 and 5 we conclude that the afternoon rainfall peaks are dominant over topographical features and in several regions far removed from elevated topography where local instability dominates.

c. Nocturnal rainfall: The role of propagating convective systems

Nocturnal rainfall over the broader West African region is associated with westward propagating MCSs as indicated by the Hovmöller diagrams in Fig. 7 and previous studies (e.g., Shinoda et al. 1999; Laing et al. 2008).

In this subsection, we examine preferred regions for the initiation of these propagating systems and their pathways across West Africa.

If the nocturnal rainfall in a given region is supported by propagation from a preferred location upstream, then rainfall in that region at the peak hour will be positively correlated with the rainfall in the upstream location with some time lag. Figure 11a shows the correlation coefficient map between the JJAS 2006 time series of 0300 UTC rainfall averaged over the NN domain and the rainfall time series of each grid point at previous (a) 1500, (b) 1800, (c) 2100, and (d) 0000 UTC. As in (a)–(d), but for the SN domain time series of 0000 UTC rainfall correlated with the rainfall time series of each grid point at previous (e) 1500, (f) 1800, and (g) 2100 UTC. Only positive correlation coefficients statistically significant at the 99% confidence level are shown. Elevation is shown as red contours for 400 and 800 m.

Fig. 11. Correlation maps relating the JJAS 2006 time series of 0300 UTC rainfall averaged over the NN domain and the rainfall time series of each grid point at previous (a) 1500, (b) 1800, (c) 2100, and (d) 0000 UTC. (e)–(g) As in (a)–(d), but for the SN domain time series of 0000 UTC rainfall correlated with the rainfall time series of each grid point at previous (e) 1500, (f) 1800, and (g) 2100 UTC. Only positive correlation coefficients statistically significant at the 99% confidence level are shown. Elevation is shown as red contours for 400 and 800 m.
Figures 11e–g show correlation coefficient maps that relate the 0000 UTC rainfall time series averaged over SN with each grid point at the previous 1500, 1800, and 2100 UTC times, respectively. The nocturnal peak over SN is associated with rainfall that is initiated near the Jos Plateau (centered at 9°N and 8°E) on the previous afternoon at 1500 UTC. Similar to the NN region, a westward-propagating pathway occurs. Additionally, for the SN and NN regions, the midnight–early morning rainfall shows no significant correlation with the rainfall...

FIG. 12. JJAS 2006 seasonal mean wind vectors (m s⁻¹) in MERRA in the lower troposphere (a)–(h) from 600 to 950 hPa at 50-hPa intervals. The three boxes are as previously defined.
on the following afternoon. To summarize, nocturnal rainfall over the NN and SN regions is found to be associated with the development of afternoon convection approximately 500 km upstream (i.e., to the east).

Next we examine the low-level flow related to this propagation. Figure 12 shows the wind from 950 to 600 hPa in JJAS 2006. Below 850 hPa, the southwesterly monsoon flow dominates, while above 850 hPa the wind is mainly easterly, which is favorable for the westward propagation of MCSs. The seasonally averaged zonal wind speed does not exhibit a diurnal cycle, and the 650-hPa zonal wind speed averaged between 1500 and 0300 UTC on the next day is 9.5 m s$^{-1}$. Our analysis suggests that in order to produce a nocturnal rainfall peak there needs to be large-scale midtropospheric easterly flow and an afternoon rainfall system that is generated at the appropriate distance upstream. This applies to the NN and SN regions as well as other regions with nocturnal rainfall peaks (Figs. 3 and 5). For example, the nocturnal peaks in the southeastern part of the domain (8°–9°N and 7°–20°E) are associated with afternoon rainfall initiated over the Darfur Mountains of Sudan about 650 km to the east (not shown on this map). The nocturnal peaks at 7°–10°N and 7°–4°W are associated with afternoon rainfall peaks over the AF region.

Generally, nocturnal rainfall peaks over West Africa are associated with rainfall systems propagating westward into a neutrally stable environment. Regions with strong nocturnal rainfall are located 3°–10° of longitude downstream (i.e., to the west) of regions with afternoon rainfall peaks. In contrast, regions with afternoon rainfall peaks are either collocated with topographical features or located far from topography allowing locally generated atmospheric instability to dominate the diurnal cycle.

Applying a 50 mm day$^{-1}$ threshold to the rainfall at 0300 UTC over the SN domain, 12 extreme events (see Table 1) from the summer of 2006 are selected. Of the 12 events, 4 are associated with African easterly waves (AEWs) as indicated by 700-hPa wind and relative vorticity fields (not shown). The lack of association between these extreme events and AEWs agrees with Nicholls and Mohr (2010). The diurnal cycle of rainfall is closely associated with the timing of extreme rainfall events.

Figures 13a–l show the propagation of rainfall from 1500 to 0300 UTC on the following day for each extreme event. Each of these events shows propagating features that relate nocturnal rainfall to rainfall generated on the previous afternoon to the east. Within the NN region, four events in 2006 (Figs. 13a,e,f) are associated with systems that formed on the windward slope that lies along the boundary between Niger and Nigeria, and one event (on 6 August) originated over the Air Mountains of Niger. The 11 August event is a large-scale squall line that formed near 1200 UTC on the previous day. It is associated with nocturnal peaks over both the NN and SN regions. In the SN region, all seven events (Figs. 13c,g–l) derive from rainfall events that were initiated over the Jos Plateau and propagated westward. The overlapping of the 0000 and 0300 UTC rainfall distributions suggests that the system became stationary near midnight. In summary, all of these extreme events show propagation following the pathway identified in Fig. 11. This examination of extreme events related to the nocturnal rainfall peak improves our understanding of mechanisms that determine the diurnal cycle of rainfall.

5. Conclusions

We investigate the geographical distribution of the diurnal cycle of West African warm season (JJAS)

<table>
<thead>
<tr>
<th>Label</th>
<th>Time</th>
<th>Domain</th>
<th>Rainfall rate (mm day$^{-1}$)</th>
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<td>NN</td>
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</tr>
<tr>
<td>(b)</td>
<td>0300 UTC 6 Aug 2006</td>
<td>NN</td>
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<td>(c)</td>
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<td>NN</td>
<td>162.63</td>
<td>Yes</td>
</tr>
<tr>
<td>(c)</td>
<td>0000 UTC 11 Aug 2006</td>
<td>SN</td>
<td>59.98</td>
<td>Yes (same one as in NN)</td>
</tr>
<tr>
<td>(d)</td>
<td>0300 UTC 22 Aug 2006</td>
<td>NN</td>
<td>101.20</td>
<td>No</td>
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<tr>
<td>(e)</td>
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<td>NN</td>
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Fig. 13. Propagation of rainfall during 12 extreme events during JJAS 2006 (see Table 1 for details) over the NN and SN domains. Rainfall contours at the same time are plotted in the same color using intervals of 50, 100, and 200 mm day$^{-1}$. Color legend is red for 1500 UTC, yellow for 1800 UTC, cyan for 2100 UTC, deep blue for the following 0000 UTC, and purple for the following 0300 UTC. Elevation (m) is shaded as a background.
rainfall and explore its underlying physical processes using the satellite-derived TRMM precipitation and atmospheric dynamics fields from MERRA.

The diurnal cycle of rainfall is analyzed in the TRMM climatology for 1998–2013. We also evaluate the 2006 warm season (JIAS) as a case study to facilitate a more detailed investigation on synoptic time scales. The high-resolution TRMM rainfall product improves the understanding of the diurnal cycle of rainfall at finer spatial scales.

The conclusions from this observational analysis are summarized as follows:

- Most regions of West Africa (98% in the climatology and 72% in 2006) have a single diurnal peak of rainfall either in the afternoon (i.e., 1500 and 1800 UTC) or at night (i.e., 2100, 0000, and 0300 UTC). The high percentage of single-peak regions in the climatology is partially due to multiyear time averaging. This finding is in contrast to several previous studies (see section 2) that suggest that the diurnal cycle of rainfall over West Africa is characterized by two peaks. This is a vestige of averaging over a large area.

- Two types of regions experience afternoon rainfall peaks: 1) regions with topographic features and 2) regions far removed from upstream topography (i.e., not within the range that is directly influenced by westward-propagating rainfall systems originated over the mountains). In these regions, local instability processes dominate. A moist static energy analysis is used to show that the afternoon rainfall peaks are associated with unstable atmospheric profiles dominated by diurnal temperature variations in the lower troposphere; boundary layer moisture variations play a minor role.

- Coherent regions with nocturnal rainfall peaks are located 3°–10° of longitude downstream (i.e., to the west) of regions with afternoon rainfall maxima. These rainfall peaks are associated with the westward propagation of rainfall systems but not with local instability.

- The diurnal cycle of rainfall is closely associated with the timing of extreme rainfall events. Applying a 50 mm day\(^{-1}\) threshold, 12 extreme events that occurred at night during the summer of 2006 are examined, and each is found to be associated with an MCS that originated on the previous afternoon to the east and propagated to the west.

An improved understanding of the diurnal cycle of rainfall is important for advancing weather and climate prediction over West Africa. To capture the diurnal cycle of rainfall correctly, climate models need to have an accurate representation of the determining physical processes. In Part II, the diurnal cycle of rainfall produced by convection-permitting simulations is analyzed to evaluate the extent to which the model correctly represents the important physical processes controlling the diurnal cycle of rainfall over West Africa.

**Acknowledgments.** The financial support for this project was provided by NSF Award 1036604. We thank Dr. Karen Mohr and the anonymous reviewers for providing suggestions and comments that improved the quality of this paper. We also thank the Texas Advanced Computing Center (TACC) for providing the high-performance computing resource.

**REFERENCES**


