Symmetric and antisymmetric Madden-Julian oscillation signals
in tropical deep convective systems

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ABSTRACT

This work studies the significance of asymmetry in systems of organized convection associated with the Madden-Julian oscillation (MJO). Satellite infrared brightness temperature data for the period 1983–2006 were decomposed into subsets symmetric and antisymmetric about the equator. Using a recent nonlinear objective method called nonlinear Laplacian spectral analysis, modes of variability were extracted representing symmetric and antisymmetric features of MJO convective systems, along with a plethora of other modes of tropical convective variability. It was found that the boreal winter MJO emerges as a single pair of modes in both symmetric and antisymmetric convection signals. Phase composites of the corresponding kinematic and thermodynamic anomalous fields were constructed using reanalysis data. It was found that the predominantly symmetric convective systems are potentially short-lived due to equatorial dry air intrusion eradicating equatorial convection when the MJO crosses the Maritime Continent. The predominantly antisymmetric convective systems, however, are less affected by dry intrusion; the strength of the MJO convective systems as well as anomalous circulations can be maintained and enhanced in the West Pacific. It was also found that the off-equatorial convective systems enhanced during the MJO are mostly deep convection and stratiform anvils, unlike the typical complex of shallow-congestus-deep convection to stratiform anvils on the equator. The multiscale interactions between the diurnal, MJO, and ENSO modes of convection were studied. It was found that the symmetric component of MJO convection is out of phase with the symmetric component of the diurnal cycle, while the antisymmetric component of MJO convection is in phase with the antisymmetric diurnal cycle. The former relationship breaks down during strong El Niño events, and both relationships break down during prolonged La Niñas.
1. Introduction

The Madden-Julian oscillation (MJO, e.g., Madden and Julian 1971, 1972) is an eastward-propagating, planetary-scale envelope of organized convective activity in the tropics. Characterized by gross features in the 20–90-day intraseasonal time range and zonal wavenumber 2–4, it dominates tropical predictability in subseasonal time scales. Moreover, through tropical–extratropical interactions, it influences global weather and climate variability, fundamentally interfacing short-term weather forecasts and long-term climate projections (Waliser 2005). Its convective, precipitation, and wind signals migrate latitudinally with the seasonal cycle; the strongest boreal winter MJO precipitation is asymmetric about the equator (e.g., Zhang and Dong 2004), implying asymmetric heat sources associated with convection. The simplest theoretical model for the MJO describes the phenomenon as the planetary-scale response of a moving heat source with prescribed propagation speed through the linear shallow-water equations for a first baroclinic mode (e.g., Matsuno 1966; Gill 1980; Chao 1987). Atmospheric responses to heat sources symmetric and antisymmetric about the equator are therefore of fundamental interest (Gill 1980). Biello and Majda (2005), in their multiscale model for the planetary-scale circulation associated with MJO, have demonstrated the differences between equatorial and off-equatorial convective heat sources on the solutions. In reality, symmetric and antisymmetric heat sources often coexist during an MJO event.

The goal of this work is to study the significance of asymmetry in the MJO convective systems by contrasting predominantly symmetric versus predominantly antisymmetric events in the observational record. In particular, we study the symmetric and antisymmetric components of satellite infrared brightness temperature ($T_B$) data over the tropical belt, extracted using an averaging method (e.g., Yanai and Murakami 1970; Wheeler and Kiladis 1999). The differences and similarities of the extracted spatiotemporal patterns open up a route to elucidate the interactions of the MJO with other important weather and climate processes, including the diurnal cycle and ENSO. These objectives require meticulous analysis procedures, for convectively coupled tropical motions are highly nonlinear and multiscaled
in time and space.

Theory has suggested that the MJO is a nonlinear oscillator (Majda and Stechmann 2009, 2011); in observations it was found that MJO may well be a stochastically-driven chaotic oscillator (Tung et al. 2011). Conventional methods for extracting the tropical waves and disturbances from observations and models are linear, including Fourier-based linear band-pass filtering, regression, and empirical orthogonal functions (EOFs) (Waliser et al. 2009). Even though substantial advances in the understanding of tropical waves and their linear theories have been guided by these diagnostics (e.g., Kiladis et al. 2009), linear filtering of a nonlinear, chaotic system may impede fundamental understanding of the system (Badii et al. 1988). Here, we address the challenges associated with nonlinearity of tropical dynamics via two approaches.

First, we apply nonlinear Laplacian spectral analysis (NLSA, Giannakis and Majda 2012a,c, 2013; Giannakis et al. 2012) to extract the spatiotemporal modes of the symmetric and antisymmetric components of the MJO from satellite infrared brightness temperature ($T_B$) data over the tropical belt. NLSA builds a set of data-driven orthogonal basis functions on the discretely-sampled nonlinear data manifold. By lag-embedding the observed data via the method of delays, those basis functions differ crucially from classical Fourier modes in that they contain information about the time evolution (dynamics) of the system under study, and are also adapted to the geometrical structure of the data in phase space. Compared to extended EOFs (EEOFs), singular spectrum analysis (SSA) and related variance-optimizing algorithms, NLSA has high skill in capturing intermittent patterns, which carry little variance but may be of high dynamical significance (Crommelin and Majda 2004). Moreover, the method applies no preprocessing such as seasonal detrending or band-pass filtering, allowing one to simultaneously study processes spanning multiple timescales (here, interannual–diurnal).

In the second stage of the analysis, nonlinear locally-smooth adaptive filtering (LSAF, Tung et al. 2011; Gao et al. 2011) is used to temporally filter $T_B$, as well as wind, temperature,
humidity from reanalysis, and their derived heat and moisture budget residuals. LSAF is applied prior to reconstructing the dynamic and thermodynamic fields associated with the significant symmetric and antisymmetric MJO events identified through the NLSA modes.

Combining NLSA and LSAF in this manner, we find that predominantly antisymmetric MJO convective systems differ prominently from their symmetric counterparts in that they are able to propagate across the Southeast Asia-Australia Maritime Continent and into the South Pacific convergence zone (SPCZ) without significant inhibition. We attribute this difference to the enhanced presence of off-equatorial deep-convective and stratiform anvil cloud types in predominantly antisymmetric MJO events, as opposed to the equatorial shallow-to-deep convection and stratiform complex which is active during symmetric events but tends to be suppressed by equatorial dry air intrusion. The symmetric and antisymmetric modes are also found to correlate in different ways with diurnal-scale processes over equatorial Africa, the Maritime Continent, and America, with ENSO also playing a role.

The paper proceeds as follows. In section 2, we describe the data and methods used in this study. We present and discuss our results in sections 3 and 4, and conclude in section 5.

2. Data and methods

a. CLAUS $T_B$: Proxy for tropical convective activity

We analyze multi-satellite infrared brightness temperature ($T_B$) data from the Cloud Archive User Service (CLAUS) Version 4.7 (e.g., Hodges et al. 2000). Brightness temperature is a measure of the earth’s infrared emission in terms of the temperature of a hypothesized blackbody emitting the same amount of radiation at the same wavelength ($\sim 10-11 \, \mu\text{m}$ in CLAUS). It is a highly correlated variable with the total terrestrial longwave emission. In the tropics, positive (negative) $T_B$ anomalies are associated with reduced (increased) cloudiness. The global CLAUS $T_B (\Lambda, \Phi, t)$ data are on a $0.5^\circ$ longitude ($\Lambda$) by $0.5^\circ$ latitude ($\Phi$) fixed grid, with three-hour time ($t$) resolution from 00 UTC to 21 UTC, spanning July 1, 1983
to June 30, 2006. The values of $T_B$ range from 170 K to 340 K at approximately 0.67 K resolution.

The subset of the data in the global tropical belt between 15°S and 15°N was taken to create symmetric and antisymmetric averages about the equator. The symmetric average is

$$ T_B(\Lambda, t) = \frac{1}{N} \sum_{\Phi=0^\circ}^{\Phi=15^\circ} \frac{T_B(\Lambda, -\Phi, t) + T_B(\Lambda, \Phi, t)}{2}, $$

where $N$ is the number of samples within the latitudinal ($\Phi$) range; negative values of $\Phi$ denote southern latitudes. Similarly, the antisymmetric average is

$$ T_B(\Lambda, t)_A = \frac{1}{N-1} \sum_{\Phi=0^\circ}^{\Phi=15^\circ} \frac{T_B(\Lambda, -\Phi, t) - T_B(\Lambda, \Phi, t)}{2}. $$

The resulting averages are longitude-time sequences sampled at $d = 720$ longitudinal grid-points in $\Lambda$ and $s = 67,208$ temporal snapshots in $t$. Prior to spatial averaging, the missing data (less than 1%) were filled via linear interpolation in time. Figure 1 shows the portion of the data for 1992–1993; a period which includes the Intensive Observing Period (IOP) of the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA COARE, November 1992–February 1993, Webster and Lukas 1992).

b. ECMWF Interim reanalysis

The European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Reanalysis (ERA-I, Dee et al. 2011) is a reanalysis product with a T255 60-level atmospheric model and four-dimensional variational assimilation system. We used the version of the ERA-I dataset archived at the National Center for Atmospheric Research Data Support Section. The data are defined on a regular $512 \times 256$ N128 Gaussian grid at 37 standard pressure levels from 1000 to 1 hPa, and sampled four times daily at 00, 06, 12, and 18 UTC. To facilitate our analyses, the horizontal wind $(u, v)$, vertical $p$-velocity $(\omega)$, temperature $(T)$, and specific
humidity (\(\gamma\)) variables were further interpolated onto a fixed 1.5° × 1.5° horizontal grid.

Following Yanai et al. (1973), we computed the apparent heat source, \(Q_1\), and apparent moisture sink, \(Q_2\), viz.

\[
Q_1 \equiv c_p \frac{p}{p_0} \kappa \left( \frac{\partial \theta}{\partial t} + \nabla \cdot \nabla \theta + \omega \frac{\partial \theta}{\partial p} \right) = Q_R + L(\bar{c} - \bar{c}) - \frac{\partial \bar{s}' \omega'}{\partial p}
\]

(3)

and

\[
Q_2 \equiv -L \left( \frac{\partial r}{\partial t} + \nabla \cdot \nabla r + \omega \frac{\partial r}{\partial p} \right) = L(\bar{c} - \bar{c} + \frac{\partial r' \omega'}{\partial p}),
\]

(4)

In the above, \(\theta\) is the potential temperature calculated from \(T\), \(r\) the water vapor mixing ratio converted from \(\gamma\), \(Q_R\) the radiative heating rate, \(L\) the latent heat released by condensation, \(s\) the dry static energy, and \(c\) and \(e\) the rates of condensation and evaporation (of cloud water) per unit mass of air, respectively. Moreover, the quantities \(p_0\) and \(\kappa\) are set to 1000 hPa, and \(R/c_p\), respectively, with \(R\) the gas constant of dry air. Overbars denote running horizontal average with respect to a large-scale area and the prime denotes the deviation from the average. The grid-scale variables of ERA-I are regarded here as “large-scale”.

The finite difference scheme used for computation is written so that the advective form is consistent with the flux form. \(Q_1\), computed via the grid-scale variables in the LHS of (3), is interpreted as the total effect of radiative heating, latent heat released by net condensation, and the convergence of fluxes of sensible heat due to subgrid-scale eddies such as convection and turbulence. \(Q_2\) corresponds to the total effect of net condensation and divergence of eddy moisture flux due to clouds and turbulence.

### c. NLSA algorithms

Nonlinear Laplacian spectral analysis (Giannakis and Majda 2012c,a, 2013) is a method for extracting spatiotemporal patterns from high-dimensional time series that blends ideas from the qualitative analysis of dynamical systems (Broomhead and King 1986; Sauer et al.)
1991), singular spectrum analysis (SSA, Aubry et al. 1991; Ghil et al. 2002), and spectral
graph theory for machine learning (Belkin and Niyogi 2003; Coifman and Lafon 2006). Unlike
principal components analysis (PCA), SSA, and related variance-maximizing algorithms,
NLSA is based on the premise that dynamically-relevant low-rank decompositions of the data
should be constructed via orthonormal basis functions which are intrinsic to the nonlinear
data manifold sampled by the observations. The basis functions in question are Laplace-
Beltrami (LB) eigenfunctions, computed via graph-theoretic algorithms (Belkin and Niyogi
2003; Coifman and Lafon 2006) after time-lagged embedding of the data to incorporate
information about time-directed evolution.

Consider an $n \times s$ data matrix $X$ consisting of $s$-samples of an $n$-dimensional variable. In
section 3 ahead, each column of $X$ will be either of the $T_B(A, t)$ or $T_B(A, t)_A$ fields from (1)
and (2), respectively, embedded over an intraseasonal lag window $\Delta t$. NLSA produces a
decomposition of the form

$$X \approx U_l \Sigma_l V_l^T \Phi_l^T,$$

(5)

with $U_l$ an $n \times l$ matrix with orthogonal columns, $\Sigma_l$ an $l \times l$ diagonal matrix of singular
values $\sigma_i$, $V_l$ an $l \times l$ orthogonal matrix of expansion coefficients, and $\Phi$ an $s \times l$
matrix of LB eigenfunction values. Each column $u_i$ of $U_l$ represents a spatiotemporal process of
temporal extent $\Delta t$ analogous to an EEOF. The corresponding temporal pattern, analogous
to a principal component (PC), is given by $\tilde{v}_i = \Phi v_i$, where $v_i$ is the $i$-th column of $V_l$.
The $\tilde{v}_i$ are orthogonal with respect to a weighted inner product with weights $\mu$ designed to
capture rapid transitions and rare events (Giannakis and Majda 2012c).

The parameter $l$ corresponds to the number of LB eigenfunctions used in the NLSA
decomposition, and in practice is significantly smaller than the number of samples. This
type of data compression is especially beneficial in large-scale applications where the ambient
space dimension $n$ and the sample number $s$ are both large (Giannakis and Majda 2013).
Each triplet $\{u_i, \sigma_i, \tilde{v}_i\}$ yields a spatiotemporal process $\tilde{X}_i = \sigma_i u_i v_i^T$ in lagged-embedding
space, which can be projected down to physical space to produce reconstructions of the input
signal (e.g., section 34).

The role of the transform matrix $\Phi_l$ in (5) is to select geometrically-preferred temporal patterns on the nonlinear data manifold $M$. Such patterns are viewed in NLSA as good candidates to produce physically-meaningful EEOFs through weighted averages of the form $u_i = X \mu v_i$. In contrast, the classical SSA decomposition,

$$X \approx U_{SSA} \Sigma_{SSA} V_{SSA}^T$$

makes no such provision, and the temporal patterns $V_{SSA}$ used for averaging can be arbitrary functions of time, without regards to the geometrical structure of $M$ other than global covariance. Note that the basis functions $\Phi_l$ can encode multiple timescales of temporal variability despite being weakly oscillatory on the data manifold. This is because time variability in $\Phi_l$ is an outcome of both their geometrical structure as functions on $M$, and the sampling of these functions along the trajectory on $M$ followed by the system under time evolution. For instance, the modes discussed in section 3 exhibit temporal variability spanning interannual to diurnal timescales, and yet are described in terms of a moderately small number of leading ("low-wavenumber") LB eigenfunctions.

The advantages of using (5) versus (6) have been demonstrated in a number of applications, including Galerkin reduction of dynamical systems where PCA is known to fail (Giannakis and Majda 2012c), and regression modeling in comprehensive climate models (Giannakis and Majda 2012a,b). Details of the NLSA methodology applied to the symmetrically averaged $T_B$, $\overline{T_B(A,t)}$, including parameter selection and computational considerations, are provided in Giannakis et al. (2012).

d. Temporal band-pass filtering using the nonlinear Locally-Smooth Adaptive Filter (LSAF)

Spatial patterns of the kinematic and thermodynamic fields associated with the MJO were reconstructed for the analysis in section 4. Prior to reconstruction, the intraseasonal
(20–90-day) component of each variable was extracted through band-pass filtering using the nonlinear LSAF. The latter was developed to recover nonlinear trends in noisy timeseries, and has been shown to be more effective than chaos-based approaches and wavelet shrinkage (Tung et al. 2011; Gao et al. 2011). Prior to processing via nonlinear LSAF, the transient eddy components of the ERA-I variables were computed by subtracting the zonal mean and the 1985–2005 temporal mean from each variable. These temporally band-pass filtered fields are symbolically denoted with a pair of brackets, [·], and called the “anomalous” fields hereafter. We use asterisks to distinguish band-pass filtered transient eddy fields (e.g., the anomalous zonal wind \([u^*]\)), from other variables (e.g., the anomalous CLAUS \([T_B]\)).

3. The NLSA modes

a. Spatiotemporal patterns revealed by NLSA

We have applied the NLSA algorithm described in section 2c using an intraseasonal embedding window spanning \(\Delta t = 64\) days. This choice of embedding window was motivated from our objective to resolve propagating structures such as the MJO with intraseasonal (30–90 day) characteristic timescales. Unlike conventional approaches (e.g., Kikuchi et al. 2012; Maloney and Hartmann 1998; Wheeler and Hendon 2004), neither band-pass filtering nor seasonal partitioning was applied to the CLAUS dataset prior to analysis using NLSA.

The singular values \(\sigma_i\) corresponding to the modes in (5) are displayed in Fig. 2. In Fig. 2a, several representative modes relevant to our analysis are identified in the symmetric \(T_B\) data, namely an annual mode \((i = 1)\), an interannual mode \((i = 2)\), a pair of MJO modes \((i = 3, 4)\), a pair of diurnal modes \((i = 5, 6)\), and a “Maritime Continent” mode \((i = 11)\). It takes at least four modes to account for the annual and seasonal cycles in the antisymmetric \(T_B\) data. In Fig. 2b, in addition to an annual mode \((i = 1)\) similar to that in the symmetric data, the subsequent three modes are likely associated with African and American monsoons \((i = 2)\), annual cycle of the eastern Pacific ITCZ \((i = 3)\), as well as the
Asian-Australian monsoons and the seasonal cycle of the SPCZ ($i = 4$). Further study of these modes is not within the scope of this paper, since the types of variability represented by these modes are best examined in the future through NLSA of the full two-dimensional (2D) fields. Further modes of interest in Fig. 2b are two MJO modes ($i = 5, 7$) and two diurnal modes ($i = 6, 8$). Unlike their symmetric counterparts, the antisymmetric MJO and diurnal modes do not appear in consecutively in the $\sigma_i$ spectrum. It is later found that the amplitude of these diurnal modes is strongly modulated by the MJO.

Representative spatial ($u_i$) and temporal ($v_i$) patterns associated with the modes described above are displayed in Figs. 3 and 4. For consistency with Fig. 1, the temporal patterns are shown for the duration of 1992–1993 in Fig. 4, along with their frequency spectra calculated from the entire $\sim 20$-year-long time series. The spatial patterns of the symmetric annual and interannual modes are nearly constant within the embedding window $\Delta t$ (Figs. 3a,b), while their corresponding temporal patterns reveal that these two modes explain the seasonal and longer-time variability of cloudiness in the tropical belt. Figure 3b features a dipole over the equatorial Pacific, suggesting that this mode is associated with ENSO. The symmetric Maritime Continent mode (Figs. 3c and 4c), the symmetric MJO modes (MJO1 and MJO2; Figs. 3d,e and 4d,e), as well as the antisymmetric MJO modes (MJO1$A$ and MJO2$A$; Figs. 3g,h and 4g,h) are eastward propagating modes with pronounced intraseasonal variability. They feature broadened peaks centered around 20–90 days in the frequency spectra (Figs. 4c–e,g,h). The spatial patterns of MJO1 and MJO2 are in quadrature, indicating that together they represent the propagation of the pattern of MJO convective systems (roughly, of global wavenumber two) over the Indian Ocean–West Pacific sector. The spatial patterns of MJO1$A$ and MJO2$A$ exhibit a similar behavior. Unlike the symmetric modes, whose maximum values tend to be confined over the oceans (Figs. 3c–e), the antisymmetric MJO modes have significant local extrema over land mass such as around $30^\circ$E, $120^\circ$E, and $75^\circ$W (Figs. 3g,h). Moreover, the global extrema of the symmetric modes are in the Indian Ocean around $90^\circ$E, while those of the antisymmetric modes extend further
east into the Pacific such as 120°E and the date line. All three symmetric modes are more or less suppressed around 90°E.

The spatial patterns of both symmetric diurnal modes (Fig. 3f) and antisymmetric diurnal modes (Fig. 3i) are most prominent over land, where the diurnal cycle of convection is most active. The major difference between the symmetric and the antisymmetric modes is seen in the Western Hemisphere over America and part of Africa. For instance, unlike the symmetric modes, the antisymmetric diurnal cycle around 60°W is as strong as that over 30°E. Such contrast can be easily explained by the antisymmetric land mass over the Western Hemisphere in the analysis domain. The more subtle difference is over the Maritime Continent area, where the antisymmetric diurnal cycle has more substantial magnitude. The temporal patterns of the symmetric diurnal modes (Fig. 4f) are drastically different from those of the antisymmetric modes (Fig. 4i). The former appear to take place year-round, while the latter are strongly modulated by the seasonal cycle. Mode Diurnal1A (not shown) is stronger in the boreal winter-spring and weaker in the boreal summer-fall. Mode Diurnal2A (Fig. 4i) is evidently a boreal winter-spring mode.


We validate our mode reconstructions using the well-studied MJO events occurring in the TOGA-COARE IOP (November 1992–February 1993). The two complete MJOs observed during that period (e.g., Yanai et al. 2000) can be seen in the longitude-time section of the symmetric CLAUS $T_B$ in Fig. 1a. Marked by two distinct envelopes of cold $T_B$ “supercloud clusters” (Nakazawa 1988), these systems propagated eastward from the Indian Ocean to the date line. The first event initiated near 75°E in late November, subsequently crossed the Maritime Continent around 100°–150°E, and disappeared near 170°W around January 10. The second event, being slightly faster than the first, started around January 5, and reached the central Pacific in early February. Concurrent signals in the antisymmetric $T_B$ can be seen for these two events (Fig. 1b). A weaker event took place in October 1992,
prior to the start of the TOGA-COARE IOP. Unlike the two strong cases, this event neither exhibited a significant symmetric component beyond the Maritime Continent (Fig. 1a), nor left a discernible trace in the antisymmetric $T_B$ (Fig. 1b).

The TOGA-COARE period was coincident with the amplifying phase of an El Niño event. Therefore, the MJO events propagated further east beyond the date line, where during normal years the cold sea surface temperature is not conducive to deep convection. The influence of ENSO on MJO propagation is particularly evident in the January–May 1992 portion of Fig. 1, where ENSO was stronger than during the TOGA-COARE IOP. MJO convection not only propagated further east beyond the date line during that period, but also exhibited significant antisymmetric characteristics.

In all of the above cases, the eastward propagating speed of the MJO was $\sim 4\text{–}5 \text{ m s}^{-1}$. The convective systems around the Maritime Continent were especially complicated before, during, and after the passages of the MJO. In addition, two regions of apparently standing convection over equatorial Africa and America were observed.

Figure 5 shows the 1992–1993 reconstruction of the modes identified with unique markers in Fig. 2 except for the symmetric annual mode. First, the symmetric interannual mode in Fig. 5a exhibits periods of reduced cloudiness over the western Pacific accompanied with enhanced cloudiness over the eastern Pacific. The latter are features associated with ENSO. The interannual mode is an amplifying phase during the TOGA-COARE IOP, and in a strong phase in January–May 1992. These features are consistent with the ENSO observations stated earlier. The symmetric Maritime Continent mode, shown in Fig. 5b, exhibits an eastward propagating disturbance with a speed of $\sim 7\text{–}8 \text{ m s}^{-1}$. It mainly comprises of two deep convective systems, each with a zonal scale of order 500 km, centered around 90°E and 135°E, respectively. This mode may represent convective activity around the Maritime Continent, which exists on its own but is modulated by the passing of the MJO.

Figure 5c shows the symmetric MJO reconstruction, which captures the succinct features of the propagating envelope of the super clusters, including the initiation of enhanced cloudi-
ness (hence cold anomalies) over the Indian Ocean, the passage over the Maritime Continent, and the arrival and demise near the date line. The two reconstructed COARE IOP MJO events propagate at a speed of $\sim 4-5 \text{ m s}^{-1}$. As seen in Fig. 5e, the antisymmetric MJO reconstruction also captures these two MJO events, although the November 1992 event has a rather weak signal. The antisymmetric MJO signal is most pronounced during the strong ENSO in January-May 1992 in Fig. 5e. Lastly, the symmetric MJO signal is likely the only mode present at the very first initiation of a chain of MJO events.

The symmetric diurnal modes are a twofold-degenerate pair (Fig. 2). Upon reconstruction, they reveal the standing diurnal convective events occurring mainly over tropical Africa and America (Fig. 5d). The signals over the Maritime Continent are relatively weak. These diurnal cycles are moderately modulated by the passing of MJO, particularly over the American continent, but generally exist year-round. On the other hand, the antisymmetric diurnal modes in Fig. 5f are obviously in phase with the MJO over Africa, the Maritime Continent, and America.

c. Comparison with SSA

For completeness, we have compared the NLSA spatiotemporal patterns with the corresponding patterns recovered through SSA using the same $T_B(A,t)$ and $\hat{T}(A,t)$ datasets and $\Delta t = 64 \text{ d}$ embedding window. The main commonalities and differences between the two approaches are as follows.

For the first few modes in the top of the spectrum, the spatial patterns $u_i$ are qualitatively similar, but the corresponding temporal patterns $v_i$ and spatiotemporal reconstructions from SSA exhibit significantly weaker amplitude modulation than in NLSA. This behavior is especially prominent in the antisymmetric MJO modes, where instead of the sharp amplitude modulation favoring winter MJO activity shown in Fig. 5, the SSA MJO patterns occur as more or less continuous wavetrains, with no clear distinction between summer and winter activity. Likewise, the antisymmetric diurnal modes from SSA persist at a constant am-
plitude throughout the year without regards to season and/or passing of an MJO. Similar statements can be made about the symmetric MJO and diurnal modes, although in this case the differences in amplitude modulation between the NLSA and SSA modes are not as significant.

As one might expect, the two methods differ qualitatively for the low-variance modes lying further down in the spectrum (for other examples, see Giannakis and Majda 2012a,c). In particular, we have found no evidence of a mode analogous to the Maritime Continent mode in the SSA spectrum. Instead, the SSA spectrum contains several wavetrain-like modes featuring simultaneous eastward- and westward-propagating structures with no obvious physical interpretation. Moreover, the NLSA spectrum of the symmetric data contains a second set of diurnal modes (not discussed in this paper), which are mainly active over the Amazon region during boreal summer. A second set of diurnal modes also arises in SSA, but these modes are active over both the Congo and Amazon regions (i.e., they are qualitatively similar to the leading set of symmetric diurnal modes), and exhibit weak amplitude modulation.

d. Frequency-Wavenumber Spectra

Figure 6 shows the frequency-wavenumber spectra of the 1984–2005 symmetric and antisymmetric $T_B$ data, respectively. The spectral power in each panel has been normalized by its the maximum value. The dispersion curves were calculated by assuming a static base state, with the marked equivalent depths for each equatorial wave type. These raw spectra indicate that eastward-moving MJO is the most dominant signal in both symmetric and antisymmetric $T_B$. Indeed, both reconstructed symmetric and antisymmetric MJO modes exhibit their strongest spectral peaks in the eastward-moving, wavenumber 1–3, and 30-90-day range (Figs. 7a,b). On the other hand, the antisymmetric MJO has relatively stronger westward-moving components, and spreads more into higher wavenumbers than the symmetric MJO. The symmetric Maritime Continent mode (Fig. 7c), has a higher frequency peak at around 30 days. According to the dispersion curves, this mode may have
convection-coupled Kelvin and Rossby wave components. However, since the averaging pro-
process to obtain the 1D spatiotemporal symmetric $T_B$ data in our analysis has compressed
information, especially over the challenging Maritime Continent area, we refrain ourselves
from further interpretation of this mode.

e. Linkage from diurnal to interannual scales

Figure 8 shows 2D phase spaces spanned by pairs of temporal patterns during 1992–1993.
The arc-like trajectories in Figs. 8a,b indicate strong periodicity in the phenomena described
by the two symmetric (MJO1, MJO2) and antisymmetric (MJO1$_A$, MJO2$_A$) MJO modes.
The convectively active periods of the two TOGA-COARE MJO events, identified visually
from Fig. 1a, are recorded here with green crosses and red dots marking the first and second
events, respectively. Define the amplitude of an event at a given time as the distance between
the origin and the point on the trajectory at that time, i.e.

$$r_S(t) = [\text{MJO1}^2(t) + \text{MJO2}^2(t)]^{1/2}, \quad r_A(t) = [\text{MJO1}_A^2(t) + \text{MJO2}_A^2(t)]^{1/2}$$

Under this criterion, the second event has obviously stronger amplitude than the first. Each
phase space can be divided into eight sections indicating eight MJO phases, which are marked
by Roman numerics in Figs. 8a,b. These phases, which are cyclic by definition, are calibrated
so that Phase I encompasses the time interval during which convective signals of the TOGA-
COARE MJOs initiate over the equatorial Indian Ocean. Spatial field reconstructions for
each phase are discussed in details in section 4.

Figures 8c,d display the two symmetric (Diurnal1, Diurnal2) and antisymmetric diurnal
(Diurnal1$_A$, Diurnal2$_A$) modes, respectively. As expected by the diurnal periodicity of these
modes and the 3 h sampling interval of the data, the phase-space coordinates lie along
rays passing through the origin, and separated by 45° polar angles. Comparing against
the respective MJO modes, an interesting pattern emerges: the relatively strong symmetric
MJO component during TOGA COARE coincides with suppression of the symmetric diurnal cycle. In contrast, the stronger antisymmetric MJO component is associated with enhanced antisymmetric diurnal cycle. Because the antisymmetric diurnal modes are slaved to the seasonal cycle (section 3a), it is natural to ask whether the simultaneous amplification of the antisymmetric diurnal and MJO modes observed during TOGA COARE implies more broadly a seasonal regularity of antisymmetric signals in MJO events. Any deviation from such regularity would manifest itself as a breach between simultaneously large values of $r_A(t)$ and the corresponding amplitude associated with the antisymmetric diurnal modes. Indeed, breaches of this type occur frequently in the two decades of available $T_B$ data, and are correlated with the amplitude and sign of the ENSO mode, as we now discuss.

Figure 9 shows the temporal pattern of the ENSO mode and the amplitudes of the symmetric and antisymmetric MJO and diurnal modes. Strong and persistent positive values in Figs. 9a1 and 9b1 indicate El Niño events, such as years 1986–1987, 1991–1992, 1994–1995, 1997–1998, and 2002–2003. On the other hand, prolonged negative values such as those from 1988–1989 and 1999–2001 mark the La Niña events. Note that the significant events in the amplitude time series for the MJO and diurnal modes are only those with large positive values. Specifically, only cases with standardized amplitude larger than one standard deviation (i.e., the top 16th percentile of a Gaussian distribution) are considered significant enough for the phase reconstruction in section 4.

Because of their regular seasonal variability, the amplitude of the antisymmetric diurnal modes in Figs. 9a3 and 9b3 is a useful indicator to distinguish between the winter–spring and summer–fall months. The latter are characterized by high and low antisymmetric diurnal amplitude, respectively (note that the amplitude of the symmetric diurnal modes has no such seasonal dependence). It is then evident that even though the MJO modes are mostly winter–spring modes in Fig. 9a2, they appear much more irregular in Fig. 9b2. Such disparities may be explained with the chain of pronounced ENSO events starting with strong El Niño followed by years of La Niña states from early 1997 to late 2001. Upon close examination, at
the amplifying stage of the El Niño event in 1997, both symmetric and antisymmetric MJO components were enhanced. Similar situations may have also taken place in 1990 and 2002. However, after the El Niño reached its full strength in 1998, the symmetric MJO component diminished, and so did the symmetric diurnal mode. The antisymmetric MJO and diurnal modes remain unaffected. This simultaneous suppression of symmetric MJO and diurnal modes also occurred in the winter of the 1991–1992 El Niño. During La Niña years, most notably from 1999–2001, both symmetric and antisymmetric MJO modes are suppressed, while the diurnal modes are unaffected or even enhanced. The interplay between the ENSO, MJO, and diurnal modes is also evident in Fig. 5.

In summary, during neutral and weak ENSO years, the symmetric MJO and diurnal modes are out of phase, whereas the antisymmetric MJO and diurnal modes are in phase. During significant El Niño events, the former relationship breaks down in the sense that the symmetric MJO and diurnal modes are both suppressed, probably due to strongly skewed MJO and convective activity in space. During significant La Niña events, the correlation between MJO and diurnal modes diminishes. This might be explained by the westward shift of warm SST in the Pacific Ocean, limiting the eastward propagation of MJOs, and thus weakening the amplitude of the associated temporal patterns. These conjectures should be further examined with NLSA applied to timeseries of 2D $T_B$ fields.

4. MJO kinematic and thermodynamic reconstructions by phase

We now use the symmetric and antisymmetric MJO mode phases identified in section 3e to create composites of kinematic and thermodynamic fields associated with MJO convective systems. In particular, we consider the anomalous (intraseasonal band-pass averaged) fields $[T_B], [Q_1], [(u^*, v^*, \omega^*)], [T^*], \text{ and } [\gamma^*]$, which were obtained via the procedure described in section 2d. These reconstructions allow one to visualize and physically interpret the 2D and
3D spatial structures corresponding to the MJO modes extracted from the 1D averaged data.

In order to create the composites, samples with large MJO modal amplitudes \( r_S \) and \( r_A \) from (7) were isolated and averaged over Phases I–VIII of Figs. 8a and 8b. The amplitude thresholds for averaging were equal to the time mean plus one standard deviation; i.e., we selected samples with amplitude exceeding 1 in Figs. 9a2 and 9b2. Note that there are MJO events with \( r_S \) and \( r_A \) simultaneously exceeding that threshold. This is because the \( T_B \) field is always positive, and therefore strongly off-equatorial signals produce symmetric and antisymmetric averages of comparable magnitude. Such cases, which we call overlap samples, amount to 42% and 56% of the symmetric and antisymmetric strong events, respectively. Because the phases of the symmetric and antisymmetric modes may differ at a given time, the number of overlap samples at each phase is smaller (28% and 37% of the total samples, respectively) but non-negligible. The resulting reconstructions are therefore called “predominantly” symmetric or antisymmetric. Table 1 summarizes the symmetric, antisymmetric, and overlap sample counts for each MJO phase. The overlap samples are included in both symmetric and antisymmetric phase reconstructions so that their likeliness to each group is accented.

Data samples of the anomalous fields corresponding to the identified strong events were selected and partitioned into the eight phases. Then, an average was taken over the aggregated samples in each phase. To minimize statistical assumptions, we did not use linear projection or correlation methods to reconstruct these fields. The selection criteria are robust-enough so that predominantly symmetric and antisymmetric convective signals associated with the MJO emerge naturally in the reconstructed \([T_B]\) fields, shown in Figs. 10 and 15. Concurrent horizontal maps of \([Q_1]\), \([\langle u^*, v^* \rangle]\), \([\gamma^*]\), and \([T^*]\) are displayed in Figs. 11 and 15. Figures 12–14 and 17–19 show vertical profiles of tropospheric \([\langle u^*, \omega^* \rangle], [\gamma^*], [T^*], [Q_1], \) and \([Q_2]\) on the equator (averaged over 3°S–3°N) and off the equator (averaged over 10.5°–4.5°S). In addition to showing anomalous diabatic heating and drying, the \([Q_1]\) and \([Q_2]\) fields in Figs. 14 and 19 are indicative of the changes of cloud populations associated with the MJO. Together,
these fields establish the 3D kinematic and thermodynamic structures associated with the
two types of MJO deep convective systems.

Recall that Phase I was chosen for the time when incipient MJO signals in \[ T_B \] are seen
over the Indian Ocean and remnant convection of a previous MJO is still seen in the SPCZ.
In the ensuing discussion, the 3D structures associated with the predominantly symmetric
MJO deep convective systems are examined in four stages in the MJO life cycle: (i) initiation
in the Indian Ocean, (ii) passage over the Maritime Continent, (iii) intensification over West
Pacific and SPCZ, and (iv) decoupling from deep convective systems (the dry phase) in the
Western Hemisphere. Then, the structures of the predominantly antisymmetric MJO deep
convective systems are discussed with emphasis on the differences from their predominantly
symmetric counterparts. Because deep convective systems do not play a major role in stage
(iv), this stage is only discussed for the predominantly symmetric events.

1) Predominantly symmetric MJO convective systems

(i) Initiation in the Indian ocean

Figure 10a shows the the eastward progression of the predominantly symmetric MJO
convective activities represented by the \[ T_B \]. An interesting feature in Phases I–III is the
transition over equatorial East Africa (\( \sim 40^\circ \)E) from a convectively suppressed state to
an active state, while convection over the equatorial western Indian Ocean (\( \sim 60^\circ–90^\circ \)E)
is enhanced. These can also be seen in the 475-hPa \[ Q_1 \] field (Fig. 10b), in which a set
of equatorially symmetric heating anomaly maxima is originally at 60\(^\circ\)E in Phase I, then
migrates to 90\(^\circ\)E by Phase III; meanwhile, the eastern Africa is gradually encroached by
increased \[ Q_1 \] heating associated with the convection indicated by \[ T_B \] (Fig. 10a). During
the same period, lower-tropospheric moisture builds up in \( \sim 60^\circ–90^\circ \)E, as seen in the 750-
hPa \[ \gamma^* \] (Fig. 11a, Phases I–III). The latter is likely associated with moisture advection from
the east indicated by the overlaid 850-hPa anomalous wind vectors.
A set of off-equatorial moisture anomalies are seen in 750-hPa $[\gamma^*]$ over Africa around 20º–40ºE in Phase I. One is centered at $\sim$ 10ºN, the other is at $\sim$ 20ºS. They move off Africa coast by Phase III (Fig. 11a). These moist anomalies are located in the convergence zones of a 850-hPa twin cyclonic system drawing moisture from the Atlantic and Indian Oceans. These twin cyclonic anomalies appear to be part of an incipient MJO quadrupole, which is more evident in the 200-hPa anomalous winds (Fig. 11b). At 200 hPa, the twin cyclonic anomalies are centered around 70º–90ºE in Phase I, associated with 400-hPa cold $[T^*]$ (Fig. 11b). They move eastward, followed by twin anticyclonic anomalies associated with 400-hPa warm anomalies; the latter system arrives at 60ºE by Phase III. In Fig. 11b, an elongated area of 400-hPa warm $[T^*]$ at $\sim$ 60º–150ºE emanates from below the 200-hPa divergence center around 60º–80ºE. It appears to be correlated with convective systems and 475-hPa $[Q_1]$ heating (Fig. 10 and in vertical profiles). According to Yanai et al. (2000), such association of enhanced convective activity and warm anomalies implies positive correlation between heating and temperature, and therefore production of eddy available potential energy.

The vertical profiles on the equator show that the aforementioned initial transition over equatorial East Africa is associated with anomalous descending motions in Phase I (Fig. 12a), hence adiabatic warming and drying (Fig. 13a). The anomalous descents change to anomalous ascents in Phase III (Fig. 12a). Meanwhile, between 60º–150ºE, an area of tropospheric ascending motions develops in a pulsating manner, with scattered ascents leading eastward of fully developed ones. Vertical gradients of $[\omega^*]$ indicate that horizontal convergence is mostly confined in the lower troposphere in 90º–120ºE in Phase I, then extends upwards to the mid-troposphere in Phases II and III (Fig. 12a). A westward-tilted anomalous circulation cell fully develops in Phase III. A few degrees to the south of the equator, however, the anomalous circulation is not as significantly tilted (Fig. 12b).

Along the equator at 60º–120ºE, the low-tropospheric convergence anomalies appear to lead low-tropospheric moistening in Phases I and II (Figs. 12a and Fig. 13a). Otherwise, on and off the equator, the upward motions are collocated with moist anomalies with maxima
at $\sim$ 800–700 hPa (Figs. 12 and 13). The $[T^*]$ profiles in Fig. 13 indicate two types of warm anomalies, respectively associated with dry and moist anomalies. The warm and dry anomalies likely result from adiabatic descents, e.g., $\sim$700–500 hPa around 120$^\circ$E in Phases I and II (Figs. 12 and 13). The warm and moist anomalies may be associated with deep convective production of eddy available potential energy, e.g., $\sim$400–200 hPa around 60$^\circ$E in Phases I and II and around 90–120$^\circ$E in Phase III (Figs. 12a,b and 13a,b).

On the equator, the anomalies in the convective systems display a conspicuous transition of dominant cloud types from east to west (Fig. 14a) likely featuring, (i) shallow clouds associated with negative $[Q_1]$ and $[Q_2]$ within the planetary boundary layer, (ii) precipitating shallow clouds and cumulus congestus with positive $[Q_1]$ and $[Q_2]$ between 1000–700 hPa, (iii) precipitating deep convection with $[Q_1]$ peaks between 600–300 hPa and $[Q_2]$ maxima between 900–700 hPa, and (iv) stratiform anvils with $[Q_1]$ and $[Q_2]$ both at around 400 hPa. Off the equator, shallower precipitating cloud types appear to be lacking, while deep convective and (especially) stratiform heating and drying account for most of the $[Q_1]$ and $[Q_2]$ anomalies (Fig. 14b).

(ii) Passage over Maritime Continent

The 850-hPa wind anomalies in the equatorial western Indian Ocean turn from easterlies in Phase III to westerlies in Phase IV, marking the eastward propagation of the MJO quadrupole (Fig. 11). The latter is coupled with deep convection signals in $[T_B]$ and $[Q_1]$ and low-tropospheric moisture anomalies in $[\gamma^*]$ at around 90$^\circ$E (Figs. 10 and 11a). The 200-hPa twin anticyclonic anomalies are enhanced in Phase IV (Fig. 11b), along with enhanced 400-hPa warm $[T^*]$. At this stage, the conversion of eddy available potential energy produced via convective heating into eddy kinetic energy associated with the MJO may also be enhanced along the equator. Downstream, equatorial deep convection exists eastward of 120$^\circ$E in $[T_B]$ and 475-hPa $[Q_1]$, and is collocated with weak 750-hPa moist $[\gamma^*]$ and 400-hPa warm $[T^*]$ (Figs. 10 and 11).
In Phase V, the deep convection center separates into two (Fig. 10), along with two 850-hPa cyclonic anomalies and moist $\gamma^*$ straddling the equator (Fig. 11a) upon reaching Sumatra around 100°E. The 200-hPa twin anticyclonic anomalies continue to be energized along with an enhanced 400-hPa warm $T^*$ (Fig. 11b). A dry westerly intrusion seen in 750-hPa $\gamma^*$ evidently develops upstream of the 850-hPa cyclonic anomalies. Downstream equatorial deep convection and low-tropospheric moist anomalies persist eastward of 120°E with some signs of enhancement (Figs. 10b and 11a). The dry westerly intrusion appears to be detrimental to further development of equatorial deep convection, which supplies perturbation available potential energy to the MJO. In particular, the low-level dry air in Phase VI penetrates across the entire equatorial Maritime Continent, dividing the deep convective systems to the north and to the south of the equator in the 100°–180°E range (Figs. 10 and 11).

The vertical profiles on the equator show that the westward-tilted anomalous circulation cell propagates eastward, despite interruptions over the Maritime Continent between 100°–120°E (Fig. 12a). In Phase IV the leading anomalous low-tropospheric convergence reaches the date line; in Phase V the major tropospheric ascent region bypasses the Maritime Continent and arrives in 120°–150°E. The Maritime Continent does not appear to be an obstacle to the anomalous circulation a few degrees south of the equator (Fig. 12b). Traces of low- and mid-tropospheric anomalous convergence are seen over 150°–180°E in Phase IV, which later join with more coherent ascents at 120°–150°E in Phase V. Unlike the equatorial signals, the enhanced off-equatorial tropospheric ascents are confined to the west of 120°E (Fig. 12). It is not until Phase VI that, as the equatorial ascents diminish, the off-equatorial anomalous ascents make a definite move to the West Pacific.

At this stage of the MJO lifecycle, the anomalous ascents are collocated with moist anomalies both on and off the equator (Figs. 12 and 13). The warm and dry anomalies divide the anomalous upward and downward motions, although the most pronounced tropospheric warm anomalies now are those collocated with moist anomalies around 90°–120°E in Phases IV and V (Figs. 12 and 13a,b). In Phase IV, the equatorial profiles of $[Q_1]$ and
[Q_2] continue to display the transition of dominant cloud types (Fig. 14a). This spectrum of transitions is disrupted in Phase V, however, leaving mostly deep convection and stratiform anvils, followed by a pronounced low-level cooling and moistening anomalies likely unrelated with convection between 60°–90°E (Fig. 14a). Off the equator, deep convective and stratiform heating and drying anomalies persist (Fig. 14b), featuring similar propagating signals as the anomalous ascents (Fig. 12b). Under the influence of equatorial dry westerly intrusion, the apparent heating and drying signals along the equator diminish entirely in Phase VI, while the off-equatorial signals propagate with weakening strength (Figs. 12 and 14). At the same time, the 400-hPa warm and moist anomalies are also weakened both on and off the equator (Fig. 13).

The dry air intrusion over the Indian Ocean has been observed during the Dynamics of Madden-Julian Oscillation (DYNAMO) field campaign in radiosonde data (Johnson, personal communication), as well as satellite and aircraft data (Kerns, personal communication). The MJO forecasted by the NOAA GFS model decayed prematurely upon reaching the Maritime Continent due to dry intrusion associated with an unrealistically symmetric MJO structure (Vintzileos, personal communication). On the other hand, enhanced stratiform heating was also observed via a radar network during DYNAMO (Powell and Houze 2013).

(iii) Intensification over the West Pacific and SPCZ

The split convective anomalies and the 200-hPa twin anticyclonic anomalies propagate eastward into the West Pacific in Phase VII, although the strength of the entire system appears weakened in the tropics apart from the SPCZ (Figs. 10 and 11). However, signs of tropical-extratropical interactions are seen in Phases VI and VII, especially in terms of 475-hPa [Q_1] and 200-hPa [(u^*, v^*)] around 30°N, 150°E–150°W (Figs. 10b and 11b). At the same time, lower-tropospheric moisture is advected northward from ~ 10°N between 120°E–150°W (Fig. 11a). A set of 200-hPa twin cyclonic anomalies originating over Africa
in Phase VI move to $\sim 60^\circ$E in Phase VII (Fig. 11b). At this stage, the Indian Ocean is dominated by anomalous low cloudiness, 475-hPa $[Q_1]$ cooling, 750-hPa dry $[\gamma^*]$, and 400-hPa cold $[T^*]$ (Figs. 10a,b and 11a,b). In Phase VIII, the convective systems over the SPCZ and the 200-hPa anomalous twin anticyclones centered around the date line are both intensified (Figs. 10b and 11b). From Phase VII to VIII, enhanced cloudiness and deep convective heating start to reemerge around $60^\circ$E, as low-tropospheric moisture is advected into the area again (Figs. 10 and 11a).

The vertical profiles on the equator show that anomalous descents, $[Q_1]$ cooling, and $[Q_2]$ moistening prevail over the equatorial Indo-Pacific warm pool region in Phases VII and VIII (Figs. 12a and 14a). Unlike the equatorial signals, the enhanced off-equatorial tropospheric ascents, $[Q_1]$ heating, and $[Q_2]$ drying propagate eastward in Phases VI and VII, then stall and intensify around the date line in Phase VIII (Figs. 12b and 14b). The anomalous descents are collocated with warm and dry anomalies in the low and mid troposphere (Figs. 12 and 13). Around the date line, 400-hPa warm and moist anomalies are seen on and (especially) off the equator. They intensify from Phase VII to VIII as the deep and stratiform convective heating is intensified in the SPCZ (Figs. 13 and 14b). From Phase VI to VIII, the anomalous circulation and thermodynamic fields gradually revert to a state similar to the initiation stage seen in Phases I and II at $0^\circ$–$60^\circ$E on and off the equator.

(iv) Dry phase in the Western Hemisphere

In Phase VIII, the 200-hPa anomalous anticyclones are located around the date line (Fig. 11b). They proceed eastward into the Western Hemisphere as seen in Phase I. In the absence of deep convective systems such as those over the Indo-Pacific warm pool, the anticyclones are replaced by a clockwise cross-equatorial anomalous circulation centered around $140^\circ$W in Phase III. This structure continues to propagate eastward, and is seen to interact with the circulation centered around $60^\circ$W over South America in Phases IV and V. These cross-equatorial circulations diminish as the quadrupole develops in the Eastern Hemispheric
warm-pool region in Phases VI and onward. The convective activity in the Eastern Pacific
ITCZ, America, and the tropical Atlantic also exhibit corresponding variability (Fig. 10).
The vertical profiles on the equator show that during the transition from the 200-hPa MJO
quadrupole to the cross-equatorial circulation, the 400-hPa warm and moist anomalies as-
associated with off-equatorial convective heating and drying over the East Pacific in Phases
I–III (Figs. 13 and 14) are replaced by warm and dry anomalies associated with anomalous
descending motions in Phases III-V (Figs. 12 and 13).

2) Predominantly antisymmetric MJO convective systems

(i) Initiation in the Indian ocean

This sequence of off-equatorial MJO convection initiation in Fig. 15 is distinctly different
from that of the predominantly symmetric events (Fig. 10). In particular, a transition takes
place over East Africa around 40°E in Phases I–III (Fig. 15a), in which suppressed convection
(positive \(T_B\) anomaly) at 15°S–0° becomes enhanced convection, while the surrounding
African continent remains in an enhanced convective state peaking in Phase II. The western
Indian Ocean experiences enhanced convection at about the same latitudinal range. The
associated heating over the ocean represented by 475-hPa \(Q_1\) (Fig. 15b) is originally a
cross-equatorial band over 40°–90°E in Phase I, stretching northeastward from the coast of
Africa to the Bay of Bengal. In Phase II, this enhanced heating band is confined to the
south of the equator at 40°–60°E, then moves eastward to 60°–90°E in Phase III, where it
merges with a broader heating region in the south Indian Ocean. The \(Q_1\) enhancement
over land is strongest in Phase II over equatorial and Southwest Africa. Overall, in Phases
I–III, the deep convective maxima initiating over the Indian Ocean make a southwestward
move to Africa, then move eastward returning to the Indian Ocean.

Unlike the predominantly symmetric events, there is a lack of significant low-tropospheric
moisture buildup in the Indian Ocean during antisymmetric Phases I–III. Even though the
spatial distribution of 750-hPa $[\gamma^*]$ at $0^\circ-90^\circ$E displayed in Fig. 16a resembles (albeit less symmetrically) that in Phases I and II of the predominantly symmetric events (Fig. 11a), the enhanced humidity does not move off Africa into the Indian Ocean in antisymmetric Phase III. The absence of a 200-hPa organized MJO quadrupole in that phase (Fig 16b) should therefore be of no surprise. Instead, an anticyclonic anomaly is seen around $30^\circ$N, $40^\circ$E, while a cyclonic anomaly is centered around $30^\circ$S, $90^\circ$E. Other than the anomalous cyclonic centers, the region $30^\circ$S–$30^\circ$N, $0^\circ–120^\circ$E exhibits weak 400-hPa warm $[T^*]$. This behavior is very different from the predominantly symmetric MJO convection, where an organized quadrupole is clearly established by Phase III (Fig 16b).

The vertical profiles on the equator indicate that the initial transitions of $[\omega^*]$ and $[T^*]$ over East Africa (Figs. 17a and 18a) are similar to those in the symmetric events (Figs. 12a and 13a). However, over the equatorial and off-equatorial Indian Ocean the anomalous ascents become weaker and more scattered as the major convective region shifts southward from Phases I to III (Fig. 17). There is still a westward-tilted anomalous circulation cell with prevailing low-tropospheric anomalous convergence along the equator at $60^\circ–90^\circ$E in Phases II and III (Fig. 17a). Similarly to the predominantly symmetric events, the anomalous convergence appears to lead low-tropospheric moist $[\gamma^*]$ at this stage (Figs. 17a and 18a), while warm and dry anomalies prevail in the mid-troposphere on and off the equator (Fig. 18). However, the area of moist $[\gamma^*]$ shrinks over time. The $[Q_1]$ and $[Q_2]$ profiles on the equator suggest that after Phase I, where a spectrum of shallow to deep convection exists at $60^\circ–120^\circ$E, shallow and congestus cloud types may prevail in the same region in Phases II and III (Fig. 19a).

(ii) Passage over the Maritime Continent

The 850-hPa wind anomalies in the equatorial western Indian Ocean turn from easterlies in Phase III to westerlies in Phase IV, while an MJO quadrupole similar to the predominantly symmetric case in Fig. 11 emerges at 200 hPa (Fig. 16). In Phase IV, off-equatorial deep
convection signals in \([T_B]\) and \([Q_1]\) are seen around 15°S–0° and 60°–120°E (Fig. 15). Low-tropospheric moist anomalies also establish over the Indian Ocean, with maxima around 15°S between 60°–120°E (Fig. 16a). The MJO quadrupole is lopsided, however, with much stronger 200-hPa anomalous flow and 400-hPa \([T^*]\) signals in the Northern Hemisphere. On the other hand, as the MJO propagates eastward in Phases IV–VI, the anticyclonic anomaly in the Southern Hemisphere interacts with the Australian Monsoon circulation, resulting an even more asymmetric quadrupole (Fig. 16b). Both features indicate that the occurrence of strong antisymmetric MJO convective systems tends to coincide with steep meridional temperature gradient in the Northern Hemisphere and active Southern Hemispheric monsoon systems.

In Phases IV to VI, the off-equatorial convective systems pass the Maritime Continent without inhibition. This is the most notable difference between predominantly symmetric and antisymmetric MJO convective systems. In the lower troposphere, the major anomalous convergence and moist \([\gamma^*]\) are centered in the region between Australia and the Southeast Asian Archipelago (Fig. 16a). Similarly to the predominantly symmetric events, the 200-hPa twin anticyclonic anomalies are energized, along with 400-hPa warm \([T^*]\) (Fig. 16b). In Phases V and VI, a 850-hPa westerly intrusion also occurs along the equator (Fig. 16a), but has minimal drying effects on the off-equatorial MJO convective systems to the south. When merging with the Australian Monsoon in Phase IV, this intrusion may draw air from the moist South Indian Ocean through the cyclonic anomalies situated around 15°S, 60°–120°E (Fig. 16a), rather than the dry Africa and Western Asia (Fig. 11a) in the Northern Hemisphere. Indeed, the 750-hPa dry \([\gamma^*]\) associated with the westerly intrusion does not penetrate as deeply through the Maritime Continent as in the case with predominantly symmetric convective systems (cf. Fig. 11a).

The vertical profiles on the equator show weak but definite westward-tilted anomalous circulation cells progressing eastward, similar to the predominantly symmetric events in Phases IV–VI (Figs. 12a and 17a). On the other hand, the off-equatorial anomalous circulation cells
are much stronger than those in the symmetric events, and propagate without diminishing strength across the Maritime Continent into the West Pacific (Figs. 12b and 17b). The relationship between anomalous ascents and moist anomalies is the same as in the symmetric events. The warm and moist anomalies, however, undergo enhancement on and off the equator (Fig. 18), as opposed to the reduction observed in the symmetric events (Fig. 13). The equatorial \([Q_1]\) and \([Q_2]\) vertical profiles have very weak signals, although some shallow to deep cloud-type transition can still be identified, especially in Phase IV. The equatorial dry intrusion confined to the west of the Maritime Continent manifests itself in Phase VI as a lower-tropospheric dry \([\gamma^*]\), as well as evaporative cooling and moistening in \([Q_1]\) and \([Q_2]\) (Fig. 18a and 19a). These features markedly contrast the extensive dry anomalies, cooling, and moistening in the symmetric events (Fig. 13a and 14a). The off-equatorial signals are stronger and spread wider than those in the symmetric events, and remain strong throughout this stage. The dominant cloud types here appear to be deep convection and stratiform anvils, with notable higher ratio of deep convection at 100°–120°E (Fig. 19b).

In summary, unlike predominantly symmetric systems, the production of eddy available potential energy via convection is uninterrupted when predominantly antisymmetric MJO convective systems pass over the Maritime Continent. Tapping into this available potential energy, the kinetic energy of the MJO can be generated continuously. Therefore, antisymmetric convective systems and the associated heating are favorable of strong MJO propagating throughout the Indo-Pacific warm pool.

(iii) Intensification over West Pacific and SPCZ

The deep convective anomalies in \([T_B]\) and 475-hPa \([Q_1]\) reach the West Pacific and the SPCZ in Phases VII and VIII (Fig. 15). In Phase VII, the 200-hPa anomalous anticyclones are centered at 15°S, 120°E, and 15°N, 150°E. Similarly to Phase III, there is no evidence of an MJO quadrupole. In Phase VIII, the 200-hPa quadrupole reemerges when a set of 200-hPa twin cyclonic anomalies is seen at 60°–90°E (Fig. 16b). The Indian Ocean is dominated
by anomalous low cloudiness, 475-hPa \([Q_1]\) cooling, 750-hPa dry \([\gamma^*]\), and 400-hPa cold \([T^*]\) at this stage (Figs. 15 and 16). Note that in Phase VIII the patterns of negative \([T_B]\) and positive \([Q_1]\) are roughly the same in the equatorial West Indian Ocean. Therefore, the cross-equatorial band of \([Q_1]\) heating in 40°–90°E is likely a genuine pattern, which becomes enhanced and more identifiable in Phase I. The vertical profiles show that, unlike predominantly symmetric events, the anomalous vertical ascents, warm and moist anomalies, and enhanced \([Q_1]\) and \([Q_2]\), continue into West Pacific both on and off the equator in Phase VII, where they are further enhanced in Phase VIII (Figs. 17–19).

5. Conclusions and future work

In this work, we have studied the predominantly symmetric and antisymmetric MJO convective systems in infrared brightness temperature \((T_B)\) satellite observations and kinematic and thermodynamic fields from reanalysis. Using nonlinear Laplacian spectral analysis (NLSA, Giannakis and Majda 2012c, 2013; Giannakis et al. 2012), a nonlinear manifold generalization of PCA, we decomposed the symmetric and antisymmetric averages of \(T_B\) over the 15°S–15°N equatorial belt into families of spatiotemporal modes for the period 1983–2006 sampled every 3 h. No preprocessing such as seasonal detrending or intraseasonal bandpass filtering was applied. As a result, the recovered modes provide a multiscale decomposition of the data into modes including ENSO, the seasonal cycle, intraseasonal convective waves, and diurnal-scale variability.

MJO modes, occurring with significant strength mostly in boreal winter to spring, were recovered in each of the symmetric and antisymmetric datasets. Through their associated temporal patterns (analogous to PCs), we created symmetric and antisymmetric MJO indices which were employed in turn to identify MJO events in the observation period with strong symmetric or antisymmetric components in the signals of deep convection. While these components may exist simultaneously, the majority (\(\sim 70\%\)) of strong MJO convective
systems were found to be either predominantly symmetric or predominantly antisymmetric.

To gain insight on the similarities and differences between the predominantly symmetric and antisymmetric MJO signals in deep convective systems, we built composites based on the identified events for a number of quantities of physical interest, including wind, temperature, humidity, and their derived heat and moisture budget residuals, after those fields were processed through nonlinear locally-smooth adaptive filtering (LSAF, Tung et al. 2011; Gao et al. 2011). The features revealed by the reconstructions establish the predominantly antisymmetric MJO convective systems as distinctly different entities from their symmetric counterparts.

First, the predominantly symmetric and antisymmetric MJO convective systems were found to differ drastically with respect to their propagation over the Southeast Asia-Australia Maritime Continent. Before reaching the Maritime Continent, symmetric MJO convective systems sustain the shallow–congestus to deep convection with stratiform cloud types near the equator. Off the equator, the most enhanced cloud type is likely stratiform clouds. Upon reaching the Maritime Continent, the convective systems split into two north-south branches separated by dry air intrusion over the equator. The mid-tropospheric conversion of perturbation available potential energy into kinetic energy due to convection is therefore interrupted over the path along the equator. These MJOs do not maintain their strength beyond this point, although some disturbances can still propagate northeastward and south-eastwards into the SPCZ. On the other hand, antisymmetric MJO convective systems often do not have a systematic cloud-type transition around the equator. They are more likely driven by off-equatorial deep convection with stratiform anvils, which are not interrupted when passing the Maritime Continent, and the energy conversion is maintained all the way to SPCZ. There is sign of dry intrusion confined over the Indian Ocean in antisymmetric events; the immunity to dry intrusion might be crucial for simulating strong MJO convective systems which traverse a long distance in the Indian Ocean–West Pacific warm pool region.

Another important distinction between the symmetric and antisymmetric MJO events
concerns their relation with ENSO and the diurnal cycle. During neutral ENSO and weak ENSO years, the symmetric MJO component is out of phase with the leading symmetric diurnal mode, while the antisymmetric MJO is in phase with the corresponding antisymmetric diurnal mode. The former relationship breaks down during strong El Niño events. Both relationships might break down during strong La Niña events.

Our phase reconstructions indicate that the first sign of MJO initiation may be over east Africa in both symmetric and antisymmetric cases. Due to the cyclic nature of the phase definition, however, what is observed may be the outcome of wavetrain initiation of the MJO. We plan to study the challenging question of intermittent MJO initiation and termination, as well as the physical mechanisms of the multiscale interactions between interannual, intraseasonal and diurnal modes in future work involving NLSA of 2D fields.

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REFERENCES


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Table 1. Number of CLAUS $T_B$ samples from 1984–2005 identified for each MJO phase as strongly symmetric, antisymmetric, or overlapping. The overlap samples are strong but asymmetric cases, hence identified as both symmetric and antisymmetric.

<table>
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<th>III</th>
<th>IV</th>
<th>V</th>
<th>VI</th>
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<td>930</td>
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<td>431</td>
<td>350</td>
<td>280</td>
<td>266</td>
<td>341</td>
<td>2743*</td>
</tr>
</tbody>
</table>

*Without the phase partition, there are 4173 temporally overlapping samples.
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1. Time-longitude section of (a) symmetric and (b) antisymmetric brightness temperature $T_B$ data (in K) from CLAUS for the period 1992–1993. Only thresholded values are shown to emphasize convective activity. The bottom map in (a) indicates that the symmetric component was obtained via averaging over $15^\circ$S to $15^\circ$N. The antisymmetric component in (b) was obtained by subtracting the values at the northern latitudes from the corresponding southern latitudes.

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Same as Fig. 6, but for (a) symmetric MJO, (b) antisymmetric MJO, and (c) symmetric Maritime Continent modes.

Phase space diagrams for 1992–1993. (a) Symmetric MJO pair (MJO1, MJO2); (b) antisymmetric MJO pair (MJO1<sub>A</sub> and MJO2<sub>A</sub>); (c) symmetric diurnal pair (Diurnal1 and Diurnal2); (d) antisymmetric diurnal pair (Diurnal1<sub>A</sub> and Diurnal2<sub>A</sub>). The green crosses (X) mark the weaker MJO event observed during TOGA COARE from mid-November 1992 to early January 1993. The red dots (O) mark the later, stronger event terminating in mid-February 1993. Roman numerics in (a) and (b) denote the eight MJO phases, calibrated against the TOGA-COARE events so that they correspond to a sequence of enhanced eastward-propagating convective activity from the Eastern to Western Hemispheres.

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Same as Fig. 10, but for (a) 850-hPa winds (vectors, m s\(^{-1}\)) and 750-hPa specific humidity (color shades, g kg\(^{-1}\)), and (b) 200-hPa winds and 400-hPa temperature (K). Note that the fields are transient eddies, i.e., with zonal mean and 1984–2005 temporal mean removed prior to band-pass filtering.

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