Modulation of Daily Precipitation over East Africa by the Madden–Julian Oscillation*

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(Manuscript received 13 November 2013, in final form 29 March 2014)

ABSTRACT

Spatiotemporal variability in East African precipitation affects the livelihood of tens of millions of people. From the perspective of floods, flash droughts, and agriculture, variability on intraseasonal time scales is a critical component of total variability. The principal objective of this study is to explore subseasonal impacts of the Madden–Julian oscillation (MJO) on tropospheric circulations affecting East Africa (EA) during the long (March–May) and short (October–December) rains and associated variability in precipitation. Analyses are performed for 1979–2012 for dynamics and 1998–2012 for precipitation. Consistent with previous studies, significant MJO influence is found on wet and dry spells during the long and short rains. This influence, however, is found to vary within each season. Specifically, indices of MJO convection at 70°–80°E and 120°W are strongly associated with precipitation variability across much of EA in the early (March) and late (May) long rainy season and in the middle and late (November–December) short rainy season. In the early short rains (October) a different pattern emerges, in which MJO strength at 120°E (10°W) is associated with dry (wet) spells in coastal EA but not the interior. In April the MJO influence on precipitation is obscured but can be diagnosed in lead time associations. This diversity of influences reflects a diversity of mechanisms of MJO influence, including dynamic and thermodynamic mechanisms tied to large-scale atmospheric circulations and localized dynamics associated with MJO modulation of the Somali low-level jet. These differences are relevant to problems of subseasonal weather forecasts and climate projections for EA.

1. Introduction

The Madden–Julian oscillation (MJO), which is a 30–60-day oscillation centered around the equator, is responsible for the majority of weather variability in the tropics (Madden and Julian 1994). The MJO appears as an eastward propagating large-scale system in convection, zonal winds, and upper-level velocity potential (Hendon and Salby 1994). The system usually develops in the western Indian Ocean, and precipitation anomalies are recognizable as it propagates eastward to the western Pacific Ocean. When it reaches the cold waters in the

eastern Pacific, it becomes nondescript. However, precipitation usually reappears as it reaches the tropical Atlantic Ocean and Africa (Madden and Julian 1971, 1972).

The MJO is strongest in winter and weakest in summer (Wang and Rui 1990; Hendon and Salby 1994). Notably, however, no matter whether it is winter or summer, the MJO influences rainfall in a number of regions in the tropics and extratropics (Jones 2000; Paegle et al. 2000; Higgins and Shi 2001; Carvalho et al. 2004; Jones et al. 2004; Barlow et al. 2005; Donald et al. 2006; Lorenz and Hartmann 2006; Jeong et al. 2008; Wheeler et al. 2009; Zhang et al. 2009; Pai et al. 2011, among many others). Pohl and Camberlin (2006a,b; hereafter PC06a and PC06b) identify equatorial East Africa (EA) as a region in which the MJO can influence intraseasonal precipitation. They diagnose an MJO influence on precipitation in both the long rains (March–May) and the short rains (October–December) for selected regions in Kenya and northern Tanzania, with an observed contrast of influence between highland and coastal areas. They attribute the MJO influence and the intraregional contrast to a suite of mechanisms related to deep convection, moisture advection, and stratiform precipitation.

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-13-00693.s1>.

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DOI: 10.1175/JCLI-D-13-00693.1

FIG. 1. Climatology of TRMM precipitation (mm day^{-1}) and wind vectors at 850 hPa (m s^{-1}) from NCEP-R1. Precipitation values less than 0.5 mm day^{-1} are suppressed. (left) Long rains for (a) March, (c) April, and (e) May and (center) short rains for (b) October, (d) November, and (f) December. The box in (a) shows the study region. (right) Map showing area of (a)–(f).

The identification of an MJO influence in EA is both intriguing and potentially quite valuable. EA is a topographically diverse region and one of the most meteorologically complex regions on the African continent (Spinage 2012; Cook and Vizy 2013). Precipitation variability on interannual, interseasonal, and intraseasonal time scales has profound and extensively documented impacts on rain-fed agriculture, pastoralism, food and water security, and human health (Epstein 1999; Funk et al. 2005; Verdin et al. 2005; Bowden and Semazzi 2007; Funk et al. 2008; Ummenhofer et al. 2009; Anyah and Qiu 2012; Lyon and DeWitt 2012; Cook and Vizy 2013). While many studies have addressed challenges of explaining and predicting climate variability on seasonal and interannual time scales (Nicholson and Kim 1997; Indeje and Semazzi 2000; Mutai and Ward 2000; Black 2005; Hastenrath 2007; Owiti et al. 2008; Funk et al. 2008; Ummenhofer et al. 2009), relatively few have addressed intraseasonal variability on time scales that are potentially explainable, and perhaps predictable, based on MJO.

It is well known that the long rains and short rains differ in their sensitivity to large-scale climate drivers and in the

characteristics of precipitation (e.g., Camberlin et al. 2009). In addition, each season exhibits systematic differences in rainfall patterns between the early, middle, and late season (Fig. 1). These seasons are transitions between winter and summer monsoons (Hastenrath 2007) and correspond to the period when the intertropical convergence zone (ITCZ) crosses the equator in its south–north and then north–south migrations, respectively (Mutai and Ward 2000; Camberlin and Philippon 2002). The ITCZ modulates the northeast trades blowing during the southern summer and the southeast trades during the northern summer (Asnani 1993, 2005). Variability in the characteristics of the ITCZ is closely associated with variability in rainfall of the region (Gitau 2011).

For the long rains, several studies have indicated that the teleconnections linked to variability also differ across the season (Camberlin and Philippon 2002; Zorita and Tilya 2002), suggesting that atmospheric processes associated with precipitation at the beginning and end of the season are not the same (Camberlin and Okoola 2003). For this reason, authors of previous studies have recommended that studies of interannual variability

consider each month of the long rainy season separately (Camberlin and Philippon 2002).

Here, we apply this reasoning to an analysis of the MJO influence on EA, using the Climate Prediction Center's (CPC's) operational MJO index and the all-season real-time multivariate MJO index (RMM) from the Centre for Australian Weather and Climate Research. The overarching objective of the study is to explore impacts of the MJO on tropospheric circulations affecting EA during the long and short rains and associated changes in precipitation on intraseasonal time scales. In this respect our analysis builds on the work presented in PC06a and PC06b, but for a larger geographic extent and more recent period, and with the analysis carried out for each calendar month, individually, during both rainy seasons. These differences allow for detailed exploration of intraregional and intraseasonal variability in the MJO influence on EA. In addition, we employ multiple datasets in the analysis and explore a number of mechanisms not specifically identified by PC06a and PC06b. The paper is organized as follows: Section 2 describes data and methods, followed by results and discussion in section 3. Finally, a summary and conclusions are offered in section 4.

2. Data and methods

a. Data

We use multiple datasets to study associations of the MJO with precipitation and tropospheric circulation. The precipitation dataset used in this study is the Tropical Rainfall Measuring Mission (TRMM) 3B42 Multisatellite Precipitation Analysis (TMPA), version 7. The dataset has a horizontal resolution of $0.25^\circ \times 0.25^\circ$ latitude–longitude (Huffman et al. 2010). Previous studies have shown that TMPA captures variability of precipitation in East Africa reasonably well, although some versions of the data have exhibited a bias in the magnitude of estimated precipitation rates (e.g., Dinku et al. 2007; Li et al. 2009; Habib et al. 2012). The version 7 multisensor product used in this study has not been evaluated in peer-reviewed publications, but its behavior is similar to earlier products, with some evidence that biases in highland regions have been reduced.

Interpolated outgoing longwave radiation (OLR) estimates derived from the Advanced Very High Resolution Radiometer (AVHRR) onboard National Oceanic and Atmospheric Administration (NOAA) polar-orbiting satellites (Liebmann and Smith 1996) were employed to examine MJO associated changes in patterns of deep convection. Negative OLR anomalies tend to correspond to positive precipitation anomalies, while positive OLR anomalies tend to correspond to negative precipitation anomalies.

Atmospheric fields [i.e., wind vector data, pressure velocity (ω), temperature, humidity, and precipitable water] and sea level pressure (SLP) were drawn from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (NCEP-R1; Kalnay et al. 1996). The wind vectors and temperature are considered as “most reliable” as they strongly depend on instrumental measurements, while ω and relative humidity are considered to be “quite reliable” since they rely more on general circulation model parameterization (Pohl and Camberlin 2006b). Both the wind vector and OLR datasets are available at $2.5^\circ \times 2.5^\circ$ latitude–longitude resolution and were obtained from the website of the NOAA/Earth System Research Laboratory (ESRL) Physical Sciences Division (PSD; <http://www.esrl.noaa.gov/psd/>). For purposes of comparison, we repeat our analyses using SLP and atmospheric fields drawn from the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim), which is the latest global atmospheric reanalysis produced by the ECMWF (Dee et al. 2011). The dataset replaces the 40-yr ECMWF Re-Analysis (ERA-40) and addresses several difficult data assimilation problems encountered during the production of ERA-40 (Dee et al. 2011). For detailed information about ERA-Interim products, the reader is referred to Dee et al. (2011). Sea surface temperature (SST) data were also acquired from the NOAA/ESRL PSD high-resolution (0.25°) analysis product. For more details about this dataset, the reader is referred to Reynolds et al. (2007).

The MJO indices used in this study are the CPC MJO index (Chen and Del Genio 2009) and the RMM (Wheeler and Hendon 2004, hereafter WH04). The CPC MJO index is generated by first applying an extended empirical orthogonal function (EEOF) analysis to pentad velocity potential at 200-hPa for ENSO-neutral and weak ENSO winters (November–April) during 1979–2000 (Xue et al. 2002; Barrett and Leslie 2009). The first EEOF consists of 10 time-lagged patterns. Then, 10 MJO indices, centered at 20° , 70° , 80° , 100° , 120° , 140° , and 160° E, and 120° , 40° , and 10° W, are constructed by regressing the daily data onto the 10 patterns of the first EEOF. Positive (negative) values represent suppressed (enhanced) convection. Each index is normalized by dividing by its standard deviation (Barrett and Leslie 2009). Several previous studies have used the CPC MJO index for analyses of MJO process and impacts (e.g., Chen and Del Genio 2009; Ridout and Flatau 2011; Del Genio et al. 2012; Straub 2013, among others). (The indices and their details are available on the CPC website at http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_mjo_index/mjo_index.shtml.)

The daily real-time multivariate MJO indices (RMM1 and RMM2) of WH04 are 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 OLR averaged over the tropics (15°N–15°S). WH04 categorized the eastward propagation of the MJO into eight phases, each corresponding to the geographical position of its active convective center (see their Fig. 7). These phases constitute a full MJO cycle that is strong in the Indian Ocean and decays over the central Pacific. On average, each phase lasts for about 6 days. WH04 developed a two-dimensional phase space diagram, with RMM1 and RMM2 as the horizontal and vertical Cartesian axes, which is used for viewing the spatial and temporal evolution of the MJO. In this phase-space representation, strong MJO events move in a large counterclockwise direction around the origin, while weak MJO variability usually appears as random movement near the origin.

Phase 1 denotes the period when the center of convective activity is over Africa. In phases 2 and 3, the convective envelope of the MJO is in the equatorial Indian Ocean; phases 4 and 5 correspond to the period when the MJO's convective envelope is in the Maritime Continent, and phases 6 and 7 correspond to the period when it is in the equatorial Pacific Ocean. The square root of the sum of the squares of RMM1 and RMM2 represents amplitude of the MJO. When the amplitude of the MJO is greater than 1, the eight phases are categorized as “strong” MJO phases; otherwise, the MJO is categorized as “weak” irrespective of the phase of the MJO. (RMM indices are available online at <http://cawcr.gov.au/staff/mwheeler/maproom/RMM/index.htm>.)

Analyses that involve precipitation are constrained by the availability of TRMM satellite data, which starts in 1998. As a result, precipitation analyses cover the period 1998–2012. OLR and dynamical analyses are presented for the modern satellite record 1979–2012, while the SST data cover the period from 1982 to 2012. To test the stability of MJO associations over time we repeated all 1979–2012 analyses using data only for 1979–97 and data only for 1998–2012. Results for these two time periods are consistent at seasonal scale and for most months. Small differences between the 1979–97 and 1998–2012 periods are noted in the results section where they are relevant.

b. Data analysis

Combinations of linear correlations and composites are employed to explore associations between MJO and precipitation in EA and corresponding changes in

tropospheric circulation. Composite and correlation analyses are performed at pentad scale for each calendar month of both rainy seasons in order to capture sub-seasonal variability. Pentads from 2–31 March, 1–30 April, 1–30 May, 3 October–1 November, 2 November–1 December, and 2–31 December are considered for the months of March, April, May, October, November, and December, respectively. MJO composites for the CPC indices are constructed using all pentads with CPC index amplitude equal to or greater than one and above, a threshold that has been used in previous studies (e.g., Chen and Del Genio 2009; Barrett and Leslie 2009). This results in between 25 and 43 composite pentads per month for the TRMM period (1998–2012). A total of 90 (204) pentads were available for each month (six per year) for 1998–2012 (1979–2012). In addition, wind and vertical velocity are analyzed at daily resolution for CPC pentads with strong MJO convection or subsidence to investigate whether the anomalies are change of strength of the prevailing motion or actual reversals. Composite figures in this paper show the difference between enhanced MJO convection and suppressed MJO convection; that is, in each month the composites are the mean of pentads with MJO index less than or equal to negative one minus pentads with MJO index greater than or equal to one.

All analyses were repeated using daily RMM indices to verify the robustness of the results obtained employing the CPC MJO index. When using the RMM index, in each month, days with MJO index of amplitude one and above are used. Results for CPC and RMM indices are overwhelmingly similar, so we focus on CPC results for simplicity. (For all figures showing CPC results we provide the equivalent RMM figures in the supplementary material.)

To calculate composites, we first compute the long-term monthly mean for a given variable for each month as the average of all the values in each month. Composites of all variables considered are computed for each calendar month based on the MJO indices as

$$F'(x, y, t) = F(x, y, t) - \bar{F}(x, y), \quad (1)$$

where the left-hand term is the pentad anomaly (daily for RMM), the first term on the right is the value of a variable on a given pentad (day) employing the CPC (RMM) index, and the last term on the right is the monthly mean of the variable considered.

Composites of OLR, SLP, vertical motion, and wind vector anomalies at different levels are calculated for each MJO index. These anomalies are examined to elucidate the physical mechanisms by which the MJO impacts rainfall on monthly time scales.

To investigate changes in components of the thermodynamic balance, we employ the hydrostatic thermodynamic energy equation given by

$$\frac{\partial T}{\partial t} = -\mathbf{V} \cdot \nabla T + S_p \omega + \frac{J}{C_p}, \quad (2)$$

where T is temperature, $\mathbf{V}$ is horizontal wind vector, S_p is the static stability parameter, C_p is the specific heat of dry air, and J denotes diabatic heating. In Eq. (2), the left term is tendency, while the first term on the right is horizontal temperature advection. Static stability is proportional to the vertical gradient of temperature, so $S_p \omega$ is the adiabatic term that represents the vertical advection of temperature and the effect of adiabatic warming and cooling with vertical motion. The diabatic heating term J/C_p is calculated as a residual.

Moist static energy H composites are also calculated at each grid point; H is found using

$$H = C_p T + gZ + l_v q, \quad (3)$$

where C_p is the specific heat of air at constant pressure, T is air temperature, g is gravitational acceleration, Z is geopotential height, l_v is latent heat of vaporization, and q is specific humidity. Lower-tropospheric buoyancy is quantified using moist static instability, which is calculated as moist static energy (MSE) at 1000 hPa minus saturation moist static energy at 700 hPa ($H_{1000} - H_{s700}$; Seager et al. 2003). Saturation moist static energy H_s is calculated in the same manner as H [Eq. (3)], but saturated specific humidity is used in place of specific humidity. Throughout much of the study region the surface lies above 1000 hPa, but anomalies of T , Z , and q at 1000 hPa and at the surface exhibit very similar patterns, so the metric can still be used to diagnose the stability of the lower troposphere (McHugh 2004).

Composites of moisture flux divergence are calculated using

$$\text{MFD} = q \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) + u \frac{\partial q}{\partial x} + v \frac{\partial q}{\partial y}, \quad (4)$$

where MFD is moisture flux divergence, q is specific humidity, and u and v are zonal and meridional wind vectors, respectively. Also, $u \partial q / \partial x + v \partial q / \partial y$ represents horizontal advection of specific humidity and $q(\partial u / \partial x + \partial v / \partial y)$ denotes the product of specific humidity and horizontal mass divergence. In all analyses that involve calculations of gradients, centered difference techniques are used.

For analyses that involve wind speed, vertical motion, temperature, or sea level pressure fields, NCEP-R1 and

ERA-Interim are both employed to confirm the robustness of findings. The datasets provided similar results in all cases, and NCEP-R1 is used in the figures because it has been used in many previous studies in the region (e.g., Mutai and Ward 2000; Camberlin and Okoola 2003; McHugh 2004; Hastenrath 2007; Hastenrath et al. 2007; Lyon and DeWitt 2012). To test the significance of correlation coefficients, a two-tailed t test is used. In the composite analysis, a procedure outlined by Terray et al. (2003) is used. This procedure is useful to overcome drawbacks associated with the normality assumption of the Student's t test.

3. Results and discussion

a. MJO-related anomalies

The region and magnitude of MJO influence in EA varies from month to month as shown in Figs. 2–5. The region of significant association in each month follows the northward and southward migration of the ITCZ as depicted in Figs. 3 and 4. For this reason, we calculate correlations between the CPC MJO indices and pentad precipitation for geographic boxes centered on the area of greatest correlation in each month (Fig. 2). Correlations vary smoothly with MJO phase: enhanced rainfall is experienced when the MJO convective center is between 20° and 140°E and dry anomalies prevail when the MJO is located in the region from 140°E to 10°W. The strength of these correlations is greatest for MJO indices at 70°–80°E and 120°W. We focus primarily on 80°E and 120°W MJO indices for the subsequent analyses.

The spatial distributions of precipitation correlations summarized in Fig. 2 are shown in the form of composite maps for MJO indices at 80°E and 120°W in Fig. 3. For comparison, we have also shown composites of precipitation based on RMM phase 3, which corresponds to MJO's convective envelope over the Indian Ocean, in Fig. 4. Composites of precipitation employing the RMM index and the CPC index are generally consistent. However, the composite results obtained using the RMM index are weaker than those found using the CPC index because the composites using RMM index are calculated by deducting the climatology for each month from the average of the strong MJO events (RMM amplitude of one and above), whereas the composites using the CPC MJO index are calculated by deducting the average of strong MJO subsidence (index value of one and above) from the strong MJO convection (index value of -1 or less). This difference in analysis is a product of differences in the way that the CPC and RMM indices are calculated.

FIG. 2. Correlations of area average precipitation over EA with each MJO index. Regions used for each month are $12^{\circ}\text{--}2^{\circ}\text{S}$, $34^{\circ}\text{--}38^{\circ}\text{E}$ for March; $6^{\circ}\text{S--}5^{\circ}\text{N}$, $30^{\circ}\text{--}38^{\circ}\text{E}$ for April and May; $2^{\circ}\text{S--}8^{\circ}\text{N}$, $30^{\circ}\text{--}38^{\circ}\text{E}$ for October; $4^{\circ}\text{S--}4^{\circ}\text{N}$, $32^{\circ}\text{--}38^{\circ}\text{E}$ for November; and $10^{\circ}\text{S--}2^{\circ}\text{N}$, $32^{\circ}\text{--}38^{\circ}\text{E}$ for December. Solid lines are correlations with concurrent MJO indices, while dashed lines are correlations with MJO indices at a lead of two pentads. Diamonds show correlations that are significant at the 90% confidence level using the reduced degrees of freedom resulting from autocorrelation.

The composites confirm the fact that the association between precipitation and these MJO indices is widespread in November–December, March, and May, covering large portions of the study region. The relationship is weaker in October and April. The EA coast is also affected by MJO at 80°E and 120°W to some degree

(Figs. 3 and 4), but in the coastal zone another MJO influence appears in October, when a strong association with MJO indices at 10°W and 120°E is evident (Fig. 5). This October MJO association is distinct from the highland MJO influence in location and timing, suggesting that a different mechanism of influence may be active.

FIG. 3. Composites of precipitation (mm day^{-1}) based MJO indices at (a)–(f) 80°E and (g)–(l) 120°W for 1998–2012, for (a),(g) March, (b),(h) April, (c),(i) May, (d),(j) October, (e),(k) November, and (f),(l) December for 1998–2012. Color shading shows results significant at the 90% confidence level. Maps of correlations at 90% significance level are nearly identical and are not shown.

FIG. 4. Composites of precipitation (mm day^{-1}) based on RMM phase 3 for 1998–2012, for (a) March, (b) April, (c) May, (d) October, (e) November, and (f) December. Color shading shows results significant at the 90% confidence level.

Maps of precipitation correlations with the CPC MJO index are very similar to the composite maps included in Figs. 3 and 5 and are not shown.

Relatively low associations between MJO indices and EA precipitation in April might be interpreted as a pause in MJO influence in the middle of the long rainy season. However, our analysis suggests that the apparent pause might simply be a result of the fact that the characteristics of the MJO vary from month to month and, as has been shown in previous studies, from decade to decade (Suhas and Goswami 2010). As such, no single, seasonally static MJO index system offers a perfect or complete representation of the MJO phenomenon. As shown in Fig. 2, there are large correlations between April EA precipitation and the MJO index at 10°W and 120°E at two-pentad lead time. These correlations are

conceptually consistent with the zero-lag correlations found in March and May, as the MJO propagates at speeds of $4\text{--}8\text{ m s}^{-1}$ when convectively coupled in the Indian Ocean, but at speeds as high as $30\text{--}35\text{ m s}^{-1}$ when uncoupled from convection in other regions (Zhang 2005). This means that a two-pentad-lead correlation with MJO activity at 10°W , for example, can be roughly indicative of zero-lag correlation with MJO activity in the vicinity of $70^{\circ}\text{--}80^{\circ}\text{E}$. Indeed, as described in section 3b, below, the mechanisms that underlie these two-pentad-lead correlations in April are quite similar to those associated with zero-lag correlations in March and May. This would seem to indicate that the lack of a strong zero-lag signal results from the fact that high and low precipitation anomalies in EA are associated with patterns of Indian Ocean convection that do not

FIG. 5. Composites of precipitation (mm day^{-1}) for 1998–2012, based on the MJO index at (a)–(c) 120°E and (d)–(f) 10°W , for (a),(d) October, (b),(e) November, and (c),(f) December. Color shading shows results significant at the 90% confidence level. Correlation maps at 90% significance level are nearly identical and are not shown.

align perfectly with CPC MJO indices at 70° and 80°E during this month. Indeed, zero-lag associations are significant for all intermediate variables linking MJO to EA precipitation (OLR patterns, vertical motion, and proposed dynamic and thermodynamic links described below) when analyses are performed for the longer 1979–2012 period or for 1979–97. This is true for both CPC and RMM MJO indices. This strongly suggests that there is a link between MJO activity and EA precipitation processes in April and that the lack of strong zero-lag association in the TRMM period of analysis results from some combination of nonstationarity in MJO behavior and in the difficulty of characterizing an evolving phenomenon like MJO with static index systems. This interpretation of statistical shifts in April is speculative and is a subject of our ongoing research. For consistency in presentation, we show zero-lag results for all months throughout this section. Lead time associations will be reintroduced in sections 3b and 3d.

OLR composites corresponding to the TRMM period (Fig. S14 in the supplementary material) are consistent with the precipitation analysis, showing strong associations between anomalies in deep convection in EA and convective activity in the Indian Ocean when MJO activity is centered at 80°E for all months that show significant associations between MJO at 80°E and precipitation. When the MJO index at 120°W is used, a pattern opposite

to the results with the MJO index at 80°E is observed. For April, on the other hand, results similar to the composites using MJO indices at 80°E (120°W) are seen employing two-pentad-lead MJO indices at 10°W (120°E). For OLR, however, we are able to perform pentad and daily analysis for a period that is not limited by the TRMM record. Figure 6 shows composites of OLR using concurrent 80°E and 120°W CPC MJO indices for the entire study period (1979–2012). These composites are consistent with the composites in 1998–2012 time period, with the exception that April composites for this period do show significant zero-lag OLR anomalies. This result is further evidence that April precipitation in the region is, in fact, affected by MJO. The lack of significant zero-lag association seen in TRMM composites reflects variability in the MJO activity, its influence on East Africa, and/or the ability of standard MJO indices to capture these connections. The possibility of variability in the strength of MJO associations in EA is discussed in section 3c. OLR composites generated using the RMM phase 3 and 6 indices give similar but weaker results for both 1979–2012 (Fig. S1 in the supplementary material) and 1998–2012 (not shown).

The consistency of TRMM and OLR results reflects strong correlation between precipitation and OLR variability in the region: at pentad scale, these correlations range from -0.81 to -0.89 in the months of both the

FIG. 6. Composites of OLR (W m^{-2}) based on MJO indices at (a)–(c), (g)–(i) 80°E and (d)–(f), (j)–(l) 120°W for 1979–2012, for (a), (d) March; (b), (e) April; (c), (f) May; (g), (j) October; (h), (k) November; and (i), (l) December. Shading shows results significant at the 90% confidence level.

long and short rainy seasons (using the averaging boxes described in Fig. 2), confirming that precipitation variability in the region is predominantly due to deep convection. Interestingly, OLR conditions indicative of deep convection in EA (i.e., strongly negative OLR anomaly) are in phase with convective activity in the MJO center of convection but are not contiguous with the core convection feature. In all cases, there is a gap or weakening in the OLR anomaly off the eastern coast of Africa that separates the EA convection anomaly from the MJO center of action. This suggests that the EA precipitation anomalies cannot simply be explained as a dynamical extension of MJO convection in the Indian Ocean. A mechanism of communication is required.

MJO-related anomalies are also evident in vertical motion fields over EA (Fig. 7). These anomalies represent an actual reversal of rising and descending motion in some cases and a weakening of prevailing motions in others. These anomalies were generated using NCEP-R1 vertical velocity fields, but the same patterns are evident when ERA-Interim vertical velocity data are used. During this period, when the convective envelope of the MJO is in the Indian Ocean (80°E), there is significantly enhanced upward vertical motion over EA in the months of March, May, November, and December, and corresponding wet anomalies. In contrast,

anomalous descent, which results in anomalous drying, is observed in the same months when the MJO migrates to the eastern Pacific Ocean (120°W). Vertical motion in April over EA does not show statistically significant association with MJO indices at zero lag in 1998–2012. However, in the 1979–2012 time period, consistent with the OLR anomalies, strong significant anomalies in vertical motion are observed in all months except in October. Similar patterns to the composites using the CPC MJO index at 80°E are found employing the RMM phase 3 index (Fig. S2 in the supplementary material). For all other months vertical motion anomalies in the 1998–2012 time period are consistent with the 1979–2012 results.

In October, OLR and vertical velocity anomalies associated with MJO activity at 80°E and 120°W are very weak, but there are strong OLR associations with MJO indices at 120°E and 10°W (Figs. 8a,b). This is consistent with the precipitation results (Figs. 5a,d): when MJO convective activity is centered at 120°E , positive OLR anomalies and negative rainfall anomalies are observed in the western Indian Ocean and coastal EA (Fig. 8a). When MJO convective activity moves to 10°W , negative OLR anomalies and wet anomalies are observed over the western Indian Ocean and the coastal EA region (Fig. 8b). These OLR anomalies are accompanied by

FIG. 7. As in Fig. 6, but for composites of vertical velocity (Pa s^{-1}).

anomalies in vertical motion: enhanced subsidence when the MJO is centered at 120°E and a tendency toward rising motion—which is sometimes actual rising motion and sometimes weakening of prevailing subsidence—when the MJO is centered at 10°W (Figs. 8c,d). MJO

activity at 120°E is also associated with warm SST and low SLP over the Maritime Continent and cool SST in the west equatorial Indian Ocean (Figs. 8e–g). The opposite holds when MJO convection moves to 10°W . The SST pattern in Figs. 8e and 8f is consistent with previous

FIG. 8. Composites of (a),(b) OLR (40 W m^{-2}), (c),(d) ω (0.05 Pa s^{-1}), (e),(f) SST (1.25°C), and (g),(h) SLP (0.67 hPa) and wind (m s^{-1}) in October based on MJO indices at 120°E for (a),(c),(e),(g) and 10°W for (b),(d),(f),(h). Shading shows results significant at the 90% confidence level. Black vectors indicate that wind anomalies are significantly different from zero at the 90% confidence level in at least one of the wind components (meridional or zonal). SST composites are for 1982–2012, while all the rest are for 1979–2012.

studies finding that positive (negative) SST anomalies tend to lead the maximum (minimum) in MJO convection by about two pentads (e.g., [Lavender and Matthews 2009](#)). As indicated in those previous studies, the two-pentad-lead SST anomalies associated with MJO convection result from the impact that the MJO convective center has on the ocean surface radiation and energy balances. In this sense the SST pattern observed here is an MJO response, but it is a direct response to the propagation of the MJO primary center of convection that does not depend on communication between the Indian Ocean MJO center of convection and East Africa.

The results shown in all panels in [Fig. 8](#) are consistent with anomalies in the zonal Walker circulation in the Indian Ocean. In October, when MJO activity is greatest at 120°E, there is a strengthened Walker cell across the Indian Ocean that enforces subsidence over the western Indian Ocean and coastal EA. When the MJO is centered at 10°W, there is a reversed or weakened Walker cell that leads to actual/anomalous upward motion in EA and downward motion in the Maritime Continent. This association is not observed in the other months of the short rains. We note that the MJO influence on the Walker cell bears some resemblance to known Indian Ocean dipole index (IOD) associations with the Walker cell and EA precipitation ([Saji et al. 1999](#)). The MJO association, however, occurs on shorter time scales than IOD variability and has an SST pattern ([Figs. 8e,f](#)) that is spatially more extensive than the classic IOD signal ([Saji et al. 1999](#)).

The strength of OLR, vertical motion, and large-scale atmospheric circulation anomalies associated with MJO-mediated precipitation variability in coastal EA in October contrasts with the findings of [PC06a](#), who found only weak OLR associations with MJO-mediated precipitation variability in the coastal region. The difference could be due to differences in dataset or period of analysis. In either event, our result suggests that MJO influence on coastal EA precipitation is due to deep convective processes influenced by variability in a Walker circulation over the Indian Ocean, and cannot be attributed solely to shallow convection or stratiform precipitation, which were the primary mechanisms identified by [PC06a](#). The Walker circulation association is also observed when composites are averaged over the entire short rainy season, but this seasonal result derives from the strong association in October.

Intraseasonal precipitation variability during the EA long rains has also been linked to variability in the strength of near-surface and midlevel westerly wind anomalies ([Camberlin and Wairoto 1997](#); [Okoola 1999](#); [Camberlin and Okoola 2003](#), among others). The most frequently cited mechanistic link between westerlies and

EA precipitation is that these winds have the potential to transport moisture from the Atlantic Ocean and Congo basin into the region ([Nicholson 1996](#)). [PC06a](#) found that anomalies in low-level zonal winds are associated with the MJO, and our analysis confirms this finding. During the long rains, strong low to midlevel westerly anomalies are associated with MJO convection at 80°E and the opposite is seen for MJO convection at 120°W ([Fig. 9](#); compare to precipitation and OLR in [Figs. 3 and 5](#)). This pattern is clear for both the CPC and RMM index ([Fig. S3](#) in the supplementary material) composites, and it is similar for both the 1979–2012 and 1998–2012 time periods (with the exception of the previously noted lack of significance in April for the 1998–2012 period). We observe a similar pattern in the short rainy season. The anomalies do not, however, necessarily represent actual wind reversals. Analyses of daily low-level winds indicate that even when MJO convection is strong at 80°E very few incursions of humid and unstable westerlies from the Congo basin occur during the long rains; the anomaly is simply a weakening of prevailing easterly winds. This suggests that near-surface westerly wind anomalies influence long rainy season EA precipitation primarily because they represent a weakening of easterly winds (mechanisms discussed below), and not because of actual incursion of humid air from the Congo basin. In November and even more frequently in December westerly wind anomalies are, in fact, associated with wind reversals and with the incursion of westerly winds into EA.

During both the long and short rains, these MJO-associated westerly wind anomalies are a product of SLP variations in the Indian Ocean. When MJO convection is centered on the Indian Ocean (80°E), a negative SLP anomaly produces a pressure gradient that weakens the easterlies and enhances westerly flow over the African continent, and the opposite holds when convection is suppressed in the Indian Ocean ([Fig. 9](#)). Similar to the SLP composites using the 80°E CPC index is found using RMM phase 3 ([Fig. S3](#)). This link between Indian Ocean SLP and both winds and precipitation over EA is consistent with previous studies ([PC06a](#); [PC06b](#); [Black et al. 2003](#); [Goddard and Graham 1999](#)).

b. Mechanisms of precipitation variability

The strong precipitation, OLR, and low-level wind anomalies described above are consistent with multiple potential mechanisms of MJO influence. Enhanced westerlies, for example, could enhance precipitation in EA through increased moisture flux to the region, stronger low-level convergence, and/or reduced stability in the lower troposphere. MJO-induced differences in

FIG. 9. Composites of SLP (hPa) and wind at 850 hPa (m s^{-1}) based on MJO indices at (a)–(c), (g)–(i) 80°E and (d)–(f), (j)–(l) 120°W for 1979–2012, for (a),(d) March; (b),(e) April; (c),(f) May; (g),(j) October; (h),(k) November; and (i),(l) December. Shading shows results significant at the 90% confidence level. Black vectors indicate that wind anomalies are significantly different from zero at the 90% confidence level in at least one of the wind components (meridional or zonal).

Indian Ocean convection, meanwhile, impact lower-tropospheric air temperature over the Indian Ocean and the potential for both moisture advection and energy advection into EA from the east. Our analysis suggests that multiple mechanisms are active, and that the relative importance of these mechanisms differs between the long and short rains and between the interior highlands and the coast.

Focusing on the mechanisms that are most evident in MJO-based composites, we observe that low-level moisture flux divergence is strongly influenced by MJO activity during the short rains (Fig. 10). The sign of the divergence composites is consistent with precipitation anomalies—negative divergence (i.e., convergence) is observed when MJO activity is centered in the Indian Ocean, westerly wind anomalies are enhanced (Fig. 9), and the EA precipitation anomaly is positive. This indicates that a mechanism of moisture convergence leading to enhanced convection is potentially quite important in the short rains. The divergence composites

are present in the long rains as well, but are somewhat less pronounced.

For Figs. 10 and 11, we show only composites for 80°E ; composites for MJO centered at 120°W are approximately equal and opposite to the 80°E composites in all cases. Additionally, for April we have replaced the 80°E zero-lag composites with two-pentad lead time composites on the MJO index at 10°W . As explained previously, the two-pentad-lead MJO activity at 10°W shows stronger correlation with EA precipitation in April than any zero-lag MJO index (Fig. 2) in the 1998–2012 time period.

Lower atmosphere moist static instability anomalies associated with MJO activity are pronounced in EA in the long rains and middle and late short rains (November and December) ($H_{1000} - H_{5700}$; Figs. 10 and 11). This is consistent with a mechanism in which anomalously warm, wet winds enter the region from the Indian Ocean—MJO convection in the Indian Ocean warming the lower troposphere (not shown)—and the slowing of

FIG. 10. Composites of (a)–(c) MFD ($\frac{1}{3} \times 10^{-7} \text{ s}^{-1}$), (d)–(f) buoyancy resulting from the MSE profile ($H_{1000} - H_{5700}$; $8 \times 10^3 \text{ J kg}^{-1}$), and (g)–(i) temperature advection (T_{adv} ; $2 \times 10^{-5} \text{ C}^\circ \text{ s}^{-1}$) and wind at 850 hPa (m s^{-1}) for the short rains for 1979–2012, for (a), (d), (g) October; (b), (e), (h) November; and (c), (f), (i) December. The composites are based on MJO at 80°E . Shading shows results significant at the 90% confidence level. Black vectors indicate that wind anomalies are significantly different from zero at the 90% confidence level in at least one of the wind components (meridional or zonal).

winds allows for further local heating of the lower atmosphere. The resulting warm, moist air mass is primed for convective precipitation events.

Indeed, analysis of the thermodynamic balance confirms this link. Figures 10g–i and 11g–i show the advective component of the lower atmosphere thermodynamic balance. In all months of both the short and long rainy seasons there is a significant association between MJO activity in the Indian Ocean and temperature advection into EA: enhanced MJO activity at 80°E (or, for April, 10°W) is associated with higher near-surface air temperature over the Indian Ocean, which is carried into EA on low-level easterly winds, and the opposite occurs when MJO activity is centered at 120°W . Even though low-level winds entering the region are anomalously slow when MJO convection is in the Indian Ocean (see the wind anomaly vectors in Fig. 9), the warmer near-surface air temperatures over the Indian Ocean result in more energy advection into the region, contributing to instability and to moisture convergence. Analysis using ERA-Interim data shows similar patterns of temperature advection. Analysis of temperature advection, moisture flux divergence, and buoyancy using the RMM phase 3 index gives similar, but weaker, results (Figs. S4–S6 in the supplementary material).

The MJO influence on coastal EA precipitation in October is distinct from the mechanisms that link MJO to the highlands. In this month, temperature and vertical

motion anomalies in the Indian Ocean that occur when MJO convection is centered on 120°E are associated with strengthening of the zonal Walker circulation while MJO convection at 10°W is associated with weakening and even reversal of that circulation (Fig. 8). The impact of this large-scale strengthening and weakening/reversal of the Walker circulation extends just to the edge of EA, such that the precipitation effects are felt on the coast but not in the interior highlands. The strengthening (weakening or reversal) of the Walker circulation when the MJO is centered at 120°E (10°W) is also associated with stronger (weaker) near-surface winds along the coast of EA sometimes referred to as the Somali low-level jet (SLLJ). A strengthening of the SLLJ has been associated with enhanced frictionally induced subsidence on the EA coast (Nicholson 1996). However, the depth and extent of the subsidence feature suggest that frictionally induced subsidence is not the sole explanation for MJO-associated precipitation effects in these phases. Similar results are found when ERA-Interim data are used.

The mechanisms of MJO influence on the long rains, short rains, and coastal rains in October are summarized in Fig. 12. The thermodynamic mechanism of influence appears to operate throughout the long rains and short rains. The dynamic, convergence-mediated mechanism is strong and widespread throughout the short rains. It is also strong in the long rains, especially in March and April. The October coastal

FIG. 11. Composites of (a)–(c) MFD ($1/3 \times 10^{-7} \text{ s}^{-1}$), (d)–(f) buoyancy resulting from the MSE profile ($H_{1000} - H_{5700}$; $8 \times 10^3 \text{ J kg}^{-1}$), (g)–(i) temperature advection (T_{adv} ; $2 \times 10^{-5} \text{ C}^\circ \text{ s}^{-1}$) and wind at 850 hPa (m s^{-1}) for the long rains for 1979–2012, for (a), (d), (g) March; (b), (e), (h) April; and (c), (f), (i) May. The composites are based on MJO at 80°E for March and May and MJO at 10°W at two-pentads lead for April. Shading shows results significant at the 90% confidence level. Black vectors indicate that wind anomalies are significantly different from zero at the 90% confidence level in at least one of the wind components (meridional or zonal).

mechanism of influence is distinct and is associated with the strengthening, reversal, or weakening of the Walker cell in the Indian Ocean.

c. Stability of associations

PC06b found that the strength of association between MJO and their study region in EA varied considerably on an interannual basis: the common variance between MJO and their smoothed rainfall time series varied from 5% to 53% on an annual basis for their period of analysis. Similarly, the influence of MJO on EA might change over longer time periods because of either evolving characteristics of MJO itself or changes in the strength of its connection to EA. To evaluate the stationarity of MJO influence on EA we compare results for the periods 1979–97, 1998–2012, and 1979–2012. Overall, we find a consistent pattern for each of these time periods, but there are some differences in monthly scale associations. As noted previously, April stands out as a month in which concurrent associations between MJO indices at 80°E and 120°W and mechanisms influencing EA precipitation are difficult to detect in the 1998–2012 period. These associations are much clearer for 1979–2012 (Figs. 6, 7, and 9–11), and they are also clear for 1979–97 (Figs. S12–S13 in the supplementary material).

In addition to April, minor shifts in the significance of MJO associations are found in November—when statistical associations are weaker for 1979–97 than they are for

either 1979–2012 or 1998–2012—and May, when OLR results are weaker for 1979–97 than for the other time periods, even though dynamic and thermodynamic composites are consistent with a significant MJO influence in all three periods. For coastal East Africa, the associations between precipitation and MJO indices at 120°E and 10°W are consistent for 1979–97, 1998–2012, and 1979–2012.

Overall, these analyses of different time periods suggest that the influence of the MJO on EA has shifted somewhat between the 1979–97 and 1998–2012 periods, and that this shift has altered the statistical association between standard MJO indices and EA precipitation variability. Nevertheless, the general patterns and mechanisms of association are consistent across time periods. This indicates that the MJO has been an important influence on EA precipitation since at least 1979.

d. Potential for prediction

As shown in Figs. 2 and 13, the CPC MJO indices are significantly correlated with EA precipitation at various lead times. For April, for example, at a lead of two pentads, we see correlations of 0.58 with the MJO index at 10°W and -0.59 with the MJO index at 120°E (Figs. 2 and 13). These correlations are both strong and widespread across the EA highlands. Significant correlations (0.41 and -0.40 with the MJO indices at 160° and 20°E , respectively) are also found in April at a lead of four pentads.

FIG. 12. Mechanisms of MJO influence on East African precipitation.

Various studies have developed statistical and/or dynamical prediction models of the MJO with skills extending to about 25 days and beyond (e.g., Waliser et al. 2003; Maharaj and Wheeler 2005; Vitart et al. 2007; Jiang et al. 2008; Lin et al. 2008; Love et al. 2008; Love and Matthews 2009; Seo 2009; Seo et al. 2009; Kang and Kim 2010; Rashid et al. 2011). Waliser et al. (2003), for example, find that in the Eastern Hemisphere useful predictability of the MJO extends out to about 25–30 days for 200-hPa velocity potential. Employing a dynamical prediction system, Rashid et al. (2011) document that the MJO can be predicted to about 21 days.

Insomuch as MJO variability can be predicted with reasonable skill on a time horizon of up to one month, the lead-time correlations presented in Fig. 13 suggest that predictions of the variability and volume of rains in April, which is the main rainy month in the study region, could be usefully attempted up to two months in advance. In other months, correlations at somewhat shorter lead times could be used to predict intraseasonal variability on the order of one month to six weeks in advance. In April and May the two-pentad-lead correlations are promising, while in November, December, and March correlations are stronger at zero to one-pentad lead. Employing the MJO indices in forecast models can help to minimize difficulties in the predictions of the long rains characteristics (Semazzi et al. 1996; Okoola 1998; Indeje et al. 2000; Camberlin et al. 2009) and to mitigate impacts of weather extremes by mobilizing resources in appropriate time.

4. Conclusions

The livelihood of millions of people in EA is critically affected by frequent droughts and floods that retard progress toward the United Nations Millennium Development Goals (<http://www.un.org/millenniumgoals/>). From the perspective of floods, flash droughts, and crop production, understanding variability on intraseasonal time scales is vital to develop weather and short-term climate forecasting tools to mitigate impacts of departures from normal weather in the region.

FIG. 13. Correlations of area average precipitation over EA, for 1998–2012, with the various MJO indices at various lead times that are significant at the 90% confidence level using the reduced degrees of freedom resulting from autocorrelation. At a given pentad lead, the highest negative and positive correlations are provided. Regions used for each month are 12°–2°S, 34°–38°E for March; 6°S–5°N, 30°–38°E for April and May; 2°S–8°N, 30°–38°E for October; 4°S–4°N, 32°–38°E for November; and 10°S–2°N, 32°–38°E for December.

Numerous studies have noted the low temporal coherence of precipitation in EA and have urged for further study of precipitation variability at subseasonal time scales (e.g., Camberlin and Philippon 2002; Zorita and Tilya 2002; Camberlin and Okoola 2003). This study examined the associations of precipitation and MJO, using the CPC and RMM MJO indices, in EA for each calendar month separately employing pentad and daily data. The region of MJO influence in EA varies within each rainy season, as the ITCZ moves through the region. The magnitude of MJO influence also varies from month to month. During the long rains, the influence is somewhat greater in the late season (May) than the early season (March), while the middle of the season (April) shows no statistically significant association with synchronous CPC MJO indices. This apparent “pause” in MJO influence in April, however, seems to be a product of modest nonstationarities in MJO that make it difficult to characterize associations across all months and time periods using standard static MJO index systems: we do find lead-time associations between April precipitation and MJO indices that are consistent with proposed MJO mechanisms of influence, and statistical relationships at zero lag are significant in April when we consider other time periods. During the short rains, the MJO modulation of EA highland precipitation is significant in November and December. In October the MJO influence on highland EA is modest, but strong associations between MJO and precipitation in the coastal regions of EA are observed.

The results show that anomalous wetness is experienced when the MJO convective center is between 20° and 120°E and dry anomalies prevail when the MJO is located in the region from 160°E and 10°W. Maximum correlations between the MJO strength and precipitation over EA are observed using 70°–80°E and 120°W MJO indices. When the MJO convective envelope is located at 70°–80°E (120°W), negative (positive) SLP anomalies over the Indian Ocean are observed that result in westerly (easterly) anomalies in low and mid-troposphere winds. In the short rains, when the MJO is centered in the western Indian Ocean, the Congo air boundary (CAB) moves east and brings moist, unstable air to EA, resulting in increased convective activity. In contrast, when the MJO moves to the eastern Pacific, the CAB moves far west and cold and dry easterlies prevail over EA.

Low-level easterly winds in the western Indian Ocean are weaker when the MJO is centered in the western Indian Ocean, but they are anomalously warm and advect significant moisture and temperature to the EA. However, as the MJO moves to the eastern Pacific Ocean, the easterlies become stronger, dry, cold, and

divergent. In November and December there is strong evidence for both a dynamic precipitation influence due to lower atmosphere convergence and a thermodynamic mechanism due to temperature advection and lower atmosphere instability. Both mechanisms are also potentially active in the long rainy season. In October a completely distinct mechanism of influence is evident in the coastal region, as MJO-mediated precipitation variability is associated with the strengthening and reversal or weakening of the Walker cell in the Indian Ocean. The fact that the MJO influence on EA depends on a diverse set of large-scale and synoptic-scale mechanisms is relevant to studies of subseasonal precipitation variability in a changing climate. Changes in either the MJO or in background humidity and winds affecting EA have the potential to modify each of these mechanisms in different ways. A robust understanding of climate change impacts on the region, then, must include an appreciation for how various mechanisms of MJO influence might evolve in a nonstationary climate.

Another potentially useful result of this analysis is the finding that precipitation over EA exhibits significant correlations with the MJO indices at reasonable lead times. Therefore, since the MJO position and strength can be predicted with skill up to a month in advance, a combination of MJO prediction and the statistical associations explored in this paper could be applied to predict rainfall anomalies in EA at time scales between long-range weather forecasts and short-term seasonal predictions. Demonstration of this predictive potential is the subject of ongoing research.

Acknowledgments. The authors thank Dr. Anand Gnanadesikan for useful discussion. This study was supported in part by NASA Applied Sciences Grant NNX09AT61G.

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