Changes in the Activity of the Madden–Julian Oscillation during 1958–2004

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(Manuscript received 13 April 2005, in final form 12 April 2006)

ABSTRACT

The Madden–Julian oscillation (MJO) is the most prominent mode of tropical intraseasonal variability. This study investigates the following questions. Is there statistical evidence of linear trends in MJO activity since the mid-1970s? Does the MJO exhibit changes in regimes of high and low activity? Are there significant seasonal differences in the activity of the MJO on time scales longer than interannual?

Positive linear trends are observed in zonal wind anomalies at 200 (U200) and 850 (U850) hPa during summer and winter seasons. Positive trends are also observed in the number of summer MJO events. Resampling statistical tests indicate that positive trends in summer U200 and U850 anomalies are statistically different from random occurrences at a 5% significance level.

A methodology based on the number of events is used to characterize low-frequency (LF) changes in MJO activity. Mean winter LF activity was characterized by nearly uniform variability from the early 1960s until the mid-1990s. In contrast, mean summer LF changes showed a regime of high activity from the mid-1960s until the late 1970s, a low regime from 1980 to 1988, and a regime of high activity from the early 1990s to early 2000. Fourier analysis of the mean summer LF index indicates that regimes of high MJO activity were separated by 18.5 yr. The substantial changes in summer MJO regimes do not appear to be related to increases in observational samplings due to satellite-derived winds assimilated in the NCEP–NCAR reanalysis. Monte Carlo experiments indicate that the observed changes in regimes of MJO activity in summer are statistically different from random occurrences at the 10% significance level only.

1. Introduction

The Madden–Julian oscillation (MJO) is the most prominent mode of tropical intraseasonal variability (Madden and Julian 1994; Lau and Waliser 2005). The MJO influences the monsoons in Asia, Australia, and the Americas (Yasunari 1979; Lau and Chan 1986; Mo 2000; Nogues-Paegle et al. 2000; Goswami and Mohan 2001; Higgins and Shi 2001; Jones and Carvalho 2002). This modulation affects rainfall and extreme events in many locations around the world (Mo and Higgins 1998; Higgins et al. 2000; Jones 2000; Bond and Vecchi 2003; Carvalho et al. 2004; Jones et al. 2004c). Since the

MJO involves intense tropical convective heating anomalies, tropical–extratropical interactions are significant during its life cycle. Therefore, some studies have detected noticeable impacts on the skills of weather forecasts on medium-to-extended ranges (Ferranti et al. 1990; Lau and Chang 1992; Hendon et al. 2000; Jones and Schemm 2000; Jones et al. 2004a) and potential predictability (Waliser et al. 2003; Jones et al. 2004d,c).

An important characteristic of the MJO is the relatively well defined seasonality of its behavior. Many previous studies have shown that the MJO is very active during boreal winter and spring and generally exhibits equatorial eastward propagation. Likewise, MJO events show similar spatial scales and periodicity during boreal summers, but the associated convective anomalies have substantial meridional propagations (Wang and Rui 1990; Madden and Julian 1994; Lawrence and

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Webster 2001; Jones et al. 2004b). To explain the seasonal behavior of the MJO, Salby et al. (1994) proposed that the MJO is controlled by the latitudinal position of tropical heat sources such that the greatest amplifications occur in the spring equinox when maximum sea surface temperatures (SSTs) are near the equator.

On time scales longer than the seasonal cycle, many studies have demonstrated that the MJO shows large year-to-year variations (Wang and Rui 1990; Gutzler 1991; Weickmann 1991; Madden and Julian 1994; Jones et al. 2004b). Because the passage of MJO events over the western Pacific Ocean can significantly modify the thermocline structure in the equatorial eastern Pacific via their connection to westerly wind bursts (McPhaden and Taft 1988; Kessler et al. 1995; Hendon et al. 1998), this latter interaction has been suggested as important in the evolution of El Niño–Southern Oscillation (ENSO; Lau and Chan 1986; Weickmann 1991; McPhaden 1999, 2004). This has motivated the idea that the MJO can act as stochastic forcing and may explain part of the ENSO irregularity (e.g., Zavala-Garay et al. 2003, 2005).

Still in the context of interannual variations, one of the outstanding issues is what drives the interannual behavior of the MJO. Observational studies indicate that before the onset of warm ENSO episodes, as SST anomalies in the central Pacific become sufficiently positive, MJO events tend to propagate farther east into the equatorial Pacific (e.g., Jones et al. 2004c; McPhaden 1999, 2004; Weickmann 1991; Zhang 2005). Despite this modulation, indexes of MJO activity in the boreal winter are uncorrelated with SST anomalies associated with ENSO (Hendon et al. 1999; Slingo et al. 1999). Recently, Teng and Wang (2003) investigated the interannual variability of the MJO in boreal summers and found that warm ENSO episodes affect the northwestward-propagating MJO mode in the western North Pacific by changes in the mean circulation. Their results indicate that MJO events occurring in May–July are strengthened in warm ENSO years and therefore suggest a seasonal dependency on the MJO–ENSO relationship.

The motivation for the present study is based on the fact that the behavior of the MJO on time scales longer than interannual is virtually unknown. In a previous study, Slingo et al. (1999) investigated interannual variations in the MJO and relationships with ENSO using the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis from 1958 to 1997. They proposed an index of MJO activity and found only weak correlations with SST anomalies associated with ENSO. This suggests that interannual variations in the

MJO during boreal winters are largely controlled by internal atmospheric dynamics rather than lower boundary conditions in the tropical Indian and Pacific Oceans. An interesting result from their study is a clear trend in their index of intraseasonal activity such that the amplitudes in the period 1958–76 were consistently lower than in those in 1977–96. They pointed out the possibility that the trend might be an artifact of the reanalysis, since the observational sampling increased significantly with the introduction of satellite-derived winds in about 1979. Nevertheless, another important result came from a 45-yr simulation with the Hadley Centre climate model forced with observed SST (1949–93). The model results reproduced the positive trend in intraseasonal activity since the mid-1970s, suggesting that the trend in intraseasonal activity could be real. Thus, Slingo et al. (1999) speculated that the MJO may become more active as tropical oceans experience prolonged warming.

The objective of this study is to present an observational analysis characterizing the activity of the MJO during 1958–2004 with emphasis on time scales longer than interannual. Specifically, this paper investigates the following questions. Is there statistical evidence of a steady increase (i.e., linear trend) in MJO activity since the mid-1970s? Does the MJO exhibit changes in regimes of high and low activity? Are there significant seasonal differences in the activity of the MJO on time scales longer than interannual? In section 2, we first present the datasets used. The methodology to identify MJO events is discussed in section 3. Section 4 shows a statistical analysis of trends in MJO activity. Changes in regimes of MJO activity are examined in section 5. Last, a summary and conclusions are presented in section 6.

2. Data

The term intraseasonal oscillation (ISO) is frequently referred to boreal summer eastward-propagating events, while MJO is associated with boreal winter occurrences (Lau and Waliser 2005). For simplicity, we refer to MJO as tropical eastward-propagating events occurring throughout the year. In this paper, the activity of the MJO was investigated using NCEP–NCAR reanalysis (Kalnay et al. 1996; Kistler et al. 2001). Pentads (5-day nonoverlapping means) of the zonal components of the wind at 200 (U200) and 850 hPa (U850) were used for the period 1958–2004 and the global Tropics (30°S–30°N latitude, all longitudes). The analysis was performed on extended winter and summer seasons separately, which were defined from 2–6 November to 26–30 April and from 1–5 May to 28 October–1 November, respectively.

It is opportune to point out a caveat when analyzing the entire record of reanalyzed fields. The NCEP–NCAR reanalyses are derived with a fixed state-of-the-art global numerical model and data assimilation system. Although reanalysis provides an extremely valuable resource to study climate variations, an inherent difficulty when using the entire data record is that the number of observations has increased in time. In particular, the introduction of satellite-derived data has dramatically increased the number of observations (see Kistler et al. 2001 for a comprehensive review). Having this potential difficulty in mind, this paper uses NCEP–NCAR reanalysis to examine possible changes in the long-term behavior of the MJO.

In general, interannual changes in the MJO based on NCEP–NCAR reanalysis are consistent with the 40-yr European Centre for Medium Range Forecasts (ECMWF) Re-Analysis (ERA-40; Slingo et al. 2005). For this reason, the results presented in this study are limited to NCEP–NCAR data, given the more extensive data record. An attempt (not shown) was also made to analyze the activity of the MJO using radiosonde observations from the Comprehensive Aerological Reference Data Set (CARDS; Eskridge et al. 1995). However, the irregular spatial distribution of the stations in the Tropics and missing data prevented a direct comparison with results obtained from reanalysis.

Outgoing longwave radiation (OLR) is used as a proxy for large-scale tropical convection (e.g., Jones et al. 1998, 2004b). Pentads of OLR (2.5° latitude $\times$ 2.5° longitude) from 1979 to 2004 (Liebmann and Smith 1996) were used to check consistency with U200 and U850 results. Last, to isolate the MJO signal, time series of U200, U850, and OLR were detrended and fast Fourier transform (FFT) was used to retain variations of 20–100 days.

3. Identification of MJO events

The MJO has been extensively studied over the years and different methods have been used to characterize the oscillation (e.g., Hendon et al. 1999; Lin et al. 2004). In general, these methods use bandpassed anomalies of a single parameter or combination of parameters and include 1) reference time series spatially averaged over specific locations (e.g., Hendon and Salby 1994; Jones et al. 1998; Kiladis and Weickmann 1992; Lau and Chan 1986; Lin et al. 2004); 2) empirical orthogonal function (EOF) analysis (e.g., Hendon et al. 1999; Jones et al. 2004d; Kessler 2001; Lau and Chan 1986; Weickmann 1983; Wheeler and Hendon 2004); 3) eastward propagation of contiguous OLR anomalies (Wang and Rui 1990; Jones et al. 2004b); 4) space–time filtering of eastward-propagating anomalies (Hendon et al. 1999); and

5) equatorial zonal mean intraseasonal index (Slingo et al. 1999; Sperber 2003). It is important to recognize, however, that some of the methods aim to specifically identify individual MJO occurrences, while others provide proxies for levels of tropical intraseasonal activity.

This study uses the EOF method (Preisendorfer 1988; Wilks 1995) to identify MJO events and is based on the following procedure. Winter and summer seasons were treated separately to account for seasonal variations in MJO activity. The spatial domain covers 15°S – 15°N and 30°E – 150°W , where the MJO is most active in both seasons (e.g., Lau and Waliser 2005). The narrow latitudinal extent serves to minimize extratropical influences on the EOF calculation, while it is still large enough to capture MJO events with maximum amplitudes on both sides of the equator. Likewise, the longitudinal extent was designed to emphasize intraseasonal variability in the Indian Ocean and the western Pacific.

In the period 1979–2004, EOF analysis of OLR anomalies was computed on the covariance matrix. In contrast, combined EOF (CEOF) analysis of U200 and U850 anomalies was performed on the correlation matrix in the period 1958–2004. Prior to EOF calculations, all time series were scaled by the square root of the cosine of the latitude. The reasons for performing CEOF of U200 and U850 anomalies and correlation matrix are discussed later in this section. Last, principal components (PCs) and eigenvectors were scaled with the square root of the corresponding eigenvalues (Wilks 1995).

Table 1 shows eigenvalues separation, sampling error, and percentage of explained variance. Sampling errors were estimated following the criterion of North et al. (1982) and degrees of freedom estimated as the number of pentads in the time series divided by 10 pentads, which is the typical time scale of intraseasonal variations. In the summer EOF analysis of OLR anomalies (top left), the first two eigenvalues are separated from each other and explain about 24.8% of the total variance. The third eigenvalue (6.42%) is separated from the fourth eigenvalue (6.42%). Somewhat similar results were obtained for the summer CEOF analysis of U200 and U850 anomalies (bottom left), in which the first two eigenvalues represent $\sim 25.6\%$ of the total variance and the third eigenvalue (8.36%) is also separated from the fourth. The winter EOF analysis of OLR anomalies (top right) indicates that the first two eigenvalues are separated from each other and account for about 26.7% of the total variance. Note that the third eigenvalue explains 6.13% of the total variance and is not separated from the fourth eigenvalue. In the winter CEOF analysis of U200 and U850 anomalies

TABLE 1. Results of EOF analysis: separation of eigenvalues, sampling error, and percentage of explained variance. (top) EOF of OLR anomalies; (bottom) combined EOF of U200 and U850 anomalies.

EOF-OLR						
<i>N</i>	Summer			Winter		
	Separation	Error	% variance	Separation	Error	% variance
1	13 482.33	5417.54	15.11	7383.19	6597.43	14.55
2	8109.81	3471.53	9.68	18 246.69	5496.81	12.13
3	3137.48	2300.98	6.42	2426.48	2776.75	6.13
4	1503.36	1848.12	5.15	4085.01	2370.31	5.23
5	3256.77	1631.13	4.55	1697.22	1761.36	3.89
6	535.00	1161.06	3.24	1065.68	1508.35	3.33
7	1172.59	1083.84	3.02	403.91	1349.49	2.98
8	1052.01	914.59	2.55	1316.32	1289.28	2.84
9	242.61	762.74	2.13	589.79	1093.05	2.41
10	496.17	727.73	2.03	862.47	1005.13	2.22
CEOF-U200 U850						
<i>N</i>	Summer			Winter		
	Separation	Error	% variance	Separation	Error	% variance
1	28.35	25.42	13.59	174.84	38.97	21.37
2	62.64	22.37	11.96	65.64	19.72	10.82
3	58.56	15.64	8.36	28.94	12.49	6.85
4	16.63	9.34	5.00	15.15	9.31	5.11
5	3.41	7.55	4.04	12.06	7.64	4.19
6	5.09	7.18	3.84	8.89	6.31	3.46
7	8.09	6.64	3.55	1.68	5.33	2.93
8	3.28	5.77	3.09	5.82	5.15	2.82
9	6.41	5.42	2.90	7.36	4.51	2.47
10	2.71	4.73	2.53	0.95	3.70	2.03

(bottom right), the first two eigenvalues are separated from each other and represent 32.2% of the total variance. Eigenvalues 3 to 6 also indicate separation, although they explain small fractions of the total variance.

Consistent with previous studies (e.g., Knutson and Weickmann 1987; Lo and Hendon 2000; Sperber 2003; Jones et al. 2004b), the results shown in Table 1 indicate that the first two eigenvalues capture the bulk of the MJO variability and together describe the eastward propagation of the oscillation. To demonstrate the spatial patterns, Fig. 1 shows the first and second eigenvectors obtained from OLR anomalies during winter and summer seasons. The dipoles of negative and positive OLR anomalies indicate the eastward propagation of enhanced and suppressed convective activity from the Indian Ocean to the western Pacific. In addition to eastward propagation, the EOFs in summer also show meridional propagation over Southeast Asia (Figs. 1c,d). Similarly, the first and second eigenvectors of U200 and U850 anomalies (Figs. 2 and 3) also show eastward propagation from the Indian Ocean to the western Pacific and illustrate the baroclinic structure of the MJO in zonal wind anomalies in the upper- and lower-tropospheric levels in both seasons.

The identification of individual MJO events was accomplished by examining the amplitudes of the first two principal components (PC1 and PC2) in winters and summers separately. Events were defined when 1) the amplitude of winter (summer) PC1 was positive and exceeded one winter (summer) standard deviation (σ), and 2) within 20 days after the winter (summer) PC1 exceeded 1σ , the amplitude of the winter (summer) PC2 was also positive and exceeded 1σ . A similar approach was used by Shinoda et al. (1998) with OLR anomalies, although they specified 1.5σ as cutoff value, which emphasize strong events. In general, 1σ of the frequency distribution of the PCs corresponds approximately to the 80th–85th percentiles. Since the objective of this study is to investigate trends and shifts in MJO regimes, we opted for 1σ as a cutoff value so that typical MJO events were identified throughout the entire record. Last, each individual MJO occurrence was registered at the pentad corresponding to the peak in the winter (summer) PC1.

Based on EOF analysis of OLR anomalies, a total of 41 winter events and 41 summer events were identified during 1979–2004. Likewise, 99 winter events and 85 summer MJO events were identified with the combined (U200 and U850) EOF approach during 1958–2004,

FIG. 1. First two eigenvector (EOF1 and EOF2) of EOF analysis of OLR anomalies during (a), (b) winter and (c), (d) summer. Eigenvectors are expressed as correlations between corresponding principal components and OLR anomalies. Solid (dashed) contours indicate positive (negative) correlations at 0.1 intervals. Zero contours omitted.

FIG. 2. Same as in Fig. 1, but of combined EOF analysis of U200 anomalies. Eigenvectors are expressed as correlations between corresponding principal components and U200 anomalies.

FIG. 3. Same as in Fig. 2, but for U850 anomalies.

such that 56 winter and 50 summer events occurred in 1979–2004. The difference in the number of winter and summer events between both approaches during 1979–2004 is not entirely surprising, since they are based on different metrics of the MJO (i.e., convective versus circulation signals). Despite these differences in the number of events, changes in MJO regimes are relatively consistent between both approaches (section 5). Lag composites of OLR, U200, and U850 anomalies during the samples of winter and summer MJO events exhibit the characteristic life cycle of the oscillation (not shown).

To summarize the amplitudes of the MJO events, an index was computed as $PC_{\text{index}} = (PC1^2 + PC2^2)^{1/2}$ (e.g., Mathews 2000). Figure 4 shows PC_{index} amplitudes of all MJO events identified with OLR (top) and (U200 and U850) (bottom) anomalies and each vertical bar represents an event registered at the peak in the corresponding PC1. The irregularity of MJO occurrences is quite evident in both records. Moreover, we recall that events identified with EOF of OLR anomalies were derived from the covariance matrix, and PC_{index} carries physical units. Also, there is no apparent linear trend in PC_{index} of OLR anomalies, although the short OLR data record may be a limiting factor to infer changes in MJO activity. In contrast, events identified with CEOF analysis of (U200 and U850) anomalies were derived from the correlation matrix. The first rea-

son for using the correlation matrix is because U200 and U850 anomalies have different variances (e.g., Wilks 1995). The second and less obvious reason is due to the nonstationarity of U200 anomalies. Inspection of time series of U200 anomalies in the tropical western Indian Ocean revealed substantial positive linear trends during 1958–2004 with large amplitudes after the mid-1970s (not shown). For this reason, if one computes EOF analysis of the covariance matrix (i.e., U200 and U850 separately or together), PC1 and PC2 also have linear trends. This nonstationarity in PC1 and PC2 implies that few events would be identified in the first part of the record (e.g., for 1σ cutoff value). Thus, CEOF analysis of the correlation matrix of (U200 and U850) becomes an essential step so that PC1 and PC2 are nearly stationary and MJO events can be identified throughout the entire record. Because the correlation matrix normalizes the time series, PC_{index} of CEOF of (U200 and U850) has no physical units and, more importantly, is also nearly stationary (Fig. 4, bottom). The drawback of using the correlation matrix in this case is that PC_{index} of (U200 and U850) loses information about potential linear trends in MJO activity.

To circumvent this difficulty, we computed the spatial variance of U200 and U850 anomalies in the domain 15°S – 15°N , 30°E – 150°W during each MJO event (i.e., the EOF domain). These two other indexes [hereafter, $S^2(\text{U200})$ and $S^2(\text{U850})$] are shown in Fig. 5, and

FIG. 4. Amplitudes of MJO events defined as $PC_{index} = (PC1^2 + PC2^2)^{1/2}$, where PC1 and PC2 are the first two principal components. (top) Amplitudes from EOF analysis of OLR anomalies ($W m^{-2}$); (bottom) amplitudes from combined EOF analysis of U200 and U850 anomalies (dimensionless). Each vertical bar represents an event and is registered at the peak in PC1.

the vertical bars are registered at the corresponding pentads of peaks in PC1. We now observe a clear positive linear trend in $S^2(U200)$ with particularly high amplitudes in the 1970s, 1980s, and 1990s. While $S^2(U850)$

does not appear to have large linear trends, we note that the plot shows winter and summer amplitudes together. As discussed in the next section, $S^2(U850)$ also exhibits positive linear trends in both seasons.

FIG. 5. Spatial variance of (top) U200 and (bottom) U850 anomalies. Variances are computed over $15^{\circ}S$ – $15^{\circ}N$, $30^{\circ}E$ – $150^{\circ}W$ during each MJO event shown in Fig. 4 and registered at the peak in PC1. Units: $(m s^{-1})^2$.

To conclude this section, we discuss some additional important points regarding the methodology to identify MJO events. First, if one applies the same exact method (e.g., EOF of correlation matrix) on U200 and U850 anomalies separately, PC1 from EOF of U200 anomalies has a different frequency distribution than PC1 from EOF of U850 anomalies (and likewise PC2). This implies that for any given cutoff value (e.g., 1σ), different numbers of MJO events are identified in both methods, even though they are based on circulation metrics. The CEOF (U200 and U850) method used in this study identifies MJO events based on coherent variations in U200 and U850 anomalies.

Second, in a previous study, Kessler (2001) discussed the importance of the third EOF in representing the MJO. In that study, intraseasonal OLR anomalies were averaged around the equator (5°S – 5°N) and EOF analysis was computed on time series of the entire record (1979–99); that is, winter and summer periods were not separately analyzed. In the Kessler (2001) analysis, the third EOF was well above the North et al. (1982) rejection criterion by a factor of ~ 3 . Kessler (2001) pointed out that the third EOF has a maximum in the region 150°E – 170°W , and the phasing between the first and third EOFs is such that during warm ENSO events, EOF3 weakens EOF1 in the Indian Ocean and strengthens EOF1 in the Pacific, which better characterizes the propagation of MJO events farther east in the central Pacific (see his Figs. 2 and 4). Opposite arguments were proposed for cold ENSO episodes. Additionally, Kessler (2001) showed that rotating the first 10 EOFs improves the characterization of the EOF loadings in the equatorial Indian and Pacific Oceans.

In this study, EOF analysis of OLR anomalies shows that the third eigenvalue is separated from the fourth eigenvalue only in the summer analysis but not in winter (Table 1). The main reasons for the differences from Kessler (2001) results are the EOF domains (15°S – 15°N , 30°E – 150°W in this study versus average around the equator) and computation of winter and summer data separately. On the other hand, the CEOF analysis of U200 and U850 anomalies indicates different numbers of separated eigenvalues in summer (3) and winter (6). In contrast to Kessler's (2001) approach (i.e., average around the equator), we have found that the spatial patterns of the first three rotated eigenvectors of CEOF analysis in the domain 15°S – 15°N , 30°E – 150°W (not shown) were very sensitive to small changes in the total number of rotated eigenvectors (e.g., first 9, 10, and 11 EOFs). Last, in a previous version of this study, we employed CEOF of U200 and U850 anomalies averaged around the equator as in Kessler (2001) and rotated the first 10 EOFs. That methodology appears po-

tentially useful in characterizing zonal displacements in winter MJO activity during ENSO years. However, we found that limiting the analysis to CEOF of time series averaged around the equator significantly misses winter MJO events whose maximum amplitudes are off the equator.

In conclusion, this study uses only the first two EOFs of OLR, U200, and U850 anomalies to identify the MJO. The main reasons for this approach are that 1) the first two EOFs and associated PCs capture the bulk of the eastward propagation of the MJO and are relatively insensitive to reasonable choices of tropical domains used in the EOF calculation; 2) additional EOF modes (i.e., EOF3 and higher), and their eigenvalues separation, do not have physical meaning and are highly sensitive to the domain used in the EOF calculation; and 3) rotated EOFs are sensitive to the domain used in the EOF calculation as well as the selected criteria for EOF rotation.

4. Trends in MJO activity

Slingo et al. (1999) used an index to capture the essential perturbations resulting from intraseasonal variations in upper-level zonal winds in the equatorial region. The index is computed by taking the zonal mean of U200 (10°S – 10°N). That time series is bandpassed from 20 to 100 days, the anomalies are squared, and a running mean of 100 days is applied to emphasize seasonal and interannual variations. Figure 6 (top) reproduces the Slingo et al. (1999) index (hereafter ZM-U200). As previously discussed, the amplitudes during 1958–79 are smaller than in 1979–2004. To gain further insight into tropical intraseasonal variability during 1958–2004, we have computed the same index but using U850 (Fig. 6, bottom). To avoid topographic influences, the zonal average was performed over 10°S – 10°N , 30°E – 100°W . Interestingly, a different behavior is observed in the U850 index (hereafter ZM-U850) and shows high amplitudes from 1958 to about 1980 and decreases to a low regime in 1989–2000. The angular coefficients of the linear trends are $B_{\text{U200}} = 1.4 \times 10^{-4}$ and $B_{\text{U850}} = -1.6 \times 10^{-5}$.

In terms of MJO activity, an important point has been made by Hendon et al. (1999), who investigated the interannual behavior of the MJO. They examined different measures of MJO activity and pointed out that the zonal mean of zonal winds at 200 hPa exhibits significant intraseasonal spectral peaks, but the MJO accounts for only about 60% of the intraseasonal variance above the red noise spectrum. This indicates that the ZM-U200 index contains a significant portion of unrelated MJO variability. Thus, the interpretation for that

FIG. 6. Equatorial zonal mean intraseasonal index from (top) U200 and (bottom) U850. Note that the ZM-U200 index was computed over 10°S – 10°N (all longitudes) and ZM-U850 over 10°S – 10°N , 30°E – 100°W . Dashed lines are linear fits. Units: $(\text{m s}^{-1})^2$.

index is that it measures the level of equatorial intraseasonal variability in U200 and indirectly captures some MJO activity.

To compare both metrics of MJO activity, we removed the dates of MJO events (1958–2004) identified with the CEOF analysis (e.g., Fig. 4, bottom) from the ZM-U200 index. For each event, we removed the amplitudes of ZM-U200 from -3 to $+4$ pentads around the peak in PC1. This accounts for 42.9% of the entire data record ($8 \text{ lags} \times 184 \text{ MJO events}/3431 \text{ pentads}$). Next, we recomputed the linear trend in the ZM-U200 index. By removing MJO events, the linear trend in the ZM-U200 index decreased by $\sim 74\%$, which indicates that the trend is mostly associated with MJO activity. In contrast, the ZM-U850 index exhibits more complex changes than a linear trend.

Possible trends in MJO activity were investigated with events identified with the methodology explained in section 3. Figure 7 displays $S^2(\text{U200})$ and $S^2(\text{U850})$ amplitudes of all MJO events separated by winter and summer seasons. Also indicated are the equations of temporal linear trends. Positive linear trends are observed in both variables and seasons, with trends in $S^2(\text{U200})$ being larger than in $S^2(\text{U850})$.

An important issue regarding trends in MJO amplitudes is whether the observed trends are higher than randomly sampled occurrences. To test this hypothesis, we randomized the $S^2(\text{U200})$ winter amplitudes (Fig.

7a) N times and linear trends fitted to each batch. The magnitude of the actual angular coefficient (1.134×10^{-4}) was compared to the 95th percentile of the frequency distribution of the N angular coefficients of data randomizations. The total number of resamplings (N) was decided based on the numerical convergence of the 95th percentile. Specifically, we resampled the amplitudes from 100 up to 2000 resamplings with steps equal to 50. The 95th percentile converged for $N \approx 1500$ times. The process was repeated for the other amplitudes shown in Fig. 7. Based on this statistical test, only positive linear trends in $S^2(\text{U200})$ and $S^2(\text{U850})$ amplitudes during summer (Figs. 7c,d) are higher than trends from random occurrences at the 5% significance level.

Figure 8 shows the distributions of the number of MJO events during winter (top) and summer (middle) as well as the total number of events per calendar year (bottom). To interpret the plot, it is important to remember that the year is divided into extended winter (2 November–26 April) and summer (1 May–28 October) seasons. For instance, the MJO was very active during the warm ENSO event in 1997 (McPhaden 1999). The CEOF method identified four events in 1997 (bottom) so that the peak PC1 amplitudes were registered on 10 February, 6 April, 11 May, and 22 December, and the events were assigned as winter (1996), winter (1996), summer (1997), and winter (1997), respectively. Note that, although some winter and summer seasons did not

FIG. 7. Temporal distribution of MJO amplitudes during (a), (b) winter and (c), (d) summer. Each vertical bar represents spatial variance of U200 and U850 anomalies (15°S – 15°N , 30°E – 150°W) during an MJO event. Equations of temporal trends are indicated on the top right of each plot. Units: $(\text{m s}^{-1})^2$.

register events, MJO events were actually observed in each year during 1958–2004 (bottom). Winter events show a small positive linear trend. In contrast, a clear positive linear trend is seen in the number of summer events. The statistical significance of the linear trends was assessed by randomizing the data points as previ-

ously explained, and the numerical convergence of the 95th percentiles was achieved for resampling sizes of $N = 700$. The positive linear trends in the number of summer and winter MJO events are not higher than random resamplings at the 5% significance level. In summary, the results presented here indicate that MJO

FIG. 8. Temporal distribution of number of MJO events during (top) winter and (middle) summer. Events are registered on extended winter (2 November–26 April) and summer (1 May–28 October) seasons. (bottom) Temporal distribution of number of MJO events during calendar years. Dashed lines indicate linear fits.

activity during summer shows statistically significant increases in $S^2(U200)$ and $S^2(U850)$ amplitudes during 1958–2004.

5. Changes in regimes of MJO activity

To summarize the seasonal activity of the MJO, one can construct an index combining the peak amplitudes and number of events:

$$MJO_{act} = \overline{PC_{index}} \times N_S. \quad (1)$$

For each season, $\overline{PC_{index}}$ is the mean PC_{index} (Fig. 4, bottom) multiplied by the number of events in the season (N_S). The activity index is displayed in Fig. 9, and one can readily see the highly variable nature in MJO_{act}

FIG. 9. MJO seasonal activity index (dimensionless).

with no apparent organized temporal changes in activity. This is not entirely surprising given that the amplitudes of MJO events vary significantly in time (Fig. 4, bottom) and dominate the activity index.

In spite of the amplitudes of the events, however, a close examination of the occurrences of MJO events (Fig. 4, bottom) reveals that the oscillation was very active during some consecutive years and subsequently separated by periods of low activity. This raises some important questions regarding the behavior of the MJO on time scales longer than interannual. Are there regime changes in the activity of the MJO? Are there significant seasonal differences in MJO regimes? Does the MJO have a low-frequency (LF) mode of variability?

To gain further insight into this problem, we characterized the activity of the MJO on different time scales in the following way.

- 1) Given the MJO occurrences (Fig. 4, bottom), we created a time series X_T , $T = 1, 3431$ pentads (1958–2004), such that $X_T = 1$ ($X_T = 0$) when there was an event (no event). This step simply discarded the amplitudes and assigned arbitrary “yes” and “no” values.

FIG. 10. Low-frequency diagram of the percentage of total number of MJO events occurring in moving scales S_K . Vertical axis represents scale S_K from 1 to 9.98 yr ($S_K = 73, 729$ pentads). MJO events were identified with (top) CEOF analysis of (U200 and U850) anomalies and (bottom) EOF analysis of OLR anomalies.

- 2) We defined a moving window S_K of size K pentads and counted the number of MJO events $M_{K,T}$, where K and T are scale and pentad (1 to 3431). The percentage of events in S_K is then $P_{K,T} = 100 \times M_{K,T}/N$, where N is the total number of MJO events (184). Similarly, the percentages of summer and winter events in S_K are $PS_{K,T} = 100 \times MS_{K,T}/N_S$ and $PW_{K,T} = 100 \times MW_{K,T}/N_W$, where $MS_{K,T}$ and $MW_{K,T}$ are the numbers of summer and winter events in S_K , and N_S and N_W are the total numbers of summer (85) and winter (99) events, respectively. The S_K scales are odd numbers and varied from the shortest (1 pentad) to the longest (3431 pentads). Likewise, steps 1 and 2 were repeated for MJO occurrences based on OLR anomalies (Fig. 4, top). The calculation described above follows the ideas used in wavelet analysis, except that in this case we computed percentages of events on different time scales. Note that wavelet analysis is not the appropriate tool here because the parameter in focus is the number of MJO events and not their amplitudes (or spectral variance), which appear to behave as a white noise process (Fig. 4).

Figure 10 (top) shows the distribution of $P_{K,T}$ during 1958–2004 and scales from 1 to 10 yr to facilitate the display. Note, however, that the resolution in the vertical axis (S_K) is in pentads and edge effects indicated by “cones of influence.” A similar diagram obtained from OLR anomalies (1979–2004) is shown in Fig. 10 (bottom). To help understand this LF diagram, one can imagine that if MJO events were evenly distributed in time, the contours would appear as horizontal lines with increasing values from short to long time scales. Furthermore, for the longest scale ($S_{3431} = 47$ yr), $P_{3431,T}$ would be equal to 100% and centered in the midpoint of the data record (30 June 1981). In contrast, the observed LF diagram of MJO variability shows a rather different behavior. Considering scales from 3 to 10 yr (Fig. 10, top), there is evidence of two regimes of high MJO activity in about 1966–78 and 1988–96, respectively, and a low regime in 1980–87. It is also interesting to observe that the steady increase in MJO activity from the early 1960s to 1970–74 and subsequent decrease in 1980–87 do not correlate with the introduction of satellite-derived winds in the NCEP–NCAR reanalysis, which can be indicative of real changes in MJO

FIG. 11. Same as in Fig. 10, but for percentage of total number of winter MJO events occurring in moving scales S_K .

regimes. A comparison with the LF diagram obtained from OLR anomalies (Fig. 10, bottom) does not indicate a close match with LF derived from (U200 and U850) anomalies (Fig. 10, top), although there is some consistency in the increased activity during 1984–96 and scales of 8–10 yr. Keeping in mind that there are seasonal changes in the MJO, Fig. 10 provides a collective view of the activity of the oscillation throughout the year.

Figure 11 shows LF diagrams ($PW_{K,T}$) of winter MJO activity obtained from (U200 and U850) (top) and OLR (bottom) anomalies. For nearly all time scales, winter activity is approximately uniform from 1963 to about 1994. However, a decrease in winter activity is observed from 1996 to 2000 and is consistent between circulation (top) and convective (bottom) metrics of MJO activity.

Summer LF changes ($PS_{K,T}$) derived from (U200 and U850) anomalies (Fig. 12, top) indicate large changes in MJO regimes on scales of 3–10 yr with high activity in about 1966–78 and 1988–2000 and separated by low activity in 1980–87. Although there is not a perfect agreement, summer LF changes obtained from OLR anomalies (Fig. 12, bottom) support an increase in MJO

activity from 1988 to 2000. The results above suggest that LF changes in summer MJO activity (Fig. 12) are different than in winter (Fig. 11), and they collectively contributed to the overall MJO behavior displayed in Fig. 10. The changes during 1958–2004 do not bear any obvious resemblance with a linear increase in observational sampling due to assimilation of satellite-derived winds in the NCEP–NCAR reanalysis. Unfortunately, the unavailability of an independent set of observations precludes further verification, and one cannot rule out the possibility that changes in regimes before 1979 are spurious.

To summarize the overall winter and summer LF changes in MJO activity (Figs. 10–12), we averaged $P_{K,T}$, $PW_{K,T}$, and $PS_{K,T}$ on scales S_K from 145 to 729 pentads (i.e., 1.98 to 9.98 yr). Note that near the edges, $P_{K,T}$, $PW_{K,T}$, and $PS_{K,T}$ have zero values inside the cones of influence, which were omitted from the averaging procedure. Figure 13 shows the mean $P_{K,T}$, $PW_{K,T}$, and $PS_{K,T}$ (hereafter P_M , PW_M , and PS_M). The overall activity of the MJO (top), P_M , points to a steady increase from early 1960s to early 1970s and a steady decrease in activity up to the mid-1980s, followed by increases until the late 1990s. The mean winter LF ac-

FIG. 12. Same as in Fig. 11, but for percentage of total number of summer MJO events occurring in moving scales S_K .

tivity (middle), PW_M , is characterized by nearly uniform activity until the mid-1990s and followed by a decrease in activity. On the other hand, the mean summer LF activity (bottom), PS_M , indicates a rapid increase in activity from the mid-1960s until the late 1970s with a steady decrease to a regime of low activity from about 1980 to 1988. Subsequently, the mean summer LF variability increased to a regime of high activity in the 1990s and early 2000. The overall amplitude of mean summer LF changes in MJO activity is $A_{SUM} = 9.2\%$ (i.e., $\sim 7\%$ in 1965 to $\sim 16\%$ in 1977).

It is interesting that the first regime of high activity in the mean summer PS_M (Fig. 13, bottom) occurred within the major climate shift in the basic state in the North Pacific Ocean and decadal ENSO variations in the tropical Pacific in the mid-1970s (e.g., Deser et al. 2004; Graham 1994; Hare and Mantua 2000; Trenberth and Hurrell 1994; Zhang et al. 1998; Yeh and Kirtman 2005), although the observed increase happened in the early to mid-1970s. Goswami (2005) investigated interdecadal changes in intraseasonal variability in the Asian monsoon with daily NCEP–NCAR reanalysis (1948–2002) and pointed out that interdecadal variability of ISO activity over the Indian summer monsoon

FIG. 13. Mean percentages of (top) total, (middle) winter, and (bottom) summer MJO events on scales S_K from 145 to 729 pentads (i.e., 1.98 to 9.98 yr).

FIG. 14. Amplitudes of first 10 harmonics from Fourier analysis of mean percentages of (top) winter and (bottom) summer MJO events. Numbers on each bar represent percentages explained by the harmonics.

region seems associated with interdecadal variability of the ENSO–monsoon mode. Goswami (2005) analysis indicates that the periods 1955–75 and 1980–95 were characterized by high and low ISO activity, and the transition between them coincided with the interdecadal shift in the mid-1970s. Furthermore, Zveryaev (2002) used U850 (NCEP–NCAR) over the Asian monsoon to characterize decadal–interdecadal variability in ISO activity and found strong linkages between interdecadal changes in SST over the Indian Ocean and intensities of ISO events over the Asian monsoon.

Fourier analysis was performed on the mean winter (PW_M) and summer (PS_M) LF MJO activity, and the results are displayed in Fig. 14. In the winter season (top), PW_M exhibits a dominant harmonic of amplitude 1.09 and 18.5-yr period, which explains 49.3% of the total variance. Likewise, the mean summer PS_M (bottom) has a dominant harmonic of amplitude 3.36 and period of 18.5 yr, which explains 77.1% of the total variance. The dominant harmonics are also shown in Fig. 13. Evidently, given the data record of 47 yr, these results do not allow inferring a periodicity of 18.5 yr but instead are used to characterize the regimes of high and low MJO activity on LF scales.

An important question regarding LF changes in winter and summer MJO activity is the statistical significance of those changes. The relevant question is whether randomly sampled time series could generate low-frequency variability sharing similar amplitudes of the

FIG. 15. Frequency distribution of amplitudes (R_{SUM}) of random mean LF summer MJO occurrences. Number of resamplings: 10 000 randomizations.

observed mean summer PS_M ($A_{SUM} = 9.2\%$) regardless of the exact timing of regimes of high and low activity. Therefore, Monte Carlo experiments were done as follows. Given the observed occurrences of summer MJO events (Figs. 7c,d), we discarded the amplitudes of the events and created a time series X_T , such that $X_T = 1$ when there was an event ($X_T = 0$ for no event) ($T = 1, 3431$ pentads). Next, we randomly shuffled summer seasons in X_T N times. For each resampling, we computed the percentage of summer events $PS_{K,T}$ in each scale S_K ($K = 1, 3431$, odd, $T = 1, 3431$ pentads). This step is equivalent to computing a plot similar to Fig. 12 (top) but based on random summer MJO occurrences. Based on this summer LF diagram, we calculated the mean $PS_{K,T}$ on scales S_K from 145 to 729 pentads (i.e., 1.98 to 9.98 yr). Finally, the amplitude of the mean LF summer curve was determined as the difference between the highest and lowest values (hereafter R_{SUM} ; 1958–2004).

A significant issue is that there is no prior knowledge on the size N of resamplings necessary to derive stable and reliable statistics. Tests were made by varying the size N from only 50 resamplings up to 10 000 randomizations and analyzing the frequency distribution of R_{SUM} . Figure 15 displays the frequency distribution of R_{SUM} obtained with 10 000 randomizations of summer occurrences, which was determined to be sufficiently large to provide stable parameters as shown in Table 2. Based on the Monte Carlo experiment described above, one can formulate the statistical test in which the null hypothesis is that the observed mean summer LF amplitude ($A_{SUM} = 9.2\%$) is not different from similar amplitudes of random MJO occurrences. Following this approach, one observes that A_{SUM} is located on the

TABLE 2. Distribution parameters of the frequency distribution of R_{SUM} shown in Fig. 15.

Random mean LF variability occurrences						
Mean	Median	Min	Max	Std dev	Skewness	Kurtosis
6.98	6.92	2.13	13.52	1.66	0.23	-0.07
Percentiles						
Lower quartile		Upper quartile	10th	90th	95th	
5.82		8.06	4.87	9.17	9.87	

upper tail of the frequency distribution of R_{SUM} , although its magnitude is only greater than the 90th percentile (Table 2). Thus, the null hypothesis can only be rejected at the 10% confidence level but not at the 5% level. Last, we note that the statistical test described above compares the observed amplitude A_{SUM} with R_{SUM} samplings without making any assumptions about time scales of changes. In other words, randomly sampled time series of summer MJO occurrences may have significantly large amplitudes but time scales different than the observed change of about 18.5 yr.

6. Summary and conclusions

The mechanisms involved in the initiation of the MJO, its maintenance, and the causes of its temporal irregularity continue to present a major challenge to develop a comprehensive theory of the oscillation. In this context, many previous studies have provided significant progress to understand some of the characteristics of the oscillation on seasonal to interannual time scales. In contrast, the behavior of the MJO on time scales longer than interannual has been unknown, largely due to a lack of long data records that can resolve intraseasonal variations. Currently, reanalysis products are the only data providing many years of consistently derived observations that can be used to infer long-term changes in the MJO. Motivated by this reason, this study used NCEP–NCAR reanalysis to investigate the activity of the MJO during 1958–2004.

First, this paper investigated if there is statistical evidence of linear increases in the intensity and number of MJO events during 1958–2004. This aspect was motivated by the study of Slingo et al. (1999), who used the ZM–U200 index to examine interannual predictability of the MJO. The ZM–U200 index exhibits more intense amplitudes since the mid-1970s. Although we cannot rule out the possibility that the trends are due to increased observational samplings assimilated in the reanalysis, general circulation model experiments forced with observed SSTs offer additional evidence that the trends can be real (Slingo et al. 1999). While the ZM–U200 index provides a concise and useful metric

for MJO activity, the index itself does not identify individual MJO events. This study used CEOF analysis of U200 and U850 anomalies to identify occurrences of the MJO. Both metrics of the MJO, that is, CEOF and ZM–U200 index, are consistent with each other as far as linear trends in U200 amplitudes.

Positive linear trends are observed in the variance of U200 and U850 anomalies in the region 15°S–15°N, 30°E–150°W during summer and winter seasons, while summer trends are higher than winter trends. Positive trends are also observed in the number of summer and winter MJO events. Resampling statistical tests indicate that positive trends in U200 and U850 anomalies in summer are statistically different from random occurrences at the 5% significance level.

Second, based on the number of events during 1958–2004, this study developed a methodology to characterize shifts in the activity of the MJO with emphasis on time scales longer than interannual. Considering events throughout the entire year, the MJO exhibited a steady increase from the early 1960s to early 1970s and a steady decrease in activity up to the mid-1980s, followed by increases until the late 1990s. The mean low-frequency activity in winter was characterized by nearly uniform activity from the early 1960s until the mid-1990s and was followed by a decrease in activity. On the other hand, mean low-frequency changes in the summer indicated large and rapid increases in activity from the mid-1960s until the late 1970s with steady decreases to a regime of low activity from about 1980 to 1988. Subsequently, the mean summer low-frequency variability increased to a regime of high activity in the 1990s and early 2000. Fourier analysis of the mean summer low-frequency MJO activity suggests that changes between the high and low regimes of activity were separated by 18.5 yr. The substantial changes in summer MJO regimes do not appear to be related to increases in observational samplings because of satellite-derived winds assimilated in the NCEP–NCAR reanalysis. Monte Carlo experiments, however, indicate that the observed changes in summer regimes of MJO activity are not statistically different than random occurrences at the 5% significance level.

The results from this study raise some important questions that deserve future investigation. Since summer and winter MJO activity involves significant feedbacks between convection and large-scale circulation, one can expect that low-frequency changes in the basic state will affect the activity of the MJO as well. What are the dynamical mechanisms forcing regime changes in the MJO? Are regime changes in the MJO related to changes in other low-frequency modes of the coupled ocean–atmosphere system (e.g., ENSO and Pacific decadal oscillation)?

This study was based on 47 yr of data and, ideally, a much longer data record would be necessary to reach definitive conclusions on the behavior of the MJO on long time scales. Unlike other phenomena (e.g., ENSO), proxy records do not provide enough temporal resolution to resolve intraseasonal variability and, therefore, low-frequency changes in the MJO. This indicates that coupled ocean–atmosphere simulations with numerical models that realistically represent the MJO might be able to provide further insight on this research aspect. Given the importance of the MJO in modulating weather variability in the Tropics and extratropics of both hemispheres, knowledge of the long-term behavior of the MJO is one of the many critical steps to be achieved before reliable projections about future global climate changes can be achieved.

Acknowledgments. This research was funded by the NOAA Office of Global Programs CLIVAR Pacific Program (NOAA/NAO30AR4310067 and NOAA/NAO5OAR4311129) and the National Science Foundation (ATM-0094387). Leila M. V. Carvalho would like to acknowledge the support from CNPq, Brazil, (proc: 302203/02-8). NCEP reanalysis and OLR data were provided by the NOAA–CIRES ESRL/PSD Climate Diagnostics Branch, Boulder, Colorado, from their Web site (<http://www.cdc.noaa.gov>). Comments and suggestions from the anonymous reviewers greatly improved this study. The authors are particularly grateful to Duane Waliser for stimulating and helpful discussions and Johnny Lin for providing his IDL varimax computer code.

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