The Madden–Julian Oscillation’s Influence on Spring Rainy Season Precipitation over Equatorial West Africa*

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(Manuscript received 22 July 2014, in final form 4 June 2015)

ABSTRACT

This paper characterizes the influence of the Madden–Julian oscillation (MJO) on spring rainy season (March–June) convection variability over equatorial West Africa (EWA) and investigates mechanisms of association. It is found that the MJO has a significant impact on convection and precipitation anomalies over the region. Over large portions of EWA, MJO impacts on rainfall constitute a difference on the order of 20%–50% from average daily rain rates for the season. This impact is primarily due to the direct influence of the eastward movement of the MJO convective core into EWA, which is associated with westerly low-level wind anomalies that advect moisture from the Atlantic Ocean to the region. In addition, equatorial Rossby and Kelvin waves triggered by MJO convection anomalies over the Indian Ocean have a significant and systematic influence on EWA spring rainy season precipitation. The Kelvin wave contribution and the relative strength of the direct MJO convective influence compared to that of equatorial wave activity differs from findings of studies that have examined MJO influence on EWA during boreal summer. In addition, MJO is found to influence precipitation extremes during spring rains in a manner that is not observed in summer. Importantly, in this analysis the influences of MJO convection and each of the MJO-associated convectively coupled equatorial waves frequently coincide, reaching EWA approximately 20 days after MJO convection initiates in the Indian Ocean. This coincident timing enhances the total MJO impact on the region, and it also implies that MJO events have potential for prediction of regional-scale convection and rainfall anomalies over EWA in this season.

1. Introduction

The Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972) is the strongest driver of intraseasonal precipitation variability in the tropics (Madden and Julian 1994). The oscillation is a planetary-scale, baroclinic disturbance trapped around the equator (Ventrone et al. 2013) that has a period between 30 and 60 days (Schreck et al. 2013). It appears as an eastward propagating large-scale anomaly in convection, zonal winds, and upper-level

velocity potential (Hendon and Salby 1994). The MJO’s convective signal, which propagates at about 5 m s^{-1} (Zhang 2005), is focused on the Eastern Hemisphere tropics (Madden and Julian 1971, 1972); however, its circulation signal affects the global tropics (Hendon and Salby 1994). These impacts vary by region and season and are stronger in boreal winter than in boreal summer (Wang and Rui 1990; Hendon and Salby 1994).

The nature of MJO impacts on local precipitation have been documented in many regions (e.g., Barlow and Salstein 2006; Hidayat and Kizu 2010; Pai et al. 2011; among many others). For Africa, previous studies have focused on regionally distinct areas such as the highlands and coastal regions of East Africa (Pohl and Camberlin 2006a,b; Berhane and Zaitchik 2014), West Africa (Matthews 2004; Maloney and Shaman 2008; Gu 2009; Janicot et al. 2009; Lavender and Matthews 2009; Mohino et al. 2012), and southern Africa (Pohl et al. 2007). These studies have shown that the mechanisms of MJO influence differ by region and season. Convective variability over the West African monsoon region, for example, is modulated by the migration of the MJO

Denotes Open Access content.

*Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JCLI-D-14-00510.s1>.

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DOI: 10.1175/JCLI-D-14-00510.1

convective center and by Rossby waves excited by MJO anomalies over the Indo-Pacific region (Janicot et al. 2009; Lavender and Matthews 2009; Mohino et al. 2012). In the highlands of East Africa, on the other hand, the MJO influences precipitation variability by modulating temperature advection to the region from the Indian Ocean, and by modifying stability of the lower troposphere and convergence of near-surface winds associated with MJO conditions over the Indian Ocean (Berhane and Zaitchik 2014).

Equatorial West Africa (EWA), like much of the global tropics (e.g., Gadgil 2003; Webster et al. 1998; Wheeler and McBride 2005), exhibits strong intraseasonal variability in precipitation. This variability has significant impacts on rain-fed agriculture, food and water security, and human health (Gadgil and Rao 2000; Sultan et al. 2005; Verdin et al. 2005), affecting the lives of millions of people. But while several studies have examined general aspects of precipitation in the region (e.g., Nguyen and Duvel 2008; Jackson et al. 2009; Dezfuli and Nicholson 2013; Nicholson and Dezfuli 2013), relatively few have focused on intraseasonal variability or on the potential influence of the MJO. Of the studies that have considered MJO influence on the region (e.g., Matthews 2004; Maloney and Shaman 2008; Janicot et al. 2009; Lavender and Matthews 2009; Mohino et al. 2012), all but one (Gu 2009) have focused exclusively on boreal summer. The findings of Gu (2009), however, show that region-wide anomalies in convection and precipitation occur during the passage of intraseasonal convective signals associated with the MJO, suggesting that further study of MJO influence during spring is warranted.

The spring rainy season accounts for up to 45% of annual precipitation in portions of EWA and is a vital cropping season in much of the region. It is also subject to significant interannual variability in total precipitation (Eugène et al. 2012) and to variability in the onset and cessation of rains (Odekunle 2004; Odekunle et al. 2005). This variability has large impacts on the livelihoods and socioeconomic activities across the region (Eugène et al. 2012).

Recognizing the importance of this season, we present a detailed investigation of MJO mechanisms of influence during the spring rains as compared to influence during boreal summer rains. The analysis builds on the work presented by Gu (2009) by including a robust wavenumber–frequency spectral analysis that allows us to distinguish between MJO influences attributable to the direct eastward propagation of the MJO convective center to the region and those attributable to Kelvin and Rossby waves triggered by the MJO in the Indian Ocean. In addition, we perform MJO analyses in a consistent manner for the spring and boreal summer rainy seasons in order to compare mechanisms of MJO

influence across these two seasons. The comparisons are useful for understanding fundamental processes of MJO influence in EWA and contribute to the scientific basis for long-range weather forecasts in both seasons. Finally, we examine the influence that MJO has on precipitation extremes across the region.

2. Data and methods

a. Data

A number of indexing systems have been used to quantify MJO activity. Here we consider several different indices in order to verify the robustness of identified associations with EWA precipitation, as follows:

- 1) Real-time multivariate MJO index (RMM), a combined measure of convection and circulation. It is calculated as the principal component (PC) time series of the two leading empirical orthogonal functions (EOFs) of combined daily mean fields of 850- and 200-hPa zonal winds and outgoing longwave radiation (OLR) averaged over the tropics (Wheeler and Hendon 2004; hereafter WH04). To view the spatial and temporal evolution of the MJO, WH04 developed a two-dimensional phase-space diagram. In this phase-space diagram, strong MJO events move in a large counterclockwise direction around the origin. In contrast, weak MJO variability usually appears as random movement near the origin of the phase-space diagram. RMM is available online at <http://cawcr.gov.au/staff/mwheeler/maproom/RMM/index.htm>.
- 2) OLR MJO index (OMI), a univariate index calculated by projecting 20–96-day filtered OLR onto the daily spatial EOF patterns of 30–96-day filtered OLR (Kiladis et al. 2014). The 20–96-day filtered OLR comprises all eastward and westward wavenumbers while the 30–96-day filtered OLR consists of the eastward wavenumbers only (Kiladis et al. 2014).
- 3) Velocity potential MJO index (VPM) is developed in the same way as WH04 except using 200-hPa velocity potential (VP200) instead of OLR (Ventrice et al. 2013). Both OMI and VPM indices can be accessed from <http://www.esrl.noaa.gov/psd/mjo/mjindex/>.

All three of these indexing systems categorize the eastward propagation of the MJO into eight phases, with each phase corresponding to the geographical position of the active convective center of the MJO (see, e.g., WH04's Fig. 7). Generally, the MJO lasts for about 6 days in each phase. In all three systems the phases correspond to periods when the center of MJO convective activity is over the Indian Ocean (phases 2 and 3), the Maritime Continent (phases 4 and 5), and the western Pacific Ocean (phases 6 and

7). In phases 8 and 1 the MJO circulation signal is in the Western Hemisphere and over Africa. The eight MJO phases are categorized as “strong” when the amplitude of the MJO is greater than 1; otherwise the MJO is categorized as “weak” irrespective of the phase of the MJO.

In addition to these established indices, we calculate an MJO index derived specifically from OLR anomalies in the African domain, following a suggestion raised by the Climate Variability and Predictability MJO Working Group (Waliser et al. 2009) to employ a 20–100-day bandpass filter with 201 weights in extracting intraseasonal variability associated with the MJO. We define this fourth index below.

- 4) First principal component of the EOF of OLR over Africa (PC-EOF1) performed on the 20–100-day bandpass filtered March–June OLR values for the domain 15°S–25°N, 30°W–40°E (Fig. 1b; results were not sensitive to moderate changes in the domain extents). The leading eigenvector from the EOF, which passes both scree (Cattell 1966) and North (North et al. 1982) tests, accounts for 18.5% of the variance and was well separated from second eigenvector, which accounts for 10.5% of the variance. The third eigenvector accounts for 7.6% of the variance. The PCs of the second and third eigenvectors do not exhibit significant associations with the PC of the first eigenvector. Therefore, the second and third eigenvectors will not be considered further. The time series of the first principal component (PC-EOF1) has been used to investigate the evolution of the MJO impacts over the study region in the spring rainy season. As we will show in the results section, positive PC-EOF1 corresponds to phase 1 of the other MJO indices considered in this study, whereas negative PC-EOF1 is phase 5. To compare the strength of the impacts of the MJO in the spring rainy season and in boreal summer, PC-EOF1 was calculated for July–September (JAS) following the same procedure to the calculation of PC-EOF1 in March–June (MAMJ).

Gu (2009) use the principal component time series of the first EOF of 20–100-day filtered OLR anomalies centered over EWA to explore impacts of the MJO on convection and precipitation over EWA in the spring rainy season. We generate PC-EOF1 index in a similar way to that of Gu (2009). This index is used to investigate further the mechanisms in detail by breaking up the total MJO influence into direct eastward propagation of the MJO to the region, and into equatorial convectively coupled Kelvin and Rossby waves. Moreover, this index is used to assess influences of the MJO

on rainfall distribution in the spring rainy season, and to compare and contrast the strengths of the MJO impacts on precipitation over EWA in the spring and summer rainy seasons.

Throughout the analyses presented in this paper we use 2.5° resolution interpolated OLR estimates derived from the Advanced Very High Resolution Radiometers (AVHRR) aboard National Oceanic and Atmospheric Administration (NOAA) polar-orbiting satellites (Liebmman and Smith 1996) as a proxy for deep convective precipitation. Negative (positive) OLR anomalies correspond to positive (negative) precipitation anomalies. High-resolution (0.25°) sea surface temperature estimates come from the National Climatic Data Center (Reynolds et al. 2007; <http://www.ncdc.noaa.gov/sst>). All other atmospheric fields [i.e., geopotential height, wind vector data, pressure vertical velocity (omega), and temperature] and sea level pressure (SLP) are drawn from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) Reanalysis-1 (R-1; Kalnay et al. 1996), which also has a resolution of 2.5°. OLR and reanalysis data are distributed by the NOAA/Earth System Research Laboratory (ESRL) Physical Sciences Division (PSD; <http://www.esrl.noaa.gov/psd/>).

Atmospheric reanalysis is notoriously difficult for Africa, due to lack of observational data and complex atmospheric dynamics. As such, the choice of reanalysis dataset does represent a source of uncertainty in our analysis. In this study, NCEP–NCAR R-1 is used because it has been verified over West Africa using pibal and rawinsonde reports (Nicholson and Grist 2001) and because it has been applied successfully in previous studies over EWA (e.g., Dezfuli and Nicholson 2013; Nicholson and Dezfuli 2013). Furthermore, in our studies of MJO impacts on EWA in boreal summer, we found general agreement between NCEP–NCAR R-1, the NCEP–U.S. Department of Energy (DOE) Atmospheric Model Intercomparison Project phase II (AMIP-II) Reanalysis-2 (R-2; Kanamitsu et al. 2002), and the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011).

The USGS Climate Hazards Group Infrared Precipitation with Station (CHIRPS) merged gauge and satellite product (Funk et al. 2014), which is available at a 0.05° latitude–longitude resolution since 1981, is employed to investigate the anomalies in precipitation over EWA associated with the MJO. CHIRPS is used because it covers a longer time period as compared to other daily precipitation datasets. The dataset is global but has a focus on Africa, and it has already been used in several studies over Africa (Bastiaanssen et al. 2014;

FIG. 1. (a) Climatology of March–June precipitation (mm) from USGS Climate Hazards Group Infrared Precipitation with Station (CHIRPS) merged gauge and satellite product for the period 1981–2013. (b) Loading of the first EOF (W m^{-2}) of 20–100-day bandpass filtered OLR centered over West Africa (15°S – 25°N , 30°W – 40°E) for March–June for 1980–2013. (c) Regionalization of (b) based on the spring rainy season CHIRPS precipitation.

Platts et al. 2015; Simane et al. 2015). Nevertheless, as CHIRPS is a recent product with limited validation studies, it must be applied with some caution. All CHIRPS analyses presented in this paper were also

performed using the Global Precipitation Climatology Project (GPCP; Huffman and Bolvin 2013) daily product for the period of GPCP availability (1997–2014). Results are generally similar and are not shown.

b. Data analysis

First, time-lagged linear regression and zero-lag composites of OLR, precipitation, temperature, SST, vertical motion, geopotential height, and wind and moisture flux anomalies at different levels are computed for MAMJ based on each of the MJO indices described above. These anomalies are used to elucidate the associations between the MJO and convection and precipitation anomalies over EWA in the spring rainy season. Temporal autocorrelation was accounted for in all significance tests through reduced degrees of freedom. We calculated reduced degrees of freedom as follows:

$$N' = N \frac{1 - \rho}{1 + \rho}, \quad (1)$$

where N' is the reduced degrees of freedom and N is the number of temporal data points while ρ is lagged autocorrelation of the MJO index.

In the composite analysis, PC-EOF1 values greater than one standard deviation (1σ) are considered enhanced intraseasonal convective events and PC-EOF1 values less than -1σ are taken as suppressed intraseasonal convective events. Over the study period, in MAMJ, there are 58 enhanced convection cases (655 days) and 64 suppressed convection cases (667 days) using PC-EOF1. Composites are calculated as the average of fields when PC-EOF1 is greater (less) than 1σ (-1σ) minus climatology.

Moisture transport anomalies from the tropical Atlantic Ocean to EWA are explored following Vigaud et al. (2009) and Nnamchi and Li (2011). The moisture flux at each pressure level is calculated by the formula below:

$$\mathbf{Q} = q_{\text{lvl}} \mathbf{v}_{\text{lvl}}, \quad (2)$$

where q is specific humidity and $\mathbf{v}$ is horizontal wind speed. The subscript lvl denotes a specific atmospheric level. Further, following Behera et al. (1999) and Nnamchi and Li (2011), the vertically integrated moisture transport is calculated by vertically integrating Eq. (1), which can be written as

$$\mathbf{Q} = \frac{1}{g} \int_{1000}^{300} q \mathbf{v} dp, \quad (3)$$

where g is the acceleration due to gravity (9.81 m s^{-2}), q is specific humidity, and $\mathbf{v}$ is horizontal wind speed. The values 1000 and 300 indicate the pressure levels (hectopascals) over which the integration is computed. To test the significance of composites, a bootstrapping procedure outlined by Terray et al. (2003) is used. This procedure is useful to overcome

FIG. 2. Wavenumber–frequency spectra of OLR component symmetric about the equator for March–June 1980–2013 summed from 15°S – 15°N divided by the background spectrum. Black lines denote shallow water equatorial wave dispersion curves for equivalent depths of 8 and 90 m. Areas outlined in blue define the filter bands used in this study. [Graphic design after Schreck et al. (2013); see Wheeler and Kiladis (1999) for details on the computation techniques.]

weaknesses associated with the normality assumption of the Student's t test.

Next, we compare the impacts of the MJO in the spring rainy season and in boreal summer by decomposing the MJO influence on tropical OLR into components associated with 1) the propagation of the MJO core convective envelope, 2) equatorial Rossby waves triggered by MJO activity in the Indian Ocean, and 3) equatorial Kelvin waves also triggered by MJO in the Indian Ocean using the wavenumber–frequency spectral analysis procedure presented by Wheeler and Kiladis (1999). The filtering is performed by creating an OLR dataset through an inverse transform that retains only the Fourier coefficients corresponding to the regions of the wavenumber–frequency domain of each signal shown in Fig. 2 [see Wheeler and Kiladis (1999) for details on the calculation techniques]. We then regressed a number of relevant time-lagged variables against PC-EOF1 to investigate the evolution of associations between convection anomalies over EWA and the MJO and to compare their relative importance in the spring rainy season and in boreal summer.

To investigate impacts of the enhanced and suppressed phases of the MJO on precipitation distribution, we started with testing the homogeneity of the

FIG. 3. Composites of March–June unfiltered OLR (W m^{-2}): (a),(b) using OMI, (c),(d) based on RMM, and (e),(f) using VPM and (g),(h) based on PC-EOF1 for (left) phase 1 and (right) phase 5, where PC-EOF1 $> 1\sigma$ in (g) and PC-EOF1 $< -1\sigma$ in (h). Results shown are significant at the 90% confidence level.

region influenced by MJO (Fig. 1b) via objective regionalization techniques described in Badr et al. (2015). As our goal was to divide the region into subregions that respond differently to large-scale forcings of intraseasonal variability, we performed the analysis on daily precipitation of each season using the “regional linkage” hierarchical clustering method, which minimizes correlation between subregions. This analysis yielded two largely independent subregions (Fig. 1c) for MAMJ. Analyses were performed in R using the HiClimR package (Badr et al. 2014). The average intraregional correlations for both regions R1 and R2 in Fig. 1c are statistically significant at the 99% significance level while the correlation between the two regions is insignificant. Then, we analyzed the influence of the MJO on precipitation extremes in each region separately.

To streamline the presentation results, we only present composites based on when PC-EOF1 time series has amplitude greater than 1σ . The analyses are performed using daily data for the both seasons and are repeated separately for each month to examine the intraseasonal variability of the impacts of the MJO on convection over the region. All analyses are performed for the time period 1980–2013 except for those involving SST (1982–2013) and CHIRPS (1981–2013).

3. Results and discussion

a. Links of the MJO to convection anomalies over West Africa

Figure 3 shows composites of unfiltered MAMJ OLR based on each of the MJO indices. Generally, all indices yield similar results. OMI, RMM, and VPM give negative OLR anomalies when the MJO is in phase 1 (Figs. 3a,c,e), which is when the MJO convective envelope is centered over Africa and the western Indian Ocean. Conversely, significant positive OLR anomalies are found during MJO phase 5 (Figs. 3b,d,f). This shows that when the MJO convective envelope is centered over the Maritime Continent, equatorial West Africa experiences anomalous dryness.

In MAMJ, composites of OLR based on PC-EOF1 provide results that are generally similar to the conventional MJO indices, and they are similar to the findings of Gu (2009; see their Figs. 3 and 4e). When the value of PC-EOF1 is greater than 1σ (Fig. 3g), we observe negative OLR anomalies over West Africa, which correspond to anomalous wetness over the region, and positive OLR anomalies in the Maritime Continent associated with MJO convective activity. The negative OLR anomalies shown in Fig. 3 represent from 45% to 70% of the standard deviation of the unfiltered OLR

FIG. 4. (a),(b) Composites of March–June raw precipitation (mm day^{-1}) from 1981 to 2013 and (c),(d) composites of March–June raw precipitation divided by the MAMJ average rain rate (%), where (left) PC-EOF1 $> 1\sigma$ and (right) PC-EOF1 $< -1\sigma$. Shading shows results significant at the 90% confidence level.

over the region covering 5°S–10°N, 10°W–25°E. This demonstrates that the MJO has substantial influence on deep convective precipitation of the region.

When PC-EOF1 is less than -1σ , the region experiences suppressed convection (Gu 2009; Fig. 3h). Therefore, generally speaking, in MAMJ, large positive values of PC-EOF1 correspond to phase 1 of the MJO and large negative values of PC-EOF1 correspond to phase 5 of the MJO as defined by the other indices considered in this study. However, the composites of OLR based on PC-EOF1 give stronger composites over West Africa and weaker values over the Maritime Continent as compared to the other MJO indices (Fig. 3). It is also worth noting that the composites of OLR over West Africa based on the different MJO indices (Fig. 3) have a similar spatial pattern to the loading of the first EOF of 20–100-day bandpass filtered OLR centered over West Africa (Fig. 1b), which is similar to Fig. 3 of Gu (2009).

We compare the impacts of the MJO in the spring rainy season and the boreal summer (JAS) using consistent data and analysis methods. Although we do find generally similar results for all MJO indices for the spring rainy season, this is not the case in JAS. In JAS, we see rather significant differences in OLR composites

based on different MJO indices (Fig. S1 in the supplemental material). In JAS, substantial difference in both EWA and Indian Ocean OLR anomalies are observed between the four different indices (Fig. S1). Comparison of Fig. 3 to Fig. S1 reveals that the choice of MJO index makes relatively little difference for analyses of the spring rainy season but makes a large difference for analyses of boreal summer. Hence the choice of MJO index represents a potential source of uncertainty in studies of MJO influence, and our use of PC-EOF1 cannot be extended to other regions and seasons without careful consideration.

The OLR composites shown in Fig. 3 also suggest that the influence of the MJO on EWA convection is somewhat stronger in the spring rainy season than the boreal summer, a period that has been studied by Janicot et al. (2009) and Mohino et al. (2012). Both our analysis (Fig. 3 and Fig. S1) and comparison with Janicot et al. (2009) and Mohino et al. (2012) indicate that the connection between MJO and EWA convection is stronger during the spring rains.

The influence of the MJO on spring rains over EWA is shown in Fig. 4. The composites confirm the fact that the MJO modulates spring precipitation over large parts of

FIG. 5. Composites of March–June longitude–height cross section of raw circulation, expressed as normalized zonal winds and vertical velocity (see text for explanation), averaged from 8°S to 12°N, from 1980 to 2013, for (a) PC-EOF1 $> 1\sigma$ and (b) PC-EOF1 $< -1\sigma$. Horizontal wind anomalies are in meters per second per standard deviation, and vertical velocity anomalies are in pascals per second per standard deviation, where the deviation for each variable is calculated over the entire cross-sectional domain. Shading shows that vertical velocity (Pa s^{-1} , negative upward) anomalies are significant at the 90% confidence level.

EWA. When PC-EOF1 is greater than 1σ , positive precipitation anomalies are observed over most parts of EWA (Fig. 4a), on the other hand, negative precipitation anomalies prevail over the region when PC-EOF1 is less than -1σ (Fig. 4b). This is consistent with Gu (2009), who finds significant results by regressing TRMM Multisatellite Precipitation Analysis (TMPA) rainfall product against PC-EOF1. Over large portions of EWA, the positive and negative composites have magnitude greater than 1 mm day^{-1} . This constitutes a difference on the order of 20%–50% from average daily rain rates for the season (Figs. 4c,d). The other MJO indices give similar results (not shown). We also checked using GPCP dataset, which covers a shorter time period, and the results are generally consistent.

To investigate MJO associations with zonal circulation patterns affecting EWA in MAMJ, we examine longitude–height cross section of circulation composites (Fig. 5). The two-dimensional wind vectors in the figure are calculated by dividing the actual zonal winds and vertical velocity by their respective standard deviation for the entire cross-sectional domain (8°S–12°N, 20°W–140°E). Therefore, the magnitude of the vectors in Fig. 5 is a measure of the local strength of the circulation relative to that across the cross-sectional domain. As shown in Fig. 5a, when PC-EOF1 is greater than 1σ , there is enhanced upward vertical motion over EWA and

subsidence over the Maritime Continent. On the other hand, when PC-EOF1 is less than -1σ , there is anomalous subsidence over EWA and anomalous ascent over the Maritime Continent (Fig. 5b). Circulation composites using phase 1 and phase 5 of the conventional MJO indices are similar to Figs. 5a and 5b, respectively (not shown). The circulation composites shown in Fig. 5 are physically consistent with the OLR anomalies presented in Fig. 3 and precipitation composites shown in Fig. 4.

To explore moisture advection anomalies associated with the MJO in MAMJ, composites of vertically integrated moisture flux are presented in Fig. 6. As Fig. 6a shows, there is enhanced moisture transport from the equatorial Atlantic Ocean to EWA when PC-EOF1 is greater than 1σ . Conversely, when PC-EOF1 is less than -1σ , the moisture transport to EWA from the equatorial Atlantic Ocean is anomalously negative (Fig. 6b). The vertically integrated moisture fluxes are shown for simplicity, recognizing that much of this signal is driven by near-surface moisture flux.

Since all MJO indices provide similar results for MAMJ, and PC-EOF1 captures maximum MJO influence in boreal summer, from this point on we use only the PC-EOF1 index to examine the mechanisms of MJO influence on convection over EWA, and to compare the relative strengths of the MJO influence over the region in both seasons. As a starting point, Fig. 7 presents

FIG. 6. Composites of March–June vertically integrated moisture transport ($\text{kg m}^{-1} \text{s}^{-1}$), from 1980 to 2013, for (a) PC-EOF1 $> 1\sigma$ and (b) PC-EOF1 $< -1\sigma$. Black vectors indicate that moisture flux anomalies are significantly different from zero at the 90% confidence level in at least one of the directions (meridional or zonal wind).

regression of raw OLR against PC-EOF1 for MAMJ and JAS. The regression maps for MAMJ are similar to the findings of Gu (2009) and the JAS results are generally similar to the findings of Janicot et al. (2009) and Mohino et al. (2012) who consider the June–September season in their analysis. We show both here to facilitate seasonal comparisons and provide a point of reference for wavenumber–frequency spectral analysis.

In Fig. 7 time T_0 refers to zero-lag regressions (i.e., the patterns of OLR correlations for when the MJO signal in EWA is greatest). Figure 7 shows the regression sequence for lags from -20 days through $+10$ days (from $T_0 - 20$ to $T_0 + 10$), with an interval of 5 days. This sequence shows the evolution of OLR associations preceding and following the time of maximum MJO influence on EWA. At T_0 OLR modulation of more than 26 W m^{-2} around the mean is observed (Figs. 3 and 7) in MAMJ, with positive PC-EOF1 values corresponding to enhanced convection over EWA. As shown in Fig. 7, at $T_0 - 20$ and $T_0 - 15$, suppressed convection prevails over EWA in both seasons, whereas from $T_0 - 5$ to $T_0 + 5$ the region

experiences enhanced convection in both MAMJ and JAS. However, we can see from the regression sequence that the MJO impact is stronger in the spring rainy season than in boreal summer. For example, at time T_0 the MJO signal over EWA in MAMJ is up to more than double of the signal observed in boreal summer.

At $T_0 - 20$, we observe that in both seasons positive OLR anomalies start to develop over the western equatorial Indian Ocean and move eastward and strengthen up to $T_0 - 5$. Subsequently, the negative OLR anomalies that start to develop on T_0 over the western equatorial Indian Ocean migrate eastward while increasing in intensity up to $T_0 + 10$. The sequence of linear regressions displayed in Fig. 7 is consistent with the composites shown in Fig. 3 and Fig. S1 for all MJO indices: when convection is suppressed over EWA, we observe enhanced convection over the western Pacific Ocean warm pool region (from $T_0 - 20$ to $T_0 - 15$), and when convection is suppressed over the warm pool region, EWA experiences anomalous wetness (from $T_0 - 5$ to $T_0 + 5$). At $T_0 - 15$ and $T_0 - 10$, we observe that the

FIG. 7. Regression of time-lagged raw OLR against PC-EOF1 for 1980–2013 for (left) MAMJ and (right) JAS. Time T_0 is a zero-lag regression, times from $T_0 - 20$ days to $T_0 - 5$ days are time-lead OLR and winds regressed on PC-EOF1 and from $T_0 + 5$ days to $T_0 + 10$ days are time-lag regressions. Shading indicates values significant at the 90% confidence level.

MJO signal is stronger over the Indian Ocean in MAMJ than in JAS. This has implications for the strength of the equatorial waves triggered in the Indian Ocean basin by diabatic heating anomalies associated with the MJO anomalies over the Indian Ocean and we will investigate this in the next sections.

In both seasons, at $T_0 - 10$ days, the Gulf of Guinea experiences anomalous positive SST (not shown). This is consistent with [Lavender and Matthews \(2009\)](#), who document that positive (negative) SST anomalies associated with the MJO, generally, lead the enhanced (suppressed) convection of the MJO by approximately 10–12 days. However, the coherence and significance of these patterns is weak, and we do not explore this mechanism further in this paper.

Analysis of the propagation of MJO equatorial waves has proved to be fruitful for studies of MJO influence on EWA summer rains ([Janicot et al. 2009](#); [Lavender and Matthews 2009](#); [Mohino et al. 2012](#)), but such analysis has not been performed for the spring rainy season, even though it is a period when MJO activity tends to be stronger. Indeed, a regression sequence of PC-EOF1 against time-lagged 925-hPa wind and geopotential height from $T_0 - 20$ to $T_0 + 10$ ([Fig. 8](#)) shows distinct wavelike patterns. On day T_0 , negative 925-hPa geopotential

height anomalies start to develop over the western equatorial Indian Ocean ([Fig. 8](#)). From [Fig. 7](#) (left panels), we see that negative OLR anomalies begin to develop over the western equatorial Indian Ocean on day T_0 , indicating the onset of the MJO over the western Indian Ocean. This is consistent with MJO dynamics, as the onset of the MJO over the western Indian Ocean is characterized by the development of a negative low-level geopotential height anomaly that leads to enhanced boundary layer convergence, convection, and precipitation over the region ([Lavender and Matthews 2009](#)). These fields have baroclinic vertical structures and change sign around 500 hPa ([Janicot et al. 2009](#)). These patterns point to the potential for the MJO to influence the EWA spring rains through equatorial wave activity.

b. Equatorial waves

Anomalies in convection over the Indo-Pacific warm pool can trigger dry eastward-moving Kelvin and westward-moving Rossby waves ([Gill 1980](#)). [Matthews \(2004\)](#), [Janicot et al. \(2009\)](#), [Lavender and Matthews \(2009\)](#), and [Mohino et al. \(2012\)](#) have documented that during boreal summer West African convection is modulated by diabatic heat anomalies associated with

FIG. 8. Regression of unfiltered geopotential height [geopotential meters (gpm)] and wind vectors (m s^{-1}) at 925 hPa against PC-EOF1 for MAMJ 1980–2013. Time T_0 is a zero-lag regression, times from $T_0 - 20$ days to $T_0 - 5$ days are time-lead geopotential and winds regressed on PC-EOF1 and from $T_0 + 5$ days to $T_0 + 10$ days are time-lag regressions. Shading indicates geopotential height anomalies significant at the 90% confidence level. Black vectors indicate values significantly different from zero at the 90% confidence level in at least one of the wind components (meridional or zonal wind).

the suppressed and active phases of the MJO over the Indo-Pacific warm pool. Diabatic heating over the region results in a dry Kelvin wave that is associated with positive equatorial tropospheric temperature anomalies, easterly wind anomalies at low levels, and descent at the wave front. Moreover, it excites a dry Rossby wave to its west that is associated with a couple of twin cyclones that straddle the equator in low levels, with westerly anomalies over the equator and descent to the west of the cyclone.

Diabatic cooling, on the other hand, triggers a dry Kelvin wave that is associated with negative equatorial tropospheric temperature anomalies, westerly wind anomalies at low levels, and ascent at the wave front. In addition, it excites a dry Rossby wave that is associated with a couple of twin anticyclones straddling the equator at low levels, with easterly anomalies over the equator and ascent to the west of the anticyclone (Janicot et al. 2009; Mohino et al. 2012).

We note that these dry Kelvin waves only become free to propagate on their own when MJO convection

dissipates. The convectively coupled Kelvin waves associated with propagation of the MJO are forced by the MJO convective core and are tied to its eastward propagation. Propagation of these convectively coupled waves is evident in propagating OLR anomalies (Straub and Kiladis 2003; Straub et al. 2006; Sobel and Kim 2012). Moreover, the convectively coupled waves are complex and not easily explainable as the dry waves since their scales do not correspond to that expected from the linear theory of dry waves (Kiladis et al. 2009; Wheeler and Nguyen 2015).

Fields plotted in Figs. 7 and 8 show evidence of this MJO-associated equatorial Kelvin–Rossby wave pattern. As shown in the left panels of Fig. 7, about 15 days before the enhancement of convection over EWA ($T_0 - 15$), there is a positive OLR anomaly over the Indian Ocean, indicative of the suppressed phase of the MJO. The Kelvin wave response is visible in Fig. 8 as a tongue of positive geopotential height anomaly that grows eastward along the equator, resulting in strengthened low-level westerly winds over the tropical Pacific. Notably, the

FIG. 9. Regression of MJO-filtered OLR against PC-EOF1 for 1980–2013, for (left) MAMJ and (right) JAS. Times from $T_0 - 20$ to $T_0 + 10$ days comprise a lagged-regression sequence, as in Fig. 7. Shading indicates values significant at the 90% confidence level.

geopotential height anomaly crosses the Atlantic Ocean and migrates over Africa, with wind anomalies reaching EWA at $T_0 - 5$. This influence appears to be specific to the spring rainy season; Janicot et al. (2009) found that the geopotential height anomaly does not propagate beyond 50°W in boreal summer, and that Kelvin wave activity has little influence on EWA rainfall in that season.

Figure 8 also shows some evidence of an equatorial Rossby wave response to Indian Ocean OLR anomalies. This is visible in easterly low-level wind anomalies and a horseshoe-shaped pattern of positive geopotential height anomalies symmetric about the equator visible in the western Indian Ocean at $T_0 - 10$. However, in the regression of raw data shown in Fig. 8, the Rossby wave features are not as clearly visible as the Kelvin wave. This can be partly attributed to the interaction of the Rossby wave response with the Kelvin wave propagating from the Atlantic Ocean and Africa to the western Indian Ocean.

At T_0 active MJO convection develops over the western Indian Ocean, and this convective center strengthens and extends eastward through $T_0 + 5$ and $T_0 + 10$. This convective activity results in negative geopotential height anomalies due to diabatic heating. The Kelvin wave associated with this atmospheric pattern is accompanied by strengthened easterly anomalies over the

entire tropical Pacific Ocean, as can be seen in Fig. 8 at $T_0 + 5$ and $T_0 + 10$. This negative geopotential height anomaly propagates to the Atlantic Ocean and Africa resulting in divergent easterly wind anomalies over the eastern Atlantic Ocean and West Africa. This pattern results in increased subsidence over the region and dry anomalies prevail over EWA. The westward-moving cyclone associated with the Rossby wave enhances this subsidence.

To quantify the relative contributions of the convectively coupled waves to the MJO influence on MAMJ EWA convection and to compare the strength of the impacts of these waves in the spring rainy season and in boreal summer, we present regression of PC-EOF1 against OLR filtered to the characteristic wavenumber and frequency of MJO, Rossby waves, and Kelvin waves (boxes in Fig. 2) in Figs. 9–11. Figure 9 shows the MJO-filtered OLR signal regressed against PC-EOF1 for both seasons. We can observe that in MAMJ, the figure retains much of the signal seen in Fig. 6, indicating that a considerable amount of the MJO influence on west equatorial African springtime precipitation is directly attributable to the migration of the MJO center of convection. This contrasts with boreal summer, when the direct influence of MJO convection on West Africa is comparatively modest. Janicot et al. (2009) and our own

FIG. 10. Regression of time-lagged Rossby-filtered OLR against PC-EOF1 for 1980–2013, for (left) MAMJ and (right) JAS. This regression sequence is shown from $T_0 - 20$ to $T_0 + 10$ days, as in Figs. 7 and 9. Shading indicates values significant at the 90% confidence level.

analysis (Fig. 9, right panels) suggest a weak impact, while Mohino et al. (2012), using different data and methods, find a direct impact that is more pronounced but still not as strong as our results for the spring rains. Even so, the magnitude of the MJO-filtered OLR anomaly in MAMJ in Fig. 9 is weaker than in the nonfiltered OLR data shown Fig. 6, suggesting that the direct impact of MJO convection is not the sole mechanism of MJO influence during the spring months.

The Rossby-filtered OLR, which has a westward propagating signal, is shown in Fig. 10. Its maximum anomaly over EWA coincides with the maximum signal of the MJO-filtered OLR which occurs at T_0 , but the Rossby-filtered signal is considerably weaker (cf. Figs. 9 and 10, noting the difference in color scale) in MAMJ. In JAS, both MJO-filtered and Rossby-filtered OLR give similar results but while the MJO-filtered OLR values are generally centered along the equator, the Rossby-filtered results propagate centered along about 10°N . This is because the Indian summer monsoon triggers Rossby waves which propagate over North Africa (Janicot et al. 2009). The mechanism of Rossby wave generation differs between the two seasons: in boreal summer, in addition to the MJO anomalies over the Indian Ocean, the active/break cycle of the Indian

monsoon influences initiation of Rossby waves that propagate westward over North Africa (Janicot et al. 2009). In March–June the Rossby waves appear to be initiated over the equatorial Indian Ocean basin as a result of the diabatic heating anomalies caused by the enhanced and suppressed phases of the MJO over the region (Figs. 7 and 10). We observe a positive OLR anomaly in MAMJ over East Africa associated with the Rossby wave at $T_0 + 5$ (Fig. 10), which could be one reason for the observed split of negative OLR signal going from T_0 to $T_0 + 5$ days (Fig. 7).

The regression time sequence of Kelvin-filtered OLR (Fig. 11) is also weaker than the MJO-filtered signal in MAMJ, but it does show clear evidence of a Kelvin wave influence on OLR over EWA. Studies of boreal summer have found that these equatorial convectively coupled Kelvin waves do not significantly impact convection over the region (Janicot et al. 2009; Mohino et al. 2012). Our analysis also agrees with previous studies that the Kelvin wave does not have significant impacts on EWA convection in boreal summer (Fig. 11, right panels). The Kelvin-filtered OLR signal in MAMJ in Fig. 11 is consistent with the Kelvin wave–associated geopotential height features shown in Fig. 8, and indicates that the Kelvin influence extends across the Atlantic Ocean and

FIG. 11. Regression of time-lagged Kelvin-filtered OLR against PC-EOF1 for 1980–2013, for (left) MAMJ and (right) JAS. This regression sequence is shown from $T_0 - 20$ to $T_0 + 10$ days, as in Figs. 7, 9, and 10. Shading indicates values significant at the 90% confidence level.

into EWA during the spring rainy season. Of course, because of possible spectral leakage on the filtered results, the contribution of the MJO, Kelvin and Rossby signals could be larger than the results shown.

The sum of the MJO-, Rossby-, and Kelvin-filtered OLR signals for both seasons is shown in Fig. 12. This sum is qualitatively similar to the unfiltered OLR regression pattern shown in Fig. 7, with a slightly lower magnitude. This study shows that in addition to the eastward-propagating MJO signal, a convectively coupled westward-propagating equatorial Rossby wave and an eastward-moving Kelvin wave are needed to explain the overall impact of the MJO on convection over equatorial West Africa in March–June. In boreal summer, the MJO impact over equatorial Africa is weaker than what is observed in the spring rainy season. Moreover, while the Kelvin wave triggered by the MJO in the Indian Ocean has significant impacts on convection in the spring rains, its impacts in the summer season are insignificant.

In the boreal summer, Lavender and Matthews (2009) found that MJO-associated diabatic heating (cooling) anomalies over the warm pool result in positive (negative) midtropospheric temperature anomalies that propagate to West Africa by dry Kelvin and Rossby waves and stabilize (destabilize) the troposphere over West Africa, resulting in enhanced subsidence (ascent).

However, in the spring rainy season, the anomalies in convection are already strong before the temperature anomalies reach the region (Gu 2009; Fig. S2 in the supplemental material). At $T_0 - 5$, we observe from Fig. S2 that there is already significant anomalous upward vertical motion and Figs. 7 and 12 also show significant anomalous OLR anomaly at $T_0 - 5$, which is consistent with the regression field of omega at $T_0 - 5$ (Fig. S2). This suggests that the tropospheric temperature anomalies that propagate to the region have a relatively small effect on the relationship between the MJO and EWA convection in this season.

It is known that the MJO undergoes strong interannual and intraseasonal variability in strength and location (e.g., Hendon and Salby 1994; Zhang 2005). For this reason, we further investigate the evolution of the MJO influence on EWA convection on a month-by-month basis within both seasons. As shown in Fig. 13, in MAMJ, MJO impacts on EWA OLR are strongest in March and decrease as the season progresses. The composite field of omega exhibits a similar pattern (not shown). This tendency likely reflects the seasonal evolution of the MJO. By June, the MJO shows northeastward migration in the Indian Ocean basin (Figs. 13g,h) that is consistent with the link between MJO activity and the Indian monsoon noted in previous studies (e.g., Lawrence and Webster

FIG. 12. Regression of PC-EOF1 against time-lagged sum of MJO-, Rossby-, and Kelvin-filtered OLR for 1980–2013, for (left) MAMJ and (right) JAS. This regression sequence is shown from $T_0 - 20$ to $T_0 + 10$ days, as in Figs. 7 and 9–11. Shading indicates values significant at the 90% confidence level.

2002). The Indian monsoon, in turn, has the potential to modulate West African convection, but the mechanisms of these teleconnections require their own, independent analysis (Janicot et al. 2009). In the JAS season, however, MJO impacts on EWA do not exhibit clear differences from month to month as they do in the spring rainy season (not shown).

c. MJO impacts on precipitation extremes

To investigate the impacts of the MJO on precipitation distribution over EWA, we first divide EWA into relatively homogenous regions and then analyze rainfall relationships with the MJO in each region separately. As shown in Fig. 14, the MJO has strong influences on precipitation extremes in the spring rainy season. Moreover, in the spring rainy season, the influence of the MJO on extremes differs systematically between the two regions shown in Fig. 1c. The enhanced phase of the MJO increases the probability of increased precipitation over region R2 as shown in Fig. 14a. In contrast, the suppressed phase mainly suppresses precipitation over region R1 (Fig. 14b). In region R2, the enhanced phase of the MJO increases daily rainfall by 41% but the suppressed phase of the

MJO reduces daily average rainfall by only 15% (Fig. 14a). On the other hand, over region R1, we observe from the box plot in Fig. 14b that the suppressed phase of the MJO decreases daily rainfall by 46%, while the enhanced phase of the MJO increases rainfall over the region by only 4%. In JAS such differences are not observed (not shown). The results in Fig. 14 suggest that the MJO has important implications for precipitation extremes and prediction of flash floods and droughts in MAMJ. Moreover, while the enhanced phase of the MJO is more important over region R2, the suppressed phase of the MJO is more important over region R1. This shows that the enhanced phase of the MJO can have stronger impacts in some regions and the suppressed phase may be of more importance in other regions. We performed the same evaluation for JAS—regionalization followed by analysis of precipitation extremes—and found no MJO impacts on rainfall extremes in that season. Previous studies have found that the MJO influences the African easterly jet (Matthews 2004; Lavender and Matthews 2009) and African easterly waves (Ventrice et al. 2011) in boreal summer. Analyzing the MJO's impacts on and interactions with synoptic and regional

FIG. 13. Composites of unfiltered OLR (W m^{-2}) for (a),(b) March, (c),(d) April, (e),(f) May, and (g),(h) June using PC-EOF1 time series for (left) PC-EOF1 $> 1\sigma$ and (right) PC-EOF1 $< -1\sigma$.

systems in the spring rainy season can help us to better understand why EWA responds differently to the MJO in spring and summer and why MJO has different impacts on precipitation extremes in different regions within EWA.

4. Conclusions

This study examined mechanisms of MJO influence on spring rainy season (March–June) convection in EWA. It also compared the strength of the impacts of the MJO over EWA in the spring rainy season and in boreal summer. To that end, wavenumber–frequency spectral analysis was employed to isolate the influences of the MJO convective center and of Rossby and Kelvin waves triggered by MJO activity in the Eastern Hemisphere. The analysis shows that the primary mechanism of MJO impact on EWA in March–June is the eastward migration of the MJO convective center into the region. This convective center modulates low-level westerlies, which advect moisture from the Gulf of Guinea and the eastern Atlantic Ocean to the region. In MAMJ, when the MJO convective envelope moves to the Maritime Continent, EWA experiences low-level divergent winds and enhanced subsidence. As a result, in this phase anomalous dryness prevails over EWA.

In addition, in MAMJ, MJO anomalies over the Indian Ocean excite convectively coupled Rossby and Kelvin waves which reach EWA at the same time with the eastward-moving MJO convective center. This

coincident timing enhances the total MJO impact over the region during the spring rainy season. Our results also suggest that the MJO influence on EWA—both direct and via wave propagation—is stronger during the spring rainy season (March–June) than in boreal summer (July–September). Within the March–June season the influence is strongest in March and weakens as the season progresses. The waning of the MJO influence in late spring and summer is a product of general weakening of MJO dynamics in boreal summer (Hendon and Salby 1994), the competing influence of Indian monsoon-generated teleconnections during the summer months, and evolving atmosphere and SST background conditions in EWA.

We have also shown that the MJO has strong impacts on precipitation distribution over EWA in the spring rainy season. The impact of the MJO on MAMJ precipitation extremes could be useful in predicting flash floods and droughts.

The present study has focused exclusively on characterizing and explaining the influence of MJO on the EWA spring rainy season and comparing the strength of the MJO influences in the spring and summer seasons. The results, however, have clear implications for long-range weather forecasting in the region. First, our results indicate that the MJO has a strong influence on spring rains relative to summer rains, such that applications to forecasting are likely to be more meaningful in the spring season. This is an important point, since most previous studies of the MJO

FIG. 14. Distribution and box-and-whisker plots of precipitation during enhanced-, suppressed-, and non-MJO days for the homogenous regions shown in Fig. 1c for (a) region R1 and (b) region R2.

influence on EWA have focused on summer. Second, the analysis shows that anomalous increase (decrease) in MJO-associated convection in the Indian Ocean basin precedes significant reduction (enhancement) of West African convection by approximately 2 weeks (Figs. 7 and 12). This provides a target time horizon and predictor region for statistical forecast systems, and our analysis of mechanisms of communication can inform development of dynamical forecast models. Third, the fact that MJO activity and EWA precipitation anomalies are associated with a 2-week lead raises the possibility of combining existing predictive models of the MJO, which have

useful skill that extends out to about 25 days (Love and Matthews 2009; Seo 2009; Kang and Kim 2010; Rashid et al. 2011; among others) with an EWA-specific model of MJO influence to extend MJO associated rainfall forecast over the region beyond 25 days. At this lead time forecast information has the potential to influence water resource decisions, agriculture management, and disaster response.

Much additional work is required to connect the atmospheric analyses presented in this paper to actionable forecast information. But the utility of incorporating MJO in long-range precipitation forecasts has been

demonstrated in other regions (e.g., Leroy and Wheeler 2008; Jones et al. 2011a,b; Johnson et al. 2011), and the magnitude and predictability of the MJO influence on EWA holds significant promise for this application.

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