Three-Dimensional Structure and Evolution of the MJO and Its Relation to the Mean Flow

ÁNGEL F. ADAMES AND JOHN M. WALLACE
Department of Atmospheric Sciences, University of Washington, Seattle, Washington

(Manuscript received 10 August 2013, in final form 27 January 2014)

ABSTRACT

The two leading principal components of the daily 850- minus 150-hPa global velocity potential in the Interim European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-Interim) (1979–2011) data are used as time-varying Madden–Julian oscillation (MJO) indices. Regression maps and meridional cross sections based on these indices are used to document the structure and evolution of the zonal wind (u) and geopotential height (Z) anomalies in the MJO cycle. The data are daily, and they are not separated by season. At upper-tropospheric levels the MJO signature is dominated by eastward-propagating planetary wave packets consisting of equatorial Kelvin waves flanked by Rossby waves centered along 28°N/S, for which the westerly jet streams serve as waveguides. At lower-tropospheric levels the pattern more closely resembles the response to a pulsating heat source over the Maritime Continent, where the Andes block the eastward-propagating Kelvin wave pulse. The contrasting upper- and lower-tropospheric patterns are made up of the same building blocks: a deep, baroclinic modal structure with a node at the 400-hPa level, which dominates the tropical signature, and a barotropic residual field consisting mainly of extratropical wave trains oriented along great circles. The extratropical wave trains emanate from the flanking Rossby waves in the baroclinic modal structure. The strongest of them, which resembles the Pacific–North America (PNA) pattern, extracts kinetic energy from the climatological-mean flow in the jet exit region. At other longitudes the jet stream seems to act as a barrier to the poleward propagation of MJO-related wave activity.

1. Introduction

The Madden–Julian oscillation (MJO, after Madden and Julian 1971, 1972) involves the coupling between planetary-scale waves and deep cumulus convection [as summarized in reviews by Zhang (2005) and Lau and Waliser (2012)]. In composites constructed by averaging many individual MJO events, a planetary-scale envelope of enhanced convective activity propagates eastward across the Indo-Pacific warm pool sector from ~60°E to the date line over a period of ~20 days—the time required for the corresponding equatorial planetary wave signature to propagate halfway around the entire tropical belt (Hendon and Salby 1994; Matthews 2000; Sperber 2003; Kiladis et al. 2005). The eastward-propagating signal in tropical convection is clearly evident in time–longitude sections of outgoing longwave radiation (OLR) even without recourse to compositing (see, e.g., Matthews 2008 and Straub 2013) and the remote signature in the planetary-scale wind field is evident in the global field of upper-tropospheric velocity potential (Knutson and Weickmann 1987; Geisler and Pitcher 1988; Slingo et al. 1996; Chen and Del Genio 2009) and in the analyses of zonal wind (Hendon and Salby 1994; Salby and Hendon 1994; Bantzer and Wallace 1996; Lin et al. 2005; Kiladis et al. 2005). The vertical structure of the planetary-scale wind field in the MJO is dominated by a deep baroclinic mode, which is characterized by a phase reversal between upper and lower troposphere (Madden and Julian 1972; Webster 1972; Gill 1980). Off-equatorial cyclonic and anticyclonic gyres are an integral part of the MJO signature over the convectively active warm pool region, as documented in Knutson and Weickmann (1987), Murakami (1988), Rui and Wang (1990), and many subsequent studies.

Individual envelopes of enhanced tropical convective activity rarely exhibit the canonical MJO structure. Often it is not obvious what triggers the convection over...
the Arabian Sea at the western end of the warm pool (Hendon and Salby 1994; Matthews 2000; Sperber 2003; Matthews 2008; Zhao et al. 2013; Straub 2013). Even when the data are averaged over many events to form composites, the time sequences of constant pressure level charts and equatorial (vertical) cross sections are quite complicated and difficult to interpret. Individual studies differ with respect to the positioning of the off-equatorial cyclonic and anticyclonic gyres and with respect to the relative strength of the gyres to the east and west of the MJO-related enhanced equatorial convection. Some previous studies have labeled these cyclones and anticyclones as Rossby waves while others have avoided using that term as a label for eastward-propagating features. Some studies have portrayed them as existing in the upper and lower troposphere with reversed phase, while others have portrayed them as upper-tropospheric features only. The situation is further complicated by the fact that the structure of the MJO is seasonally and longitudinally dependent and has been depicted in some studies as evolving through the life cycle keyed to flareups in the equatorial convection.

During the boreal winter the tropical MJO perturbations force an extratropical wave train originating in the North Pacific sector that resembles the Pacific–North America (PNA) pattern (Weickmann 1983; Lau and Phillips 1986). The wave train appears at the phase of the MJO cycle when the enhanced equatorial convection is propagating eastward across the Maritime Continent. The excitation of the PNA pattern has been widely attributed to tropical forcing: specifically the advection of planetary vorticity by the divergent, upper-tropospheric outflow from the enhanced convection as proposed by Sardeshmukh and Hoskins (1988); for example, see Matthews et al. (2004), Mori and Watanabe (2008), and Frederiksen and Lin (2013).

In this paper, we examine Interim European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-Interim) and OLR data with hopes of gaining a more complete understanding of the MJO life cycle. The data are not separated by season. We make extensive use of the leading empirical orthogonal function (EOF) of the global velocity potential (\( \chi \)) field. Analyzing zonal wind (\( u \)) in combination with geopotential height (\( Z \)) allows for a more definitive identification of the modal structures that appear in various phases of the MJO life cycle than in studies based on streamfunction. In contrast to previous analyses of the MJO, rather than choosing a single level or set of levels to represent the fields, we consider data for all levels from the surface up to 50 hPa.

The paper is structured as follows. The next section describes the datasets and methods of analysis used in the paper. Section 3 describes and justifies our use of the leading principal components of the global, daily 850- minus 150-hPa global velocity potential field as time-varying indices to represent the phase of the MJO. It also describes the evolving patterns of upper-tropospheric zonal wind and geopotential height anomalies observed in association with the MJO cycle, including the flanking Rossby waves centered just equatorward of the westerly jet streams. Section 4 contrasts the structures of the MJO-related \( u \) and \( Z \) fields in the upper and lower troposphere. We describe a method of analyzing MJO regression fields on an array of pressure levels to obtain a modal structure that appears with reversed polarity in upper-tropospheric zonal wind and geopotential height fields, together with the corresponding residual patterns that account for the remaining structure in the regression patterns in each layer. The analysis in this section offers insight into the marked differences between upper- and lower-tropospheric MJO signatures. Section 5 relates the tropical and extratropical components of the upper-tropospheric MJO signature. Section 6 offers a global perspective on the MJO based on variance maps of zonal wind and geopotential height. The final summary and discussion section highlights the most important new results in this study. The appendix demonstrates that the modal structure and regression patterns derived from our MJO index based on velocity potential are quite similar to those based on other MJO indices.

2. Data and methods

Three datasets are extensively used in this study. The first is the \( 1.5^\circ \) longitude \( \times 1.5^\circ \) latitude horizontal resolution, four times daily, ERA-Interim fields (Dee et al. 2011) for the 33-yr time period 1979 through 2011. The horizontal wind components are used in this study as a field and to calculate the velocity potential (\( \chi \)) field, while the zonal wind component \( u \) is also analyzed as a scalar field. Geopotential height (\( Z \)) is also used in this study, in particular to document the Kelvin and Rossby-waves’ components of the MJO signature and to define the MJO’s impact on the extratropical circulation. The convective envelope associated with the MJO is represented by OLR from National Oceanic and Atmospheric Administration (NOAA)’s polar-orbiting satellites (Liebmann and Smith 1996).

The seasonal cycle is removed from the dataset using a method extensively used in previous studies dating back to Wheeler and Kiladis (1999). We subtract the mean and first three harmonics of the annual cycle, based on the 1979–2011 period, at every grid point. We remove the ENSO signal by regressing out the first two EOF modes of equatorial Pacific sea surface temperature. We also subtract the running mean of the last 120
days from the time series at each data point as in Lo and Hendon (2000). Unless specifically indicated, all variables used in this study have their seasonal cycle and other low frequency components removed in this manner. While not all these steps in the preprocessing of the data are necessary to isolate the MJO signal, we perform them in order to make our results easily comparable with those of previous studies. This procedure also allows our MJO index to be calculated in real time.

The results of this study are based on regression analysis. For a two-dimensional matrix $\mathbf{S}$ that represents a variable field $\mathbf{S}$, the equation for linear regression takes the form

$$
\mathbf{D} = \mathbf{S}\mathbf{P}^T/N,
$$

where $\mathbf{D}$ is a regression pattern with dimensional units, $\mathbf{P}$ a standardized MJO index, and $N$ is the sample size in days. All the results presented in this paper are based on data for all 12 calendar months.

The statistical significance of these correlations was estimated by applying a two-tailed test at the 95% level under the null hypothesis that the time series are uncorrelated. A critical value of the Fisher transformed correlation coefficient is estimated using the relation

$$
B = 1.96\sqrt{N^*-3},
$$

where $N^*$ is the effective degrees of freedom of the time series. The $N^*$ value is calculated using the method described by Chen (1982), in which the decorrelation time scale is defined by the product of the autocorrelation coefficients of the two fields being compared, summed over all time lags. The decorrelation time scale and $N^*$ are calculated at each individual grid point. Through the use of this method, a critical correlation coefficient is calculated for every spatial point, and correlations above this value are treated as distinct from zero. The decorrelation time varies with latitude, ranging from 8–10 days in the tropics to $\sim3$ days in the midlatitudes. The longest decorrelation time scale is 10 days, so the smallest effective number of degrees of freedom is $36\text{ yr}^{-1}$. The critical correlation coefficient obtained with this value is $r_c = 0.08$. The contour intervals for the regression coefficients in the figures in this paper correspond to correlation coefficients of 0.08 or larger. Hence, the patterns shown in these figures all qualify as statistically significant. The robust year-round structure and dynamical consistency of these patterns also attests to their statistical significance.

3. Velocity potential representation of the MJO

Velocity potential ($\chi$) is the inverse Laplacian of the divergence field. Taking the inverse Laplacian acts as a spatial smoothing operator, so the features in $\chi$ correspond to the largest-scale structures of the divergence field (Hendon 1986; Geisler and Pitcher 1988). The use of velocity potential as an MJO index dates back to Lorenc (1984), who analyzed daily values of the 200-hPa $\chi$ field during the Global Atmospheric Research Program (GARP) Global Weather Experiment (1979) and found that the two leading EOFs assumed the form of an eastward-propagating zonal wavenumber-1 pattern that circles the equator in a period of 30–50 days. Following that study, EOFs of $\chi$ were used as a diagnostic for estimating the zonal propagation of the MJO by Knutson and Weickmann (1987), who found that 250-hPa $\chi$ anomalies coherently propagate eastward together with anomalies in OLR and 850- and 250-hPa zonal wind. An alternative MJO index was developed by von Storch and Xu (1990) based on principal oscillation pattern analysis of 200-hPa $\chi$. Hsu et al. (1990) and Weickmann and Khalsa (1990) questioned the use of $\chi$ on the grounds that it does not accurately represent the centers of strongest divergence and that the $\chi$ anomalies propagate smoothly around the globe while the OLR anomalies are largely restricted to the Indo-Pacific warm pool region. Thus, OLR, which was first used to describe the MJO in the studies of Weickmann (1983), Lau and Chan (1985), and Weickmann et al. (1985), began to be favored as the indicator of the status of the MJO, and it has been widely used in the development of indices for use in the analysis and prediction of the MJO (Walisser et al. 1999; Lo and Hendon 2000; Wheeler and Hendon 2004; Kiladis et al. 2014). The all-season real-time multivariate MJO index (RMM) proposed by Wheeler and Hendon (2004), based on the EOFs of 850- and 200-hPa zonal winds and OLR, has been widely used in observational studies.

Recently, the use of $\chi$ as a diagnostic tool for studying the MJO has received renewed attention. Chen and Del Genio (2009) developed an MJO index based on extended EOF analysis applied to pentad-mean 200-hPa $\chi$ anomalies, and Ventrice et al. (2013) proposed a variant of the widely used multivariate (zonal wind and OLR) index of Wheeler and Hendon (2004) that uses velocity potential in place of OLR. Results based on their velocity potential multivariate (VPM) index show a larger amplitude of MJO-associated anomalies over the western hemisphere as well as a more robust correlation with tropical cyclone activity. As a complement to the VPM, Kiladis et al. (2014) have proposed an index based on OLR alone. In this paper we further document the structure of the $\chi$ field and use it as the reference variable in describing the planetary-scale MJO structure.

Consistent with results of previous studies (Lorenc 1984; von Storch et al. 1988; Knutson and Weickmann 1987), we have confirmed that EOF analysis performed on the horizontal distribution of $\chi$ based on global data...
Table 1. Statistics for various MJO indices derived from velocity potential. The RMM (Wheeler and Hendon 2004), VPM (Ventrice et al. 2013), and OMI (Kiladis et al. 2014) are also included for comparison. The second and third columns show the explained variance of the PCs and the fourth column shows the mean coherence (coh) squared between the two leading PCs in the 30–80-day frequency band as in Ventrice et al. (2013). The fifth column shows the maximum cross correlation between the two leading PCs and the last column shows the time lag at which the maximum cross correlation occurs. For all indices shown, the period of record 1979–2011 is used except for the VPM for which the period 1989–2011 is used. Because the OMI is derived using a 121-day sliding window centered on each day of the calendar year, the range of variance explained by its PCs reflect seasonal cycle variations (see Kiladis et al. 2014, their Fig. 1).

<table>
<thead>
<tr>
<th></th>
<th>%Var</th>
<th>%Var</th>
<th>(Mean coh)^2</th>
<th>Max cross correlation</th>
<th>Lag (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PC(χ_{850})</td>
<td>21</td>
<td>18</td>
<td>0.73</td>
<td>0.51</td>
<td>6.75</td>
</tr>
<tr>
<td>PC(χ_{200})</td>
<td>24</td>
<td>19</td>
<td>0.81</td>
<td>0.56</td>
<td>6.75</td>
</tr>
<tr>
<td>PC(χ_{150})</td>
<td>28</td>
<td>27</td>
<td>0.82</td>
<td>0.56</td>
<td>6.75</td>
</tr>
<tr>
<td>PC(Δχ)</td>
<td>29</td>
<td>26</td>
<td>0.83</td>
<td>0.57</td>
<td>6.75</td>
</tr>
<tr>
<td>VPM</td>
<td>22</td>
<td>20</td>
<td>0.84</td>
<td>0.59</td>
<td>8</td>
</tr>
<tr>
<td>RMM</td>
<td>13</td>
<td>12</td>
<td>0.76</td>
<td>0.56</td>
<td>9</td>
</tr>
<tr>
<td>OMI</td>
<td>27–33</td>
<td>26–32</td>
<td>0.69</td>
<td>0.65</td>
<td>10</td>
</tr>
</tbody>
</table>

(90°S–90°N), defined on a number of different pressure levels yields two leading modes whose standardized principal components are suitable as MJO indices. For the purposes of this study we use the difference between the χ fields on the 850- and 150-hPa levels, denoted as Δχ. The 850-hPa level is chosen to represent the lower troposphere by virtue of its widespread use in previous studies. The 150-hPa level is chosen because it has the strongest upper-level signature. The index tends to be dominated by χ_{150}, which is usually much larger in absolute value than χ_{850} and is positive in regions of upper-level divergence and low-level convergence. Statistics for different combinations of χ are summarized in Table 1 and compared with those for RMM, the VPM, and the OLR-based MJO index (OMI; Kiladis et al. 2014).

The two leading EOFs of the daily Δχ field shown in Fig. 1 explain 29% and 26% of the total globally integrated variance, respectively, for a combined total of 55%. As in other EOF-based analyses of the MJO that include the upper-tropospheric χ field, EOF1 exhibits a zonal wavenumber-1 (k = 1) structure in the equatorial belt, with a strong positive center over the Maritime Continent and a more diffuse negative center of action over South America, while EOF2 exhibits a prominent dipole centered over the Maritime Continent and also resembles a wavenumber-1-like pattern in the equatorial belt, in quadrature with the wave in EOF1. When the positive polarities of EOF1 and EOF2 are defined as in Fig. 1 the corresponding expansion coefficient time series, defined as principal components (PCs) 1 and 2, vary in quadrature with one another, with PC2 lagging PC1, indicative of eastward propagation of the k = 1 pattern. That the EOF patterns exhibit tighter Δχ gradients over the warm pool than over the remainder of the equatorial belt indicates that the amplitude of the Δχ perturbations is larger and the rate of eastward propagation is slower in that sector. The coherence between PCs 1 and 2 of the 150-hPa Δχ averaged over the 30–80-day period range field is slightly stronger than that in the more widely used 200-hPa Δχ field. For reasons that are not entirely clear, the cross correlations between PC1 and PC2 of Δχ peak at 6.75 days, compared to 9 days for the RMM indices. A more detailed comparison of MJO structure and evolution, as revealed by various indices, is presented in the appendix.

The OLR pattern observed in association with PC1 (Fig. 1a) is characterized by negative values, indicative of active convection centered over the Maritime Continent and positive values, indicative of suppressed convection, centered over South America. The 150-hPa wind anomalies are mainly zonal. The pattern is suggestive of an equatorial Kelvin wave signature that extends from the western Pacific to South America, but rather than being equatorially trapped, the band of westerlies extends out to well beyond 20°N/S. At this time in its cycle, the MJO exhibits a remarkable amount of symmetry with respect to the Maritime Continent (~130°E), with the patterns west of the Δχ maximum of nearly equal amplitude and opposing polarity to the patterns east of it. These results are qualitatively consistent with those obtained in previous studies of Knutson and Weickmann (1987), Hendon and Salby (1994), Matthews (2000), Sperber (2003), and Kiladis et al. (2005). In the patterns based on PC2 of Δχ, which reaches a maximum 6.75 days later (Fig. 1b), the enhanced convection has propagated eastward into the western and central Pacific. The band of westerlies that was over the Pacific at the time of maximum PC1 become elongated and its front edge has nearly completed its circuit around the equatorial band and is seen returning to the Indian Ocean sector. At this time in the MJO cycle, westerlies span ~3/4 of the tropical belt, and atmospheric angular momentum is at its peak (Weickmann and Sardeshmukh 1994; Weickmann et al. 1997, 2000).

A number of different MJO-related phase speeds can be inferred from the patterns in Fig. 1 and the lag correlations between the associated PC time series:

- The zonally averaged phase speed for the Δχ perturbations is 360° of longitude per 27 days, which is equivalent to ~16 m s⁻¹.
- The Δχ extrema in the dipole pattern in EOF2 are separated by about 120° of longitude. Features within
the interior of the dipole propagate through that distance over a time interval of $1/2$ MJO cycle while features propagating eastward from the eastern pole, across South America and Africa to the western pole, cover twice that distance in the same time. It follows that the eastward phase speed of features in the $\chi$ pattern averages $\sim 12 \, \text{m s}^{-1}$ over the warm pool sector versus $\sim 20 \, \text{m s}^{-1}$ around the remainder of the equatorial belt.

The corresponding OLR extrema in the pattern for EOF2 of $\Delta \chi$ are separated by only about $60^\circ$ of longitude. Hence, their phase speed is only $\sim 6 \, \text{m s}^{-1}$.

These estimates are roughly comparable to those in previous studies of Knutson and Weickmann (1987), Hendon and Salby (1994), Matthews (2000), and Ventrice et al. (2013).

Because we did not choose any particular amplitude threshold for the PC time series used in the regression, the amplitudes shown here are only about half as large as those estimated in previous studies that have employed compositing to capture strong MJO events.

Of particular interest in this study is the three-dimensional planetary-scale structure of the MJO, as manifested in the zonal wind ($u$) and geopotential height ($Z$) fields. To show how we obtain this structure, we regress the 150-hPa $u$ and $Z$ fields upon various linear combinations of the two leading PC time series of $\Delta \chi$ to produce animations of the MJO cycle. Figure 2 shows a series of frames from these animations at intervals of $1/8$ cycle. Figure 2a corresponds time when the $\Delta \chi$ maximum is passing over the Indian Ocean sector (i.e., the negative peak in PC2), Fig. 2c to the time when the $\Delta \chi$ maximum is passing over the Maritime Continent (i.e., maximum PC1), and Fig. 2e to the time when the $\Delta \chi$ maximum is over the western Pacific (i.e., maximum PC2). Note that the patterns in Figs. 2a and 2e are identical apart from the reversal in polarity. The red circles in this and subsequent figures correspond to the region of maximum $\Delta \chi$, which may be viewed as marking the centroid of the region of large-scale ascent.

At any given time in the MJO cycle the $u$ field, indicated by the contours in Fig. 2, exhibits a single eastward propagating, easterly–westerly couplet, symmetric about the equator with a meridional half-width of about $28^\circ$ of latitude. The couplet extends all the way around the equatorial belt but the easterly and westerly swaths that constitute it range in length from as little as $120^\circ$ of longitude when they pass over the Indo-Pacific warm
pool to as much as 240° when they pass over the cool sector of the tropics. The individual swaths tend to be more narrowly focused on the equator on their eastern end and widen and split into Northern and Southern Hemisphere branches on their western end. This characteristic shape is consistent with Matsuno’s solution for the response to longitudinally periodic equatorial heating anomalies (his Fig. 9), which is dominated by narrow Kelvin waves to the east of the anomalous heating and wider Rossby wave gyres to the west. The zonal wind anomalies in the poleward branches of the gyres merge with the Kelvin waves to the east to form continuous swaths of westerlies and easterlies like the ones shown in Fig. 2. The centers of action in the Δχ field, indicated by red and green circles in Fig. 2, mark the leading edge of successive easterly and westerly Kelvin wave swaths. The tracks of the corresponding OLR centers of action (not shown) are more erratic and not as closely related to the u field.

The eastward-propagating swaths of easterly and westerly anomalies are flanked by Z anomalies centered along 28°N/S: negative anomalies flanking westerly swaths and positive anomalies flanking easterly swaths, in accordance with geostrophy. These features correspond to the subtropical cyclonic and anticyclonic gyres pointed out in Knutson and Weickmann (1987), Rui and Wang (1990), Hendon and Salby (1994), and subsequent studies. These “Rossby wave gyres” can be tracked all the way around the world, but they are strongest when they pass over the warm pool sector. Within 10° of the equator the westerly swaths in Fig. 2 tend to be accompanied by positive Z anomalies and the easterly swaths by negative Z anomalies, which is consistent with the structure of equatorially trapped Kelvin waves (Wheeler et al. 2000; Straub and Kiladis 2002; Yang et al. 2007).

Figure 3 shows latitude–height cross sections of u and Z over the Indian Ocean, the Maritime Continent, and...
the Pacific Ocean, following an upper-tropospheric swath of westerly winds as it propagates eastward around the equator. In all three sectors a feature reminiscent of a Kelvin wave (i.e., an equatorial Z maximum) is clearly discernible, flanked by upper-tropospheric Z anomalies centered near 28°N/S, of opposite polarity to the equatorial anomalies. The centers of the anomalies in the Z field near 28°N/S coincide with the shear zones along the edges of the swaths of westerly anomalies. They are centered at or just below 150 hPa and slightly below the Z extrema in the equatorial Kelvin waves and they slope poleward with height. It is clear from Fig. 3 that the subtropical centers along 45°N/S coincide with the shear zones along the edges of the swaths of westerly anomalies. They are centered near 28°N/S, of opposite polarity to the equatorial Kelvin wave (i.e., an equatorial Z maximum) is clearly discernible, flanked by upper-tropospheric Z anomalies centered near 28°N/S, of opposite polarity to the equatorial anomalies. The centers of the anomalies in the Z field near 28°N/S coincide with the shear zones along the edges of the swaths of westerly anomalies. They are centered at or just below 150 hPa and slightly below the Z extrema in the equatorial Kelvin waves and they slope poleward with height. It is clear from Fig. 3 that the subtropical features in the Z field in Fig. 2 are largely confined to the upper troposphere, where the zonally averaged climatological-mean flow is from the west and in excess of the highest of the MJO-related phase speeds documented earlier in this section. Hence, they can be interpreted as “flanking Rossby waves” and we will refer to them by that term throughout the remainder of this paper.

In the sections for the Maritime Continent and central Pacific sectors in Fig. 3 a set of secondary, midtropospheric centers is evident along 45°N/S and a third set of centers along ~65°N/S. We will discuss these features in further detail in sections 5 and 6. Readers who are mainly interested in that aspect of our study can proceed directly to section 5. In the next section we will examine the vertical structure of the MJO-related variability in the u and Z fields.

4. Modal decomposition of the MJO’s vertical structure

Like other tropical weather systems, the MJO exhibits an internal baroclinic vertical structure with a tendency for a phase reversal between upper- and lower-tropospheric geopotential height and wind fields. The baroclinic modal structure \( Q(p) \) has been described as an analytic function in theoretical studies such as Gill (1980) and Kiladis et al. (2009) or it can also be defined empirically. The most straightforward empirical approach is to perform EOF analysis on a single variable or maximal covariance analysis on a pair of variables in daily gridded data on a stack of pressure levels in an equatorial or tropical domain and extract the leading mode. In this section we use a more MJO-centric approach in which we obtain it as the leading mode in the expansion:

\[
S(x, y, p, t) = \sum_{m=1}^{2} \sum_{n=1}^{N} R_{mn}(x, y) P_m(t)Q_n(p) \quad (m = 1, 2)(n = 1, 2, \ldots, N),
\]

where the \( Q_n \) are an ordered set of mutually orthogonal modal structures and the \( R_{mn} \) are the corresponding regression maps, two for each mode: one corresponding to PC1 of \( \Delta_X \) and the other to PC2.

The analysis protocol is described in Fig. 4. The data matrix consists of stacks of the \( u \) and \( Z \) fields within the prescribed domain regressed upon PC1 and PC2 of \( \Delta_X \). The patterns derived from PC1 and PC2 of \( \Delta_X \) are merged to form single rows in the data matrix. Since the frames in the MJO animations are based on linear combinations of PC1 and PC2, the merged PC1–PC2 data matrix samples all of them. In our analysis, MCA is performed on paired but separate data matrices for \( u \) and \( Z \) obtained in this manner. The domain of the analysis limited to the equatorial belt (10°S–10°N), though the results that we will focus on here are nearly perfectly reproducible in domains extending out to 30°N/S.

The vertical structure function \( Q_n(p) \) for the leading mode derived from this analysis is shown in Fig. 5. Accounting for 80% of the squared covariance between the \( u \) and \( Z \) regression patterns, it corresponds to a deep baroclinic mode, with a node near 400 hPa. The largest upper-tropospheric amplitude is located at 125 hPa for both \( u \) and \( Z \). The largest amplitude in the upper troposphere is located near 750 hPa for \( u \) and near the surface for \( Z \). The \( Z \) profile in Fig. 5 is virtually identical to the leading EOF of the \( Z \) anomalies for the array of grid points in the equatorial belt (10°S–10°N) on randomly selected days.

Patterns of \( u \) and \( Z \) fields in the modal structure, shown in Fig. 6, are analogous to those shown in Fig. 2, but rather than representing data for a single level (150 hPa) they represent data for all 29 levels on the ERA-Interim, weighted in accordance with the \( Q(p) \) profiles shown in Fig. 5 and scaled to their respective values at the 150-hPa level to facilitate comparison with Fig. 2. Comparing Figs. 2 and 6 it is evident that at tropical latitudes the patterns associated with the modal structure derived from our analysis are almost identical to the patterns based on 150-hPa data alone. Hence, it can be said that the modal structure dominates the upper-tropospheric MJO signature.

Figure 7 offers a global perspective on the modal structure, as expressed in the upper-tropospheric (100–300 hPa) \( Z \) field, presented in the format of Fig. 1. The shaded fields in the panels in the right column represent the modal structure and can thus be viewed as an extension of the patterns shown in Figs. 6c,e. The contours...
in those panels show the residual field obtained by subtracting the modal structure from the regression fields shown to the left. It is evident that within 20° of the equator the modal field is dominant, consistent with the inference drawn from Fig. 6, whereas poleward of 45° the residual field, consisting mainly of extratropical wave trains oriented along great circles, is dominant. The extratropical wave trains also show up clearly in the regression patterns based on total upper-tropospheric $Z$ field, shown in the panels in the left column of Fig. 7. We will discuss these relationships further in the next section, after we consider the lower-tropospheric structure of the MJO in light of the modal structure derived from Eq. (2).

Figure 8 shows total, modal, and residual fields for the layer-averaged 500–1000-hPa geopotential height field in the same format as the previous figure. If the modal pattern were dominant at both levels, the upper- and lower-tropospheric patterns would be very similar, apart from a sign reversal. It is clear that this is not the case. The regression patterns for the (total) lower-tropospheric field, shown in the left column, are dominated by an equatorial signature over the Pacific sector that is wider in meridional extent than the upper-tropospheric feature in the modal structure that we have identified as a Kelvin wave. The only features reminiscent of flanking Rossby waves are the equatorially symmetric patches of negative anomalies over the Indian Ocean sector in the pattern based on PC1 of $\Delta \chi$. In combination, the Kelvin and

![Fig. 4. Schematic describing the methodology for decomposing the $u$ and $Z$ fields into modal structures as discussed in section 4. In this paper, the leading mode in the MCA expansion is referred to as the modal structure.](image)

![Fig. 5. Expansion coefficients (ECs) of $u$ (solid) and $Z$ (dashed) in the modal structure derived from the MCA of the vertical profiles of $u$ and $Z$ generated by regressing those fields onto PC1 and PC2 of $\Delta \chi$. The analysis is based on grid points in the 10°S–10°N latitude belt. See text for further explanation. The ECs are standardized. The dashed horizontal lines correspond to the upper-level maxima (150 hPa) in $u$ and the approximate level of the node (400 hPa) in the ECs.](image)
Rossby wave–like features in the lower-tropospheric $u$ and $Z$ fields resemble Gill’s equatorial planetary wave response to an isolated equatorial heat source, in contrast to the more wavelike upper-tropospheric patterns in Fig. 7. However, the Rossby waves to the west of the heat source are wider than in the Gill solution. It is evident from the decomposition in the right column that these elements in the lower-tropospheric pattern project upon the modal structure. The residual field is instrumental in canceling the flanking Rossby waves over the Pacific sector in the modal signature and at the same time widening the swath of low pressure over the equatorial Pacific, which appears to reflect off the Andes in the pattern for PC2 of $\Delta x$. That the residual fields for the upper- and lower-tropospheric regression patterns are so similar indicates that the residual field is mainly barotropic. Hence, it appears that the vertical structure of the MJO signature in $Z$ is mostly represented by only two degrees of freedom, one of which can be represented by a deep, baroclinic modal structure and the residual by a barotropic structure. We have verified that the horizontal patterns associated with the latter strongly resemble those derived from the second mode derived from the decomposition in Eq. (2), from regression patterns based on the 400-hPa $Z$ field, the level of the node in Fig. 5, and from a simple, pressure-weighted average of the $Z$ data at all 29 levels from 1000 to 50 hPa.

The modal structure is also of interest in interpreting the vertical structure of the MJO signature in the $u$ and $Z$ fields in the equatorial belt. Figure 9 shows five-panel sequences analogous to those in Fig. 6 but in the form of equatorial longitude–height sections. The total $u$ and $Z$ anomalies and the residuals formed by subtracting the leading mode in the MCA expansion from the regression fields are also shown. The modal contribution is dominated by the eastward-propagating Kelvin wave–like signature. However, in contrast to the linear Kelvin wave structure [e.g., as depicted in Fig. 8 of Matsuno (1966)], the features in the $u$ field are slightly displaced to the east of the corresponding features in the $Z$ field. This systematic displacement is indicative of a tendency for zonal flow directed down the horizontal pressure gradient and a generation of kinetic energy (KE), as noted by Seiki and Takayabu (2007) and Zhou et al. (2012).

As in the corresponding $\Delta x$ patterns shown in Fig. 1, zonal wavenumber 1 is dominant in the modal structure shown in Fig. 9, but as successive $u$ and $Z$ maxima and minima pass over the warm pool sector their zonal wavenumber increases and they amplify. A weak signature of the Andes, manifested as an interruption of the eastward propagation in the $Z$ signature near 80$^\circ$W longitude, is discernible in a few of the panels. The $u$ and $Z$ patterns associated with the residual in the equatorial belt are weaker and more difficult to interpret, displaying both eastward-propagating and standing components. At the lower levels they are mainly barotropic and in the tropical tropopause transition layer (TTL) and lower stratosphere they exhibit features reminiscent of the downward-propagating planetary waves (Kiladis et al. 2005; Tian et al. 2007; Virts and Wallace 2013). Figure 9 provides strong support for the notion that the $(x, y)$ and $p$ dependences of the MJO are separable in the equatorial plane and that a single mode is sufficient to capture the simplest and most essential features of its vertical structure.

5. MJO-related extratropical features

If the extratropical wave trains in the upper-tropospheric $Z$ field (Fig. 7, left columns) were directly forced by the
divergent outflow from regions enhanced equatorial convection, as is often presumed (Jin and Hoskins 1995; Mori and Watanabe 2008; Seo and Son 2012; Frederiksen and Lin 2013), they should emanate from the sectors of the subtropics poleward of the equatorial maxima in the $\Delta \chi$ field where the upper-level (meridional) outflow from the anomalous convection is crossing the climatological-mean jet stream (Sardeshmukh and Hoskins 1988). But it is clear from the panels in the left column of Fig. 7 that the wave trains do not emanate from these regions but from the flanking Rossby waves, which are centered at the longitude of the nodes in the $\Delta \chi$ field.

For example, the strongest of the extratropical wave trains, the one that resembles the PNA pattern, is most prominent at the time when the region of enhanced equatorial convection and the associated maximum in $\Delta \chi$ is passing over the Maritime Continent 90° of longitude to the west (i.e., the time of maximum PC1). In this phase of the MJO cycle, the upper-tropospheric zonal outflow from the convection is strongest over the central Pacific and it is flanked by a cyclonic gyre to the north of Hawaii, which is the first of the four centers of action in the PNA pattern (Wallace and Gutzler 1981). While many previous studies have documented the existence of a PNA-like pattern occurring in association with the MJO (Ferranti et al. 1990; Mori and Watanabe 2008; Roundy et al. 2010), Fig. 7 shows clearly that the anomalies originate over the tropical central Pacific (e.g., as reported by Magaña and Yanai 1991). The Southern Hemisphere wave trains in Fig. 7 resemble the Northern Hemisphere patterns described above, and also appear to emanate from the flanking Rossby waves.

At most longitudes the climatological-mean jet streams, with their strong meridional gradient of potential vorticity, are analogous to a set of bookends bracketing the MJO signature in upper-tropospheric $u$ and $Z$. The flanking Rossby waves described in this study correspond to the outermost pair of books. It is only over certain limited sectors of the globe that the MJO excites a strong extratropical response. In the Northern Hemisphere this active sector corresponds to the climatological-mean jet exit region, as shown in Fig. 10. In this five-panel sequence the same upper-tropospheric $Z$ anomalies and the associated $u$ anomalies averaged over the same layer are shown superimposed upon the annual climatological-mean $u$ field in the same layer as in Fig. 7, which indicates the position of the jet streams. It can be seen that as the MJO-related flanking Rossby waves propagate eastward...
toward the climatological-mean jet exit region, they speed up (Figs. 10a,b) and then they appear to slow down as they pass through it (Figs. 10b–d), while an extratropical PNA-like wave train develops and amplifies along their poleward flank. The sequence is repeated with reversed polarity when the flanking Rossby wave anticyclones propagate into the same region a half-cycle later, starting in Fig. 10e. At the time when PC1 is a maximum (Fig. 10c), the MJO-related easterly anomalies along 32°N at the date line overlie the climatological-mean jet exit region and are thereby capable of extracting kinetic energy from the mean flow, as described in Simmons et al. (1983) and Nakamura et al. (1987). This coincidence might explain why the MJO forces a strong extratropical response in the Northern Hemisphere only over the central Pacific. In Fig. 10c strong flanking Rossby waves are also seen over the Indian Ocean sector just to the south of the climatological-mean jet streams, causing the jet stream to vary in strength, as noted by Yang and Webster (1990), Matthews and Kiladis (1999), and Moore et al. (2010). In this case, however, there is no extratropical response.

At the time of maximum PC2 there are indications of extratropical wave trains emanating from the flanking Rossby waves along 120°E in both hemispheres. A dynamical interpretation of these and other more subtle features in Fig. 7 is beyond the scope of this study.

6. A more global perspective

All the diagrams in the preceding sections are representative of conditions observed at specific phases of the MJO cycle. We can gain further insight into the relation between the MJO-related $u$ and $Z$ perturbations and the background flow by considering the root-mean-squared (rms) variance of the respective fields averaged over the MJO cycle. For this purpose it is sufficient to calculate the square root of the sum of the squared amplitude of the anomalies in regression maps ($D$) based on PC1 and PC2 of $\Delta X$, ($D_{rms} = \sqrt{D_1^2 + D_2^2}$), since other “phases” in the MJO cycle are all based on linear combinations of these two sets of regression patterns. Figure 11 shows the rms amplitude of the $Z$ field averaged over the 100–300-hPa layer, calculated in this manner, superimposed upon the annual-mean, climatological-mean zonal wind field averaged over the same layer. The pattern is dominated by elongated equatorially symmetric maxima centered along 28°N/S, just equatorward of the climatological-mean jet streams and well within the westerly wind belts, where the MJO-related perturbations are propagating westward relative to the background flow. Consistent with Figs. 3, 6 and 7, these subtropical features are most clearly defined over the Indo-Pacific warm pool region. The weaker equatorial maximum, which is most prominent at longitudes remote from the
warm pool, appears to be the signature of a Kelvin wave. The zonal wind anomalies shown in the bottom panel span the latitude band equatorward of the variance maximum in $Z$ along $\sim 28^\circ$N/S. The locations of the MJO-related variance maxima and their relation to the climatological-mean background flow in Fig. 11 suggest the existence of equatorially symmetric upper-tropospheric “Rossby wave corridors” centered along $\sim 28^\circ$N/S in the region of anticyclonic shear, just equatorward of the jet streams.

The corresponding variance patterns for the lower-tropospheric layer (500–1000 hPa) shown in Fig. 12 are quite different. The most prominent feature in the $Z$ field is the meridionally broad, equatorially symmetric maximum over the Pacific sector. The Rossby wave signature over the Indian Ocean sector in the top left panel of Fig. 8 is barely visible in Fig. 12 and there is no indication of flanking Rossby waves at other longitudes. The effect of the Andes in blocking the eastward propagation of the MJO is much more clearly apparent in the lower troposphere than at upper levels. The corresponding features in the $u$ field are weaker and more regional than their upper-tropospheric counterparts, the
most prominent being the maximum over the warm pool.

Root-mean-squared MJO-related variance maps for $u$ and $Z$, analogous to those shown in Figs. 11 and 12, were generated for each pressure level and zonally averaged to create the meridional section shown in Fig. 13. Consistent with previous figures, the $u$ signature, indicated by the contours, is mainly confined to the upper troposphere, equatorward of the centers of the flanking Rossby waves, indicated by the colored shading. Secondary $u$ centers, where the fluctuations are of opposing polarity to those in the tropics, are located poleward of the flanking Rossby waves at the latitude of the climatological-mean westerly jet stream (W). A variance maximum in $Z$ corresponding to the MJO’s equatorial Kelvin wave is clearly discernible, centered at the 125-hPa level. Extratropical “centers of action” in the $Z$ field centered along 45°N/S are associated with the
extratropical wave trains discussed in the previous section. In contrast to the tropical features, which are mainly upper tropospheric, these extratropical features extend all the way downward to Earth’s surface. There is also a hint of a stratospheric signal centered ~65°N/S.

7. Discussion

The main findings of this paper relate to what we have referred to as the “flanking Rossby waves.” The existence of these features has been noted in numerous studies dating back to Knutson and Weickmann (1987), Rui and Wang (1990), and Hendon and Salby (1994). In some studies they have been referred to as cyclones (anticyclones) or cyclonic (anticyclonic) gyres, evidently because authors have been hesitant to identify eastward-propagating features as Rossby waves. They have sometimes been referred to as equatorial Rossby waves and sometimes as subtropical. Signatures of these features are evident in the distribution of tracers in the TTL: positive ozone anomalies and negative anomalies in relative humidity and carbon monoxide, suggestive of a lowering of the tropopause, above the cyclonic Rossby gyres and vice versa (Tian et al. 2007; Wong and Dessler 2007; Virts and Wallace 2013).

We have confirmed that these nondivergent circulations are indeed planetary-scale Rossby waves. Their track, as defined by the MJO-related variance maxima in the upper-tropospheric geopotential height field, lies along 28°N/S, just equatorward of the climatological-mean westerly jet streams. At this latitude, the upper-tropospheric westerlies are much stronger than the average phase speed of the MJO. Hence, even though the phase speed of the flanking Rossby waves is eastward relative to Earth’s surface, it is westward relative to the zonally averaged flow in which they are embedded. The flanking Rossby waves are thus trapped in a westerly waveguide, and in this sense are somewhat different from the equatorially trapped Rossby waves in the Matsuno–Gill theoretical solutions for the linear, steady state response to equatorial heat sources and sinks. Their track is about twice as far from the equator as that of equatorial Rossby waves [e.g., see Figs. 1 and 8 of Yang et al. (2007)]. The structure of MJO-related wave packets consisting of Kelvin waves and Rossby waves propagating eastward in tandem is summarized in the wind and height regression maps shown in Fig. 14. Despite their
much larger meridional widths, the observed, flanking Rossby waves bear a surprisingly strong resemblance to the equatorially trapped Rossby waves in the theoretical solutions. Not only do they propagate eastward in tandem with the Kelvin waves, they lengthen in longitude as the Kelvin waves lengthen and vice versa. The meridional placement of the flanking Rossby waves appears to be determined not by the Matsuno–Gill solution, but by the zonal-mean zonal wind climatology.

Ours is not the first study to attribute the presence of flanking Rossby waves to the presence of a westerly waveguide. On the basis of numerical experiments with a two-layer shallow-water wave equation model, linearized about the observed November–April basic-state flow and forced by an MJO-like equatorial heat source, Barlow (2012) obtained a realistic solution that contains features with tracks similar to those of the flanking Rossby waves. When the same forcing was prescribed in the Matsuno–Gill model with a resting basic state, the induced Rossby waves were more narrowly confined to the equatorial belt.

The MJO has been widely regarded as a tropical phenomenon that, under some conditions, is capable of forcing extratropical wave trains. When the flanking Rossby waves are taken into account, planetary wave propagation in the westerlies is seen as playing a role in the dynamics of the MJO. Extratropical wave trains such as the PNA pattern are no longer necessarily interpreted as being forced almost exclusively by the divergent outflow from the MJO’s equatorial convection [e.g., as in the recent studies of Mori and Watanabe (2008) and Frederiksen and Lin (2013)] but as emanating from the flanking Rossby waves to the east and west of the convection. At most longitudes, the jet streams appear to function like bookends, confining the MJO signature to the tropics and subtropics. However, when the flanking Rossby waves pass through the climatological-mean jet exit regions, the zonal wind anomalies poleward of the highs or lows in the flanking Rossby waves are able to extract the kinetic energy from the climatological-mean flow to force a poleward-propagating wave train. This interpretation is substantiated by results of recent numerical experiments by Bao and Hartmann (2014) with a nonlinear shallow-water model in which they specified equatorial mass sources designed to mimic the MJO-related diabatic heating. They imposed an extratropical basic-state flow with a jet exit region over the North Pacific that mimics the wintertime climatology. They found that when their simulated heating passes over the Maritime Continent it excites an extratropical response.

**FIG. 14.** Geopotential height (Z) (contours) and wind vector anomalies (arrows) in the modal structure regressed on PC1 and PC2 of Δχ arranged as in Fig. 1. Colored shading represents upward (green shading) and downward (pink shading) motion at 400 hPa. Contour interval is 2 m and shading interval is 10 hPa day$^{-1}$. The red (green) circles indicate Δχ maxima (minima) and are sized in accordance with the amplitude of the Δχ maximum (minimum).
emanating from the North Pacific jet exit region that resembles the observed response. In interpreting their results they argue that the amplification in this region is mainly a consequence of the rapid propagation of the Rossby wave within the jet itself and the extraction of kinetic energy from the mean flow just downstream of it.

Our study’s main conclusions follow from conventional regression analysis performed upon the \( u \) and \( Z \) fields. There are two novel elements in our analysis protocol, but the results described above are in no way dependent on them. The first new element is our use of a univariate MJO index based upon the \( \chi \) field, as opposed to using an existing multivariate index such as the RMM and VPM. Our main motivation for experimenting with a \( \chi \)-based index was to be able to examine the interrelationships between the unadulterated \( u \), \( \chi \), and OLR fields. Kiladis et al. (2014) have already experimented with a univariate index based on the OLR field; concomitantly and independently we have performed a similar analysis on the \( \chi \) field.

Many authors have expressed reservations concerning the dynamical interpretation of the EOFs of the \( \chi \) field. Based on the results of this study, we have concluded the following:

- Although the MJO-related convection is most highly organized over the warm pool sector, it is not confined to those longitudes. It is clear from Fig. 1 that patches of negative OLR anomalies appear within the region of positive \( \Delta \chi \) anomalies as it propagates all the way around the equatorial belt.
- In the upper troposphere the leading edges of successive easterly and westerly Kelvin wave surges can be tracked as they propagate all the way around the equatorial belt in tandem with the \( \Delta \chi \) extrema, as documented in Figs. 2 and 6. This strong correspondence is indicative of a robust relationship between the \( u \) and \( \chi \) fields and it supports the notion that the convergence–divergence associated with these planetary waves, the \(-\partial u/\partial x\) term in particular, plays an important role in defining conditions for the occurrence of MJO-related equatorial convection throughout the equatorial belt. A study of the planetary wave–related lower-tropospheric divergence field and its relation to the shape of vertical motion field in the MJO cycle is currently underway.
- Although the patterns in the leading EOFs of the \( \chi \) field appear rather bland and featureless, regression patterns based on their PCs contain information relating to a wide range of scales. For example, it is shown in the appendix that they reveal the structure of the \( u \) anomalies almost as well as the zonal wind and OLR-based RMM index of Wheeler and Hendon (2004) or variants of this index based on zonal wind alone (Straub 2013).

Our results based on the \( \chi \) field presented in Fig. 1 serve to highlight the diversity of MJO-related eastward phase speeds, which range from 6 m s\(^{-1}\) for the convection over the warm pool sector to as high as 20 m s\(^{-1}\) for the \( \chi \) and \( u \) anomalies in the less active sector of the tropics.

The other novel element in this study is an analysis of the vertical structure of the global MJO signature in the \( Z \) field. We have shown that the basic patterns in the MJO-related \( Z \) anomalies can be represented as a baroclinic modal structure with a node at the 400-hPa level that appears with opposing polarity in the upper and lower troposphere, plus a residual field that is predominantly barotropic. In the upper-tropospheric \( Z \) field the modal structure, characterized by eastward-propagating equatorial Kelvin waves and flanking Rossby waves, is dominant in the tropics and more barotropic wave trains oriented along great circles are dominant at higher latitudes. In the tropical lower-tropospheric \( Z \) field the two components are of comparable amplitude and in combination they form a pattern reminiscent of the response to a pulsating heat source over the Maritime Continent in a domain bounded on the east side by the Andes. That the upper-tropospheric pattern is more wave-like and exhibits more pronounced eastward propagation could be due to the presence of the westerly waveguides and/or to the diminished influence of the Andes in blocking the equatorial Kelvin wave pulse at the higher levels.

An analysis of the MJO-related temperature, vertical velocity, and horizontal divergence fields based on the methodology used in this paper is performed in Adames and Wallace (2014, manuscript submitted to J. Atmos. Sci.). These fields exhibit signatures that are closely associated with the modal structure in the zonal wind and geopotential height field identified in this paper.

Acknowledgments. We thank Brian Smoliak and Katrina Virts for their time and support of this project. We would also like to thank Tsubasa Kohyama for contributing the idea and first draft of Fig. 4. We appreciate the suggestions and comments from Chidong.
Zhang, George Kiladis, and three anonymous reviewers. We would also like to thank Beth Tully for improving the graphics. This research was supported by the National Science Foundation’s Graduate Research Fellowship Program (NSF-GRFP) Grant DGE-0718124. The work of the second author was funded by the National Science Foundation under Grant ATM 1122989.

**APPENDIX**

**Intercomparison of MJO Indices**

Here we compare the patterns and the modal structure derived in section 4 with different MJO indices. For this comparison we use our differential velocity potential

---

**FIG. A1.** The leading expansion coefficients for (left) \( u \) and (right) \( Z \) as in Fig. 5, derived from the MCA of the vertical profiles of \( u \) and \( Z \) generated from regression fields of PC1 and PC2 of the MJO indices shown in the legend.

**FIG. A2.** Regression patterns based on PC1 of (top to bottom) the different MJO indices listed in the top-left corner of the left-column panels. (left) As in Fig. 1a, and (middle) as in Fig. 6c. (right) The 100–300-hPa-layer-mean \( Z \) for the Northern Hemisphere, based on data for all 12 calendar months. Contour interval is 2.5 m.
index $\Delta \chi$, an index solely based on $\chi$ at 150hPa ($\chi_{150}$), the real-time multivariate MJO index (RMM; Wheeler and Hendon 2004), the velocity potential multivariate MJO index (VPM; Ventrice et al. 2013), and the OLR-based MJO index (OMI; Kiladis et al. 2014). For all indices the period of record 1979–2011 is used, except for the VPM, for which the period 1989–2011 is used. For each index the modal structure is calculated following the same procedure as depicted in Fig. 4. The regression coefficient profiles for each different MJO index are shown in Fig. A1. The modal structure $Q$ is nearly the same for all indices, corresponding to the deep, baroclinic mode discussed in section 4.

Regression patterns based on PC1 of the different MJO indices are shown in Fig. A2. The panels are arranged such that the patterns derived from the $\chi$-based indices are at the top, those derived from multivariate indices are at the bottom. The distinctions between the patterns derived from the different indices are subtle. The regression pattern for the OMI, which is based on OLR alone, are somewhat weaker and more confined to the tropics, the flanking Rossby waves are less prominent, and the PNA pattern is less clearly recognizable than in the patterns based on the other indices. The $u$-based indices exhibit a slightly stronger equatorial Pacific Kelvin wave signature in $u$ than patterns based on the other indices. The corresponding regression patterns based on PC2 of $\Delta \chi$ (not shown) are also remarkably similar, except that the OMI displays a stronger OLR signature than the other indices do. The overall lack of sensitivity of the patterns to the choice of index on which they are based lends credence to the results presented here and to studies based on other indices.

REFERENCES


