Seasonality in the Madden–Julian Oscillation

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(Manuscript received 15 July 2003, in final form 17 February 2004)

ABSTRACT

This study, using observational data and global reanalysis products, synthesizes existing knowledge on the seasonality in the Madden–Julian oscillation (MJO) and discusses its possible dependence on the mean background state. The seasonality in the MJO is documented in terms of its components in low-level zonal wind and precipitation. A single peak season (boreal winter) exists near the equator in the Indian and western Pacific Oceans. For the broad tropical region, however, the seasonality in the MJO is featured by a latitudinal migration across the equator between two peak seasons. The primary peak season is in boreal winter (December–March), during which MJO signals are mainly confined to the Indian and western Pacific Ocean and reach their maxima in the South Pacific convergence zone. The secondary peak season is boreal summer (June–September), during which the strongest MJO occurs north of the equator from the Bay of Bengal to the South China Sea and another, separated, region of strong MJO signals is located in the eastern Pacific warm pool off the Central American coast. The seasonal cross-equatorial migration is the strongest in the western Pacific in both wind and precipitation and much weaker in the Indian Ocean in precipitation. Strong seasonal activities of the MJO appear to prefer mean westerlies or weak zonal winds at the surface and low level (850 hPa) and mean lower-level moisture convergence in the western Pacific. The relationship between the MJO and its mean background state is much less clear in the Indian Ocean. Implications of these observations are discussed.

1. Introduction

Seasonality is one of the most fundamental characteristics of the Madden–Julian oscillation (MJO; Madden and Julian 1971, 1994). Near the equator, the strongest MJO signals are observed in boreal winter and spring; the weakest signals are in boreal summer (Madden 1986; Gutzler and Madden 1989). In a broader tropical region as a whole, MJO signals in the low-level zonal wind migrate in latitude, with its peak activity located south of the equator in boreal winter and north of the equator in summer (Salby and Hendon 1994). The onset and breaks of the Australian summer monsoon are closely related to the phase of the MJO during boreal winter (Hendon and Liebmann 1990). The western Pacific MJO signals in boreal summer are partially related to the intraseasonal variations in the Asian summer monsoon (e.g., Yasunari 1979; Lau and Chan 1986; Lawrence and Webster 2002). MJO signals can also be found in the eastern Pacific over the warm sea surface off the Central American cost in boreal summer (Knutson and Weickmann 1987; Maloney and Kiehl 2002).

Our current knowledge on the seasonality in the MJO has been obtained fragmentally from various studies. These studies focused on particular regions and seasons and used different data and methods to detect the MJO and define its seasonality. Many uncertainties remain in these studies. For example, Gutzler and Madden (1989) concluded from rawinsonde observations of zonal winds that the strong seasonal cycle of the MJO exists mainly west of the date line. But the MJO signals in the eastern Pacific (Maloney and Kiehl 2002) appear to be much stronger during boreal summer than winter (Knutson and Weickmann 1987). Knutson and Weickmann (1987) showed that maximum signals of the MJO in convection as depicted by outgoing longwave radiation (OLR) always reside on the summer hemispheric side of the equator, implying a seasonal migration in latitude. Such a seasonal migration was indeed found by Salby and Hendon (1994) but only in zonally averaged low-level zonal wind. Hartmann and Gross (1988), however, failed to detect intraseasonal signals from rawinsonde observations of zonal winds in India during the summer monsoon season. The mechanism for the seasonal variation in the MJO remains unknown. It is convenient to assume that sea surface temperature (SST) is responsible for the seasonal migration. The relative roles of SST and the large-scale circulation in the seasonality in the MJO have yet to be explored.

The objectives of this present study are 1) to synthesize the existing knowledge on the seasonality in the
MJO by applying a single approach to the same data and by treating MJO signals equally in the entire Tropics and in all seasons and 2) to explore the extent to which the seasonal variations in the MJO and its mean background state are related. The approach adapted takes account of all unique and basic features of the MJO: its temporal and spatial scales, eastward propagation, and coupling between the zonal wind and precipitation. A special method is applied to explicitly reveal the longitudinal dependence of the seasonality. The mean background state of the MJO is represented in terms of SST, surface and low-level zonal winds, and low-level moisture convergence. Results from this study would provide an observational benchmark, as a higher standard than simple spectrum analyses, for validating simulations of the MJO by models that are capable of reproducing eastward propagating intraseasonal signals.

The data and analysis methods are described in section 2. Results are presented in section 3. Discussions are given in section 4.

2. Methodology

a. Data

Observational data used in this study include SST (Reynold and Smith 1994), precipitation (Xie and Arkin 1997), and surface winds observed from the European Remote Sensing (ERS) satellites (Bentamy et al. 1999). The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) is used as a surrogate of observations in low-level (925 and 850 hPa) winds and humidity. We do not use high-level (e.g., 200 hPa) winds because we believe the key to understanding the MJO, including its seasonality, is to understand its dynamics and thermodynamics in the lower troposphere and in the atmospheric boundary layer where its deep convective component is rooted. Signals of the MJO in upper-tropospheric winds, in our opinion, reflect more responses of the atmosphere to MJO convection than mechanisms for the MJO. The data analysis period covers 20 years from 1979 through 1998. The data are averaged into zonal and meridional resolutions of 10° and 5°, respectively, and averaged or interpolated into the pentad time series before they are analyzed to extract MJO signals.

b. Method

In this study, the MJO is defined as the planetary-scale (zonal wavenumber 1–5), intraseasonal (period of 20–90 days), and eastward propagating components in low-level (850 hPa) zonal wind coupled with precipitation in the Tropics (20°N–20°S). This is a rather restrictive definition of the MJO. It does not include the part of the MJO existing in the wind field outside its convective region in the tropical Indian and Pacific Oceans (Gutzler and Madden 1989). This definition distinguishes the MJO from the general concept of the tropical intraseasonal variations (TIV), which includes everything in the intraseasonal frequency band. For example, a prominent phenomenon of the TIV during the Asian summer monsoon season is its northward propagation (e.g., Yasunari 1979; Lau and Chan 1986; Wang and Rui 1990a). This is not considered as the MJO in this study, even though it was discussed by Madden and Julian (1971, 1972, 1994). Only the eastward-propagating intraseasonal signals during the Asian summer monsoon season are considered as of the MJO. Because the northward and eastward propagating components of the TIV related to the summer monsoon are not at all dynamically separated (Lawrence and Webster 2002), isolating the eastward propagating component and ignoring the northward propagating counterpart of the summertime TIV might be considered artificial. The separation is, however, necessary to emphasize the seasonality in the mechanism for the eastward propagation of the TIV, which is, in our term, the MJO.

The temporal and spatial scales of the MJO are selected using bandpass filtering in time (20–90 days) and low-pass filtering in longitude (wavenumber 1–5). The coupling between wind and precipitation is established using a singular vector decomposition (SVD) method (Wallace et al. 1992) applied to the 850-hPa zonal wind ($U_{850}$) and precipitation ($P$). The most commonly used method of detecting the MJO is the EOF analysis. The first pair of leading EOF modes of an intraseasonally bandpassed time series are usually taken to represent the MJO, provided they are well separated from other modes and they are in quadrature with each other in time and longitude (e.g., Knutson and Weickmann 1987; Lau and Chan 1988; Zhang and Hendon 1997). The SVD method works similarly but with one advantage: it takes the wind–precipitation coupling of the MJO into consideration. This matters, especially when the dynamical aspect of the MJO is the main target of the analysis, as in the present study. The leading SVD modes, defined as those that are separated from the other modes according to the rule of North et al. (1982), are selected to represent the intraseasonal components that are coherent between the wind and precipitation on large scales. Time series of the two fields reconstructed (through linear regression) using the leading SVD modes are regarded as the representation for the MJO, which are hereafter denoted as $U_{850}$ and $P$. Here, we include four, instead of two, leading modes for the following reason: While the leading pair of SVD modes can usually isolate eastward propagating signals, they alone do not fully depict the seasonal migration in latitude of the MJO. This seasonal migration of the MJO can be well described only when higher modes are included, as long as they are separated from the others. In this approach, the MJO is represented by the totality of the group of the four SVD leading modes. In the group, no single or pair of SVD modes alone would be
considered to bear any specific physical meaning as far as the MJO is concerned.

The eastward propagation of $U^*_{850}$ and $P^*$ is verified using a time–space power spectrum method (Hayashi 1979). An unambiguous dominance of their eastward propagating power over their westward propagating power at the same frequencies and zonal wavenumbers is the testament of their eastward propagation. It needs to point out that isolating all eastward propagating signals in an intraseasonally bandpassed time series is not sufficient to correctly represent the MJO, unless the westward propagating power is negligibly weak. Some of the eastward propagating signals, if coherent with their westward propagating counterparts (at the same frequencies and zonal wavenumbers), belong to stationary components (Hayashi 1979), which in our definition do not represent the MJO. When the dominance of eastward propagating power over westward propagating power is not obvious, the MJO should be considered too weak to be discerned. As will be shown, this does not happen in our analysis.

Once MJO signals, $U^*_{850}$ and $P^*$, are identified, a wavelet analysis in time is applied to produce wavelet spectral power at each grid point. This wavelet spectral power is thus a function of longitude, latitude, time, and frequency. The seasonality in the MJO is visualized either as the seasonal variation in this wavelet spectral power or in the total variance of $U^*_{850}$ and $P^*$.

3. Results

a. Seasonality

Following the procedures described in section 2b, four leading SVD modes are found well separated from the others. They explain 49% of the total covariance. The time–space spectra of $U^*_{850}$ and $P^*$, reconstructed using the four leading SVD modes, demonstrate beyond doubt the dominance of eastward propagation (Figs. 1a and 1b). The maximum eastward propagating power (at positive frequencies) is four times greater than the maximum westward propagating power for $U^*_{850}$ and three times greater for $P^*$. This confirms that the reconstructed time series indeed represent the MJO. The MJO thus extracted accounts for a large fraction of the total intraseasonal variance in regions known to host strong MJO activities: 60%–90% in the western Pacific, greater for $U^*_{850}$ than for $P$ (Fig. 2). But regions of large total intraseasonal variance may not necessarily show large fractional variance of the MJO (e.g., southern Indian Ocean in boreal summer for $U^*_{850}$), illustrating the necessity of distinguishing one from the other.

The seasonal contrast of the MJO between boreal winter (December–March) and summer (June–September) is also shown in Fig. 1. For both $U^*_{850}$ and $P^*$, the eastward propagating power in winter (Figs. 1c and 1d) is stronger and centered at frequencies closer to (60 days)$^{-1}$ than in summer (Figs. 1e and 1f), during which there is a slight tendency for the maximum power to shift toward higher frequencies. The maximum eastward propagating power for $P^*$ is centered at zonal wavenumber 2 in winter (Fig. 1d) but 1 in summer (Fig. 1f). Figure 3 shows wavelet spectral power averaged over 20°N–20°S as functions of season and longitude. Throughout the entire year, the MJO signals in $P^*$ (Fig. 3b) are much more confined in longitude than the signals in $U^*_{850}$ (Fig. 3a). The strongest MJO power is in the western Pacific (150°E–180°) in boreal winter (December–March). But, there is a second peak in $P^*$ in the Indian Ocean, which remains there most of the time. The separation between the two local peaks in winter may explain the maximum spectrum power of $P^*$ at zonal wavenumber 2 (Fig. 1d). In summer, the MJO maximum in the western Pacific moves westward to 120°–150°E and weakens slightly. Meanwhile, MJO activities in the eastern Pacific (240°–270°E) become obvious (Maloney and Kiehl 2002). But they are too weak to make the maximum eastward-propagating power move away from zonal wavenumber 1. It is interesting that the maxima in $U^*_{850}$ and $P^*$ are always collocated in the western Pacific, but not in the Indian Ocean. This will be further discussed in section 4.

Figure 3 clearly demonstrates that through the entire year main activities of the MJO, as represented by coupled wind and precipitation, are in the Indian and western Pacific Oceans. Focusing on this region, we present in Fig. 4 zonally (60°–180°E) averaged wavelet spectral power of the MJO (contours) as a function of month and latitude. The most striking feature in this figure is the latitudinal seasonal migration of the MJO across the equator between two peak seasons. The primary peak season is in boreal winter (December–March), the secondary one in summer (June–September). Interestingly, the strongest MJO signals are always found away from the equator. The MJO peaks south of the equator (6°S for $U^*_{850}$ and 10°S for $P^*$) in winter and north of the equator (10°N) in summer. This summertime MJO peak is partially associated with the eastward propagating component of the intraseasonal variation of the Asian summer monsoon (Lau and Chan 1986; Lawrence and Webster 2002). The winter spectral peak is 40% stronger than the summer peak for $U^*_{850}$ (Fig. 4a) and 60% stronger for $P^*$ (Fig. 4c). Only near the equator can one find a single peak season in winter. This equatorial seasonal cycle of the MJO with a single peak, commonly cited in the MJO literatures, does not reflect the totality of the seasonality in the MJO. In a spectral analysis, Salby and Hendon (1994) found the cross-equatorial migration between two peak seasons in their MJO signals of zonal winds. Here we confirm that such a seasonal migration also exists in the precipitation component of the MJO. We will discuss Fig. 4 in more detail later.

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1 In this latter case, isolating the eastward propagating signals becomes unnecessary.
The spatial distributions of the MJO during its two peak seasons are shown as contours in Figs. 5 and 6. The strongest MJO activities occur over the South Pacific convergence zone (SPCZ) in winter. In summer, they move northwestward to the South China Sea. While the regions of maximum MJO activity in the western Pacific in both winter and summer are away from the equator, the local maximum in $P^*$ in the Indian Ocean is always close to the equator (Fig. 6). The latitudinal migration is therefore greater in the western Pacific than in the Indian Ocean. In summer strong MJO activities in the eastern Pacific occur over the warm sea surface off the Central American coast (Maloney and Kiehl 2002), which has been referred to as part of the Western Hemisphere warm pool (Wang and Enfield 2003). They, however, do not appear to be directly related to the strongest MJO activities in the western Pacific. Between the two regions there is a gap where strong MJO signals are absent along the ITCZ. The lack of a continuous propagation in deep convection from the western to east-
ern Pacific in boreal summer is also evident in the MJO composite of Knutson and Weickmann (1987). As will be discussed in section 3b, these seemingly missing signals of the MJO in between the western and eastern Pacific may provide a hint on the role of mean background state in the MJO.

Explaining the observed seasonality in the MJO is not as trivial as it might appear. The equatorial Kelvin and Rossby waves, the dynamic backbones of the MJO (e.g., Wang and Rui 1990b), are both symmetric about the equator (Matsumo 1966). One, therefore, is attempted to expect the MJO to be the strongest in the season when it is the most symmetric about the equator (May and October). In reality, that is when the MJO is weakest (Fig. 4). On the other hand, the seasonal migration of the MJO in latitude can be considered as part of the general seasonal cycle in the western Pacific. Next, we seek possible connections between the observed seasonality in the MJO and its mean background state.


**b. Relation to the mean background state**

The main signatures in the seasonality in the MJO previously discussed are in a sense embedded in the seasonal cycle of its mean background state. In this section, seasonal variations in the MJO and its mean background state are compared to explore the degree to which they are related. The mean background state is represented by monthly means of SST, low-level (850 hPa) zonal wind ($U_{850}$), surface zonal wind ($U_s$), and lower-tropospheric (925 and 850 hPa) moisture convergence (calculated using monthly mean winds and humidity).

Seasonal variations in the mean background state averaged over the Indian and western Pacific (60°E−180°W) are overlaid with that in $U_{850}$ and $P^*$ in Fig. 4. Their seasonal mean spatial distributions in the two peak seasons of the MJO are plotted in Figs. 5−10. The seasonal migration of the MJO is in phase with that of SST but their amplitudes do not match (Fig. 4a). The MJO reaches its second peak season when tropical SST is on average relatively low. Spatially, the maximum MJO activity is located over the warmest sea surface in the western Pacific but not necessarily so the Indian Ocean (Figs. 5 and 6). The most intriguing feature is that the MJO is weakest in May near the equator when SST is on average the highest and both are the most equatorially symmetric (Fig. 4a). (The annual maximum in SST in May is perhaps due more to the weakening of the equatorial cold tongue in the central Pacific than warming in the Indian and western Pacific warm pool.) There is no reason to expect this if SST is important to deep convection and if the Kelvin wave is coupled with deep convection. Apparently, other factors must be in play.

The seasonality in the MJO and mean zonal winds is compared in Figs. 4b, 4d, 7, and 8. First, it is clear that the two peak seasons of the MJO are also seasons of maximum mean westerlies in the lower troposphere (Fig. 4b) and at the surface (Fig. 4d). The MJO and the mean zonal winds, however, do not match each other so well in their zonal distributions. MJO activities in both peak seasons are generally in regions of mean westerlies at 850 hPa (Fig. 7) and at the surface (Fig. 8) in the Pacific Ocean. But their maxima appear to prefer weak or zero mean zonal winds. In boreal summer, for example, the strongest mean westerly occurs in the Arabian Sea, apparently associated with the monsoon where the MJO is rather weak. Meanwhile, the MJO always avoids strong mean easterlies. In summer, strong MJO activities are found in the western and eastern Pacific north of the equator, both feature mean westerlies or weak mean zonal winds. In between, there is no obvious MJO signal at the ITCZ latitudes where SST remains high (Figs. 5b and 6b) but mean easterlies prevail (Figs. 7b and 8b).

Finally, we compare the seasonality in the MJO and mean moisture convergence at the 850-hPa and 925-hPa levels. Notice that the mean moisture convergence calculated here is not total moisture convergence but is the component contributed by monthly mean winds and humidity. At the 850-hPa level, there is very good collocation between regions of the most active MJO and largest mean moisture convergence in the western Pacific in both winter and summer (Fig. 9). No such collocation exists in the Indian Ocean or the eastern Pacific. At the 925-hPa level, the two are not related in any obvious way (Fig. 10). The possible relationship between the MJO and mean 850-hPa moisture convergence is better illustrated in Fig. 4e, where in both peak seasons of the MJO its strongest activities are well collocated in latitude with maximum moisture convergence. Meanwhile, the seasons of the weakest MJO (e.g., May) are marked with the weakest moist convergence. This good relationship comes mostly from the western Pacific (Fig. 9). Again, no convincing relationship can be found between the seasonality in the MJO and 925-hPa moisture convergence (Fig. 4f). The discrepancies between mean moisture convergence at the 850-hPa and 925-hPa levels are intriguing, which will be further discussed in section 4.

**4. Summary and discussion**

The seasonality in the MJO has been shown to have the following main features:

1) The MJO has a single peak season near the equator in the Indian and western Pacific Ocean (boreal win-
Fig. 4. (a), (b) Wavelet spectral power of $U'_e$ (white contours) and (c)–(f) $P'$ (white contours) averaged over 60°E–180° and each month for 1979–98. The background colors (with zero marked by black contours) are for mean (a) SST (°C), (b) $U_{850}$ (m s$^{-1}$), (c) $P$ (mm day$^{-1}$), (d) surface zonal wind $U_s$ (mm day$^{-1}$), (e) 850-hPa moisture convergence (g kg$^{-1}$ m$^{-1}$), and (f) 925-hPa moisture convergence (g kg$^{-1}$ m$^{-1}$).
ter/spring) and in the eastern Pacific (summer). In the broad tropical Indian and western Pacific Oceans, strongest MJO signals migrate in latitude between $5^\circ$–$10^\circ$S in the primary peak season in winter (December–March) and $5^\circ$–$10^\circ$N in the secondary peak season in summer (June–September). Such seasonal cross-equatorial migration is much stronger in the western Pacific than in the Indian Ocean.

2) Among the mean fields examined, the MJO seems to prefer mean westerlies or weak mean zonal winds at low level (850 hPa) and the surface but prefer low-level mean moisture convergence only in the western Pacific.

Result 1) synthesizes the existing knowledge on the seasonality in the MJO from previous studies. The method applied to identify the MJO extracts all its fundamental natures (e.g., eastward propagation, circulation–convection coupling) and treats the MJO equally in the entire Tropics and during all seasons. The results from
this study are mostly consistent with the previous studies but also clarify some remaining uncertainties. The seasonal latitudinal migration of the MJO across the equator, as found by Salby and Hendon (1994) for the entire Tropics is mainly in the western Pacific Ocean. The second peak season of the MJO in summer, that they found in the low-level zonal winds also exists in precipitation. In the sector of the Indian Ocean, strong MJO signals in the zonal wind remain over the ocean during the summer monsoon season and thus are absent from the rawinsonde observations over land (Hartmann and Gross 1988). The MJO signals in the eastern Pacific (Maloney and Kiehl 2002) are disconnected from those in the western Pacific. The intraseasonal signals there in boreal winter (Knutson and Weickmann 1987) are too weak to be discerned as of the MJO.

These results can serve as a higher-order validation against simulations of the MJO by global models that are capable of producing intraseasonal, eastward propagating signals (e.g., Sperber et al. 1997). One can assert
that unless the models get the seasonality in the MJO right, the validity of their mechanisms for simulated MJO remain in question. While simulating MJO signals in global models is already a challenging ordeal (Slingo et al. 1996; Wang and Schlesinger 1999), asking for the correct seasonality in the MJO from models might be considered too demanding. We think not. Only when a model reproduces well the fundamental features of the MJO, including its seasonality, can it be trusted as a tool to study the dynamics of the MJO.

Recent studies have suggested the importance of the mean background state to the MJO (e.g., Hendon 2000; Kemball-Cook et al. 2002; Inness et al. 2003). We have examined the extent to which this is applicable to the seasonality in the MJO. The preference of the strongest MJO activities at their peak seasons to westerly or weak
mean zonal wind at the surface does not come as a total surprise. Only in regions of mean westerly or weak surface zonal wind can surface evaporation be enhanced by the surface westerly component of the MJO, often found in its large-scale convective center (e.g., Zhang and McPadden 2000). Such enhanced evaporation can serve to supply additional moisture for deep convection.

The similarity between the seasonality in the MJO and its mean background in moisture convergence at 850 hPa is intriguing (Figs. 4e and 9). One would probably guess this similarity existing between the MJO and mean moisture convergence in the boundary layer (e.g., at 925 hPa). It is doubtful that mean 850-hPa convergence would be a main moisture supply to the MJO. But it may help lessen the suffocating effect of dry air on deep convection, which is common in the western Pacific (e.g., Mapes and Zuidema 1996; Brown and Zhang 1997; Yoneyama and Parsons 1999). It is also possible, on the other hand, that the observed similarity is an atmospheric response to the MJO. The regions of the most active MJO are also regions of large mean precipitation (not shown), presumably from deep convection, which induces large-scale convergence most efficiently in the lower troposphere above the boundary layer (Schneider and Lindzen 1977; Wu 2003).

The lack of moisture convergence at 850 hPa near the eastern Pacific and Atlantic ITCZ (Fig. 9b) should not be left unexplained. Convection there is in general shallower than in other regions of the Tropics (e.g., Nesbitt et al 2000), which leads to a meridional circulation shallower than the classical deep Hadley-type circulation (Tomas and Webster 1997; Trenberth et al. 2000; Zhang et al. 2004). This shallow meridional circulation gives rise to divergence atop the boundary layer (e.g., 850 hPa), which is in sharp contrast to the strong convergence within the boundary layer (e.g., 925 hPa, Fig. 10). It has yet to be determined whether this lack of lower-tropospheric moisture convergence is another reason for the absence of strong MJO signals along the ITCZ in the central Pacific.

There are major caveats in this study. The first one is the quality of the global reanalysis product used. We have been cautious enough not to use surface or 1000-hPa moisture convergence out of the consideration of known problems in the surface meridional wind of the reanalysis. But we put our blind faith in the reanalysis. But we put our blind faith in the reanalysis moisture field may not be negligible (Trenberth 1999). Between the choices of having no data of mean moisture convergence at all and using data that might be erroneous, we hesitatingly took the latter.

The second caveat is the inconsistency in some of our results. The apparent similarity between the seasonality in the MJO and its mean background state is the best in the western Pacific, less so in the eastern Pacific, and does not apply mostly in the Indian Ocean. The monsoon circulation and convection in the Indian Ocean are not very complicated in boreal winter and should not be the reason for the lack of connection between the MJO and its mean background state there. Our study failed to identify any single mean background variable that can explain well the seasonality in the MJO in both the Indian and western Pacific Oceans. It is also intriguing that the precipitation component of the MJO in the Indian Ocean shows much less seasonal migration than the zonal wind component (e.g., Figs. 5 and 6). The discrepancies between the two, even though they are derived from an SVD analysis, suggest that the wind–precipitation relationships vary substantially between the two ocean basins and among different seasons. The Indian Ocean is where the MJO is initiated and the Pacific is where it deceases (at least in convection). One cannot help wonder whether the contrast between the observed seasonality in the two MJO components and between the MJO and its mean background state in the two ocean basins reflects possible differences in the dynamics of the MJO during its different stages of life cycle. Effects of land may provide an alternative explanation. Maximum MJO activities appear to stay away from landmass, whether it is the Maritime Continents, Australia, the Indian subcontinent, Southeast Asia, or Central America (Figs. 5 and 6). Several hypotheses have been proposed to explain this (Wang and Li 1994; Zhang and Hendon 1997) but none has been verified. To address these problems, we need numerical models that can reproduce the correct seasonality as well as structure and life cycle of the MJO. It would be interesting to see how many models can meet this strenuous requirement.

Acknowledgments. Abderrahim Bentamy kindly made the ERS surface wind data available. This study was supported by NSF through Grant ATM9912297.

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