The Linear Response of ENSO to the Madden–Julian Oscillation

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(Manuscript received 29 March 2004, in final form 21 October 2004)

ABSTRACT

The possibility that the tropical Pacific coupled system linearly amplifies perturbations produced by the Madden–Julian oscillation (MJO) is explored. This requires an estimate of the low-frequency tail of the MJO. Using 23 yr of NCEP–NCAR reanalyses of surface wind and Reynolds SST, we show that the spatial structure that dominates the intraseasonal band (i.e., the MJO) also dominates the low-frequency band once the anomalies directly related to ENSO have been removed. This low-frequency contribution of the intraseasonal variability is not included in most ENSO coupled models used to date. Its effect in a coupled model of intermediate complexity has, therefore, been studied. It is found that this “MJO forcing” ($\tau_{\text{MJO}}$) can explain a large fraction of the interannual variability in an asymptotically stable version of the model. This interaction is achieved via linear dynamics. That is, it is the cumulative effect of individual events that maintains ENSOs in this model. The largest coupled wind anomalies are initiated after a sequence of several downwelling Kelvin waves of the same sign have been forced by $\tau_{\text{MJO}}$. The cumulative effect of the forced Kelvin waves is to persist the (small) SST anomalies in the eastern Pacific just enough for the coupled ocean–atmosphere dynamics to amplify the anomalies into a mature ENSO event. Even though $\tau_{\text{MJO}}$ explains just a small fraction of the energy contained in the stress not associated with ENSO, a large fraction of the modeled ENSO variability is excited by this forcing. The characteristics that make $\tau_{\text{MJO}}$ an optimal stochastic forcing for the model are discussed. The large zonal extent is an important factor that differentiates the MJO from other sources of stochastic forcing.

1. Introduction

The very important notion of the two-scale separation between weather and climate introduced by Hasselmann (1976) provided us with a paradigm to study and understand El Niño–Southern Oscillation (ENSO) variability. The basic assumption in Hasselmann’s hypothesis is that the state of the atmosphere–ocean–land–cryosphere system $z$ can be divided into two sub-systems $z = (f, s)$, where the fast ($f$) and slow ($s$) components affect each other but are characterized by two very different response times. Time averaging of the equations describing the slow component gives then a deterministic system of equations that is decoupled from the fast component. For this reason, these models cannot simulate one very important process in ocean dynamics: the accumulation of fast time-scale random impulses of momentum and heat within the ocean mixed layer. As suggested by Hasselmann in the climate context, a natural extension can be obtained by including a random forcing term representing the fast component in the dynamical equations. A review of this approximation as applied to ENSO is discussed by Penland (1996). In this context, the slow component represents the coupled ocean–atmosphere variability with time scales of the order of seasons to years, whereas the fast component represents the high-frequency variabili-
ity with characteristic subseasonal time scales. Since the decorrelation times of the high-frequency events are much smaller than the time scales associated with ENSO, this kind of variability is often referred to as stochastic forcing (SF).

The use of stochastic models of ENSO, introduced by Lau (1985), has proved to be a possible explanation for several observed characteristics of ENSO such as its irregularity (e.g., Penland and Sardeshmukh 1995; Blanke et al. 1997; Moore and Kleeman 1999b; Thompson and Battisti 2000; Roulston and Neelin 2000), its predictability (e.g., Eckert and Latif 1997; Kleeman and Moore 1997; Zavala-Garay et al. 2004), some characteristics of its probability density functions (e.g., Penland et al. 2000), and its decadal variability (e.g., Moore and Kleeman 1999a). While all these characteristics are probably affected by several other factors, it has become clear that inclusion of the fast variability adds an important component of realism to the ENSO models.

The SF in the Tropics is composed of westerly wind bursts (WWBs), the Madden–Julian oscillation (MJO; Madden and Julian 1971), cyclones associated with midlatitude cold surges (Yu and Rienecker 1998), atmospheric waves, etc.

Strong SF activity in the western Pacific has been observed to precede ENSO events (McPhaden et al. 1998; Webster and Palmer 1997) leading to the generation of downwelling Kelvin waves that could potentially trigger or modify ENSO events (Fedorov 2002; Zavala-Garay et al. 2003, hereafter ZG03). However, the specific connection between the MJO and ENSO is a topic of current debate. Global MJO indices based on the first two principal components (PCs) of intraseasonal variability (IV) have been shown to be uncorrelated with ENSO (Slingo et al. 1999; Hendon et al. 1999). However, consideration of the third PC, which describes the meandering of the MJO around the warm pool edge, has been shown to solve the apparent lack of correlation between these two events (Kessler 2001). In an early work, Lau and Chan (1986) suggested that MJO and ENSO should be considered as strongly linked phenomena. Indeed, interannual anomalies in seasonal variance of the MJO forcing of oceanic Kelvin waves have been shown to lead ENSO events by several months (Zhang and Gottschalck 2002).

The coupled ocean–atmosphere system describing ENSO can amplify MJO perturbations via nonlinear and linear interactions. For example, Kessler and Kleeman (2000) found that an idealized MJO forcing can produce changes in the thermal structure of the oceanic mixed layer that could potentially modify the evolution of ENSO. In this scenario, the thermal changes are the result of nonlinear interactions that are therefore rectified to low frequencies by the coupled system. The second possibility, that the coupled system linearly amplifies MJO perturbations, is motivated by the fact that the most disruptive forcing structures of empirical (e.g., Penland and Sardeshmukh 1995) and numerical linear models (e.g., Moore and Kleeman 1999a) tend to be concentrated in the western Pacific where strong MJO surface forcing is observed. In this case, it is the low-frequency tail of the MJO that is relevant (Roulston and Neelin 2000; ZG03), with nonnormal processes playing an important role (Penland and Sardeshmukh 1995; Moore and Kleeman 1999b).

All these studies suggest that the MJO, as a component of the SF, could be an important source of ENSO variability. However, a systematic study on the impact of the MJO as differentiated from other sources of SF has not been provided. Several questions remain to be answered about this scenario. Is the MJO a source of optimal stochastic forcing for ENSO? If so, how much of the observed interannual variance can be explained by this forcing? How are the MJO perturbations linearly amplified by the coupled system? What characteristics differentiate the MJO from other sources of SF?

In this work we concentrate on the possible linear effects of the MJO on ENSO. This requires an estimate of the low-frequency tail of the MJO. For this reason, in section 2 we represent the equatorial surface zonal wind pseudostress as the sum of different contributions with special emphasis on the MJO and its low-frequency tail. In section 3 we study and discuss the impact of these estimates using an ENSO model of intermediate complexity as a first approximation to the true coupled ocean–atmosphere system. Finally, in section 4 we summarize and discuss our findings.

2. Identification of the different contributions of surface wind stress

a. Data and methods

Surface zonal winds from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses (Kalnay et al. 1996) and weekly SST analyses (Reynolds and Smith 1994) for the period 1980–2003 were used. The spatial domains considered are different for these two datasets. For zonal wind, the domain considered is the global equatorial strip (15°S–15°N), which encompasses most of the convective variability of the MJO (Salby and Hendon 1994), while for SST we concentrate on the equatorial Pacific (20°S–20°N, 112°E–85°W), which embodies most of the ENSO SST anomalies. The spatial resolution of the NCEP–NCAR dataset is 2.5°, and
that of the Reynolds SST is 1°. The SST was interpolated in time to daily values for compatibility with the atmospheric data. A climatological annual cycle for each calendar day was computed and smoothed with a 90-day running mean to stabilize and avoid aliasing problems (Madden and Jones 2001). The resulting climatology was well approximated by the first three harmonics of the annual cycle and was then subtracted from the daily data to yield the anomalies.

The NCEP zonal stress anomalies are less noisy than those of the Florida State University (FSU) subjective analysis and weaker than FSU and other newer and more reliable stress estimates (Wittenberg 2004). Conceding this limitation, this product provides gridded data at high temporal resolution, allowing the study of scale interactions such as that reported here. The use of other reanalysis products definitely deserves future investigation.

We analyze the anomalies using standard signal detection techniques such as singular value decomposition (SVD; section 2b) and Hilbert SVD (HSVD; section 2c). In each case the rule of thumb proposed by North et al. (1982) is a suitable criterion to determine whether a given eigenmode has been well resolved by the dataset. The number of degrees of freedom (DOF) used in the SVD analysis was 10, corresponding to the number of warm and cold ENSO events in the 1980–2003 period. For the HSVD analysis, DOF was estimated as the length of the daily time series divided by 70 days, a typical period for the MJO. Only those modes that clearly exceeded the North criteria (North et al. 1982) were considered.

The spatial and temporal scales of the recovered signals were analyzed with the aid of two-dimensional Fourier transforms of the latitudinal mean \(15°S–15°N\) of the fields. The resulting frequency–wavenumber power spectral densities have a nominal bandwidth of 0.0001 cycles per day. Statistical stability is assured with 30 passes of a 1–2–1 frequency convolution, resulting in an effective bandwidth of 0.001 cycles per day.

b. Coupled ENSO contribution

It has been shown that the interannual anomalies in seasonal variance of Kelvin wave forcing by the MJO lead ENSO by several months but are not correlated at zero lag (Zhang and Gottschalck 2002). The following procedure removes the part of the wind that can be linearly related to the contemporaneous SST. This is a reasonable assumption since the atmospheric adjustment time in the Tropics is short (a few days) compared to the slowly evolving SST. This is the rationale for the use of steady atmospheric models (both statistical and dynamical) that are “slaved” to the ocean in several ENSO coupled models.

Some coupled models that use empirical atmospheres are based on regressions between observed stress and SST anomalies. However, in the present case, we want to remove from the observations the part that the coupled model will reproduce by itself. Since the coupled model uses a Gill-type dynamical atmosphere, our regression is based on wind versus SST anomalies. We loosely refer to this variability as the ENSO contribution in the sense that this is the part reproduced by most ENSO coupled models used to date. Similarly, the residual is referred to as the “non-ENSO” contribution.

To isolate the ENSO contribution, we constructed a statistical model based on the SVD of zonal wind and SST anomalies. It was found that the covariance of these two fields is dominated by two modes of variability (85% of squared covariance) that are well resolved by the dataset according to the North criteria (North et al. 1982). The projection coefficients of the first two SST modes were then used as predictors for the zonal wind anomalies. The non-ENSO contribution is then defined as the residual of this regression.

We define the total power at ENSO frequencies as the integral of the spectral density in the window with periods of 3–7 yr. Figures 1a–c show the power at ENSO frequencies for the observed zonal wind anomalies, those recovered by the empirical model, and the residual, respectively. The empirical model explains most of the observed interannual anomalies in the equatorial Pacific but not all. The residual (one order of magnitude smaller) is located primarily in the western Pacific and eastern Indian Ocean. ZG03 have shown that a similar residual field [constructed from the domain \(30°S–30°N, 124°E–80°W\) and considering both zonal and meridional winds] can produce realistic ENSO variability in an asymptotically stable coupled model. This variability is mainly excited by stochastically induced Kelvin waves. We will show that similar results were found for the residual field constructed here, since the Kelvin waves are mainly forced by the zonal wind in the \(5°S–5°N\) region. Therefore, the small residual shown in Fig. 1c can produce SST anomalies similar to those observed and therefore winds similar to those summarized in Fig. 1b. In the next section we show that a large fraction of this low-frequency variability has the same structure as that of the MJO.

c. MJO contribution

From the resulting residual wind, denoted as \(u_{\text{res}}\), the non-ENSO surface forcing was defined as \(\tau_n = u_{\text{res}}^2|u_{\text{res}}|\) since this is that part of the stress not produced by most
ENSO models. The alternative definition, $\tau_n = \tau - u_E[u_E]$, where $u_E$ is that part predicted by the empirical model, was found to contain a much stronger low-frequency tail. The definition used here has a low frequency with most of its energy over the eastern Indian Ocean and western Pacific, where MJO dominates (see Fig. 1c). In contrast, the amount of low-frequency variance found by using the second definition (not shown) was about twice as large and had large amplitudes over the central Pacific, where ENSO signal dominates.

1) FIRST ESTIMATE: MJO AT INTRASEASONAL FREQUENCIES ($\tau'_{\text{MJO}}$)

Following previous works to isolate the MJO, our first estimate is based on intraseasonally filtered data. We use a Lanczos filter with 49 weights and half-power cutoff frequencies of 30 and 90 days $^{-1}$. To emphasize the large scale of the MJO, a spatial four-point (two point) running mean was applied in the zonal (meridional) direction to $\tau_n$. This implies an effective spatial resolution of $10^\circ$ and $5^\circ$ in the zonal and meridional directions, respectively. We refer to these space- and time-filtered anomalies as $\tau'_{\text{MJO}}$.

The MJO is identified by pairs of modes whose projection coefficients are temporally in quadrature. One way of finding these modes is via HSVD. This method has been proven to be very efficient in isolating the eastward-propagating variability of the MJO (Zhang and Hendon 1997). For a detailed description of the method, see Barnett (1983). An HSVD analysis of $\tau'_{\text{MJO}}$ leads to four statistically significant Hilbert singular vectors (HSVs) according to the criterion of North et al. (1982), with the leading HSV (30% of the variance) being clearly separated from the rest. Figure 2a shows the meridional average ($15^\circ$N–$15^\circ$S) of the real part (thin dashed line) and imaginary part (thin continuous line) of the first HSV of $\tau'_{\text{MJO}}$ [denoted as $\phi_1(\tau'_{\text{MJO}})$].

The spatial structure of the first pair shows the well-known global structure of the MJO that has been identified in other variables such as outgoing longwave radiation (OLR; e.g., Fig. 2 of Kessler 2001) and 200-mb velocity potential (e.g., Fig. 1 of von Storch and Xu 1990). Local maxima can be observed over the Eastern Hemisphere, where strong coupling with convection occurs. In the Western Hemisphere a weaker, yet significant amplitude is observed as a result of a Kelvin-like “radiating response,” which propagates away from the convective anomaly. The result is a dynamical anomaly that is concentrated at wavenumber 1 (Hendon and Salby 1994).

The eastward extension of the MJO pointed out by Kessler (2001) was found to be represented by the second and third HSVs for the spatial domain ($15^\circ$S–$15^\circ$N). When a narrower domain confined to near the equator was used (e.g., $5^\circ$S–$5^\circ$N), the eastward extension was explained just by the second HSV. We found that this eastward extension had an important contribution to the low-frequency tail of $\tau_n$ (see next section), and therefore our estimates of MJO are based on the
first three HSVs. The meridional average (15°N–15°S) of the amplitudes of the second and third HSVs of \( \tau_n \) are shown in Fig. 2b (continuous and dashed thin lines, respectively).

The MJO estimate, which optimally accounts for traveling features in \( \tau_n \), is given by (Barnett 1983):

\[
\tau_{MJO}(x, y, t) = \text{Re} \left[ \sum_{i=1}^{3} a_i(t) \phi_i(x, y) \right]
\]

with

\[
a_i = \mathcal{H} \mathcal{A} (\tau_n^i),
\]

where \( \mathcal{H} \mathcal{A}(\tau_n^i) \) is the Hilbert transform of \( \tau_n^i \), \( \phi_i \) denotes the complex HSVs, and \( \mathcal{A} \) denotes the transpose of the complex conjugate. An example of the recovered signal is shown in Fig. 3a for the period January 1987–December 1992. A clear MJO signal is apparent, characterized by large-scale disturbances propagating eastward at speeds of about 5 m s\(^{-1}\) from the Indian Ocean to the western Pacific, and with its strongest activity during boreal winter (Zhang and Dong 2004). This estimate does not have any low-frequency component and therefore cannot be used to test the hypotheses proposed by linear stability analysis. We show this estimate, however, to compare and better understand the linear response to our next estimate, which also considers MJO contributions at low frequencies.

2) SECOND ESTIMATE: MJO AT ALL FREQUENCIES (\( \tau_{MJO} \))

We repeated the methodology of the previous section but this time retaining the low frequency of \( \tau_n \). That is, \( \tau_n \) was just low-pass filtered for frequencies lower than 30 days\(^{-1}\). In addition, the same space filter used in the definition of \( \tau_n^i \) was applied (see previous section). We will refer to this space- and time-filtered pseudostress as \( \tau_n^f \) to emphasize that frequencies lower than those of the MJO have been retained. Similar results were obtained without doing any space and time filtering. However, this filtering naturally removes the short lived WWBs frequently observed in the equatorial Pacific (e.g., Vecchi and Harrison 2000). The eigen values and associated patterns (HSV) were very similar to those obtained with the analysis of the intraseasonal anomalies. Figure 2a shows that the meridional mean (15°N–15°S) of the leading HSV of \( \tau_n^f \) [denoted as \( \phi_1(\tau_n^f) \), explaining 26% of the variance] is almost identical to that of our previous estimate \( \{\phi_1(\tau_n^i)\} \). The same is observed for the second and third HSVs (Fig. 2b). The spatial correlation between both sets of HSVs was greater than 0.92 for the three HSVs used.

The estimated MJO is given then by (1), with \( \tau_n^f \) replaced by \( \tau_n^i \). An example of the recovered signal, \( \tau_{MJO} \), is shown in Fig. 3b. It appears to be very similar to \( \tau_{MJO}^i \) (Fig. 3a), the main difference being that some westerly events last longer (e.g., in 1992). Similar characteristics can be observed for other periods (see Figs. 8 and 9 for other examples). Comparison with \( \tau_n \) revealed that most of the “gaps” separating westerly wind anomalies in \( \tau_{MJO} \) (Fig. 3a) are artificial features enforced by the filtering process. It will be shown in section 3 how this filtering has enormous consequences for the intermediate model response.

A space–time spectral analysis highlights the differences between the ENSO contribution and that of our two estimates of the MJO. Figure 4 shows the space–time spectra of the meridional mean of \( \tau_n \), \( \tau_{ENSO} \), \( \tau_{MJO}^i \), and \( \tau_{MJO} \). Positive frequencies correspond to eastward propagation whereas negative frequencies correspond to westward propagation. As shown in Fig. 1, the empirical model effectively removes the low-frequency stresses in the central Pacific associated with ENSO, reflected as energy concentrated around wavenumber 2 in Fig. 4b. Indeed, when the ocean component of the intermediate model used here was forced with \( \tau_{ENSO} \), SST anomalies very similar to those observed developed.
The space–time spectra for $\tau_{MJO}$ (Fig. 4c) show energy concentrated in the vicinity of wavenumber 1 for the eastward-propagating variance. Similar space–time spectra have been reported previously by other authors for the intraseasonal 850-mb zonal wind (e.g., Salby and Hendon 1994; Zhang and Hendon 1997). As noted by these authors, this is in contrast with the MJO signal in convection (as depicted by OLR), which tends to be
spread over wavenumbers 1–3. The preference of the dynamical component of the MJO to be more zonally extensive than that of OLR (Hendon and Salby 1994) made the separation between ENSO and MJO more natural. Indeed, the spectra of $\tau_{\text{MJO}}$ (Fig. 4d) show a clear deviation of their energy at low frequencies from that directly forced by equatorial Pacific SSTs (Fig. 4b). Here, $\tau_{\text{MJO}}$ mainly contributes to wavenumber 1.
whereas $\tau_{\text{ENSO}}$ mainly contributes to wavenumber 2. This is a direct consequence of the fact that ENSO anomalies are mainly concentrated in the central Pacific (e.g., Fig. 1a) whereas MJO anomalies tend to be of global scale (i.e., wavenumber 1; e.g., Fig. 6a).

Readers familiar with the MJO may wonder why we have included the second and third HSVs in our estimates. The methodology used effectively isolates the eastward-propagating variance, and it was found that these HSVs describe aspects of the MJO, such as its eastward extension during ENSO years, and some aspects of its irregularity in shape and duration. As such they are important contributors to the low-frequency variability of the MJO. Figure 5 exemplifies these aspects. Figure 5a shows the meridional mean of the $\tau_{\text{MJO}}$ recovered when just the leading HSV is considered (shades) along with its 90-day running mean (contours). The period shown is from January 1997 to May 1998, and vertical lines indicating the zonal boundaries of the Pacific Ocean and the climatological warm-pool edge (180°) are shown for clarity. The recovered MJO does have a low-frequency component concentrated west of the date line, as depicted by the 90-day running mean. Figure 5b shows the same plot for the $\tau_{\text{MJO}}$ recovered when the second and third HSVs are considered. It can be seen that consideration of these HSVs increases the magnitude of the low-frequency tail in the estimate. The eastward extension during the 1997/98 ENSO event is also apparent. The association of the second and third HSVs with the first HSV was also observed for other periods. This fact is evident in Figs. 5c and 5d, where the seasonal variance (defined as the 3-month running variance) of the amplitudes of the second and third HSVs are compared to that of the first HSV. The correlation between the seasonal variance of Figs. 5c and 5d is 0.5 (the same in the two cases), which is significant at the 95% level. For clarity, vertical dotted lines are drawn at the beginning of each year. The seasonality of the MJO is evident as is the fact that the second and third HSVs tend to occur during periods of increased MJO activity. The same behavior was observed for the amplitudes of the estimate presented in the previous section.

d. Non-MJO contribution

We define the non-MJO contribution as $\tau_{\text{NMJO}} = \tau_n - \tau_{\text{MJO}}$. This roughly includes all the westward-propagating and steady variance, eastward-propagating variance with time scales shorter than 30 days, and all the variance with wavenumbers larger than 2. The space–time spectra of $\tau_{\text{NMJO}}$ (not shown) has no preferred mode of variability with a small preference at low frequencies for wavenumbers higher than 2. Figure 6 compares the variance for $\tau_{\text{MJO}}$ and $\tau_{\text{NMJO}}$. Most of the energy in $\tau_n$ is contained in $\tau_{\text{NMJO}}$, and increases poleward (Fig. 6b). The $\tau_{\text{MJO}}$ estimate explains less than half of the variance of $\tau_n$ in the eastern Indian Ocean–western Pacific sector. Along the equator, $\tau_{\text{MJO}}$ explains about 30% of $\tau_n$ in the same sector (Fig. 6c).

3. The response of an intermediate model to the MJO estimates

a. The model

The simple intermediate coupled model used here (Kleeman 1993) has skill predicting ENSO SST anomalies comparable to coupled GCMs and other intermediate models [e.g., the Zebiak and Cane (1987) model]. It computes the interannual atmospheric and oceanic anomalies about a prescribed annual cycle. The use of this intermediate model was motivated by the feasibility of performing several sensitivity experiments while considering most of the physics that is thought to be fundamental for ENSO. In particular, it is used here to demonstrate and study in great detail the linear effects of the MJO estimate on ENSO. A detailed description of the coupled model can be found in Kleeman (1991, 1993).

The atmospheric component of the model is a steady, two-pressure-level model (250 and 750 hPa) on a $\beta$-plane as described in Webster (1972). The model is forced by latent and direct thermal heating anomalies, with the former dominating in the western Pacific. The model was shown by Kleeman (1991) to give a reasonable description of the tropical precipitation and circulation anomalies observed during various El Niño and La Niña events when observed SST anomalies were prescribed.

The ocean component of the coupled model solves the linear shallow water equations for a single baroclinic mode on an equatorial $\beta$-plane using the long-wave approximation. The dynamical variables are approximated as the sum of Kelvin waves and the first five equatorial long Rossby waves (e.g., Moore and Philander 1977). The model domain is limited to the Tropical Pacific between 124°E and 80°W. The phase speed of the first baroclinic mode was chosen to be $c_p = 2.4$ m s$^{-1}$, which is consistent with the observed phase speed of Kelvin waves forced by intraseasonal variability (Kessler et al. 1995).

The coupled model possesses two physically important nonlinearities in the form of thresholds. The first relates to latent heat release by deep penetrative convection, which occurs when the moist static energy of
air parcels exceeds a critical value corresponding to an SST of approximately 28°C (see Kleeman 1991). The second nonlinearity is a threshold on the effects of the thermocline anomalies on SST that prevents runaway coupled instability. When the thermocline is very deep or very shallow, further changes of thermocline depth do not influence SST (see Kleeman 1993).

In the stochastically forced experiments to be de-

Fig. 5. Meridional mean (15°S–15°N) of $\tau_{\text{MJO}}$ recovered (a) from the first HSV and (b) from the first three HSVs. Shades show the daily pseudostress (m$^2$s$^{-2}$), and contours show the 90-day running mean with a contour interval of 2 m$^2$s$^{-2}$ starting from 2 m$^2$s$^{-2}$. The boundaries of the Pacific Ocean and the date line are indicated with continuous and dashed vertical lines, respectively. (c) Seasonal variance of the amplitude of the first and second HSV. (d) Seasonal variance of the amplitude of the first and third HSVs. In (c) and (d), the correlation between the time series of each case is shown in the upper-left corner, and the dotted vertical lines indicate 1 Jan of each year.
scribed next, the SST anomalies resulting from the ocean model are used to compute the heating anomalies that drive the atmospheric model. The resulting surface wind anomalies are converted to surface wind stress anomalies using a linear stress law with a coupling coefficient of $6.8 \text{ m s}^{-1}$ and then augmented by the prescribed stochastic forcing. These combined stress anomalies are then used to drive the ocean model. In the experiments referred to as “stand-alone ocean model,” no atmospheric model is used, and therefore the ocean forcing is just the externally imposed stochastic variability.

**b. Response**

To incorporate the different surface momentum forcings into the coupled model, the stress estimates in the Pacific are extrapolated poleward of 15° applying an $e$-folding scale of 10° to cover the domain of the ocean model (30°S–30°N). The use of different $e$-folding lengths did not change the basic results. For the set of parameters chosen, the model is asymptotically stable with a decay time of about 3 yr, and all runs were started from a state of rest and null SST anomalies (SSTA; i.e., no dynamical and thermal anomalies about the prescribed annual cycle). Therefore all the interannual variability shown in our results is induced by the imposed external forcing. The pseudostress data were converted to stress assuming a drag coefficient $C_D$ of $1.5 \times 10^{-3}$.

Figure 7 shows the coupled model response to $\tau_n$, $\tau_{\text{MJO}}$, and $\tau_{\text{NMJO}}$. Several interesting conclusions can be derived from these experiments. The observed noise estimate can produce and maintain realistic ENSO variability (Fig. 7a) in terms of its amplitude, spectra, and irregular occurrence (e.g., ZG03). An ocean stand-alone model run forced by $\tau_n$ did not develop such large anomalies. Therefore most of this variability is amplified by coupled processes.

A new contribution from this work is that most of that variability is excited by our $\tau_{\text{MJO}}$ estimate (Fig. 7b), which is just a small fraction of $\tau_n$ (Fig. 6). A run forced by the 90-day running mean of $\tau_{\text{MJO}}$ revealed that it is the low-frequency component of $\tau_{\text{MJO}}$ that forces most of the variance shown in Fig. 7b. The low-frequency component of $\tau_{\text{MJO}}$ does not account for all of the low-frequency variability contained in $\tau_n$, but rather that of largest zonal extent (i.e., wavenumber 1). Disturbances with smaller scales (i.e., wavenumbers larger than 2)
Fig. 7. Intermediate model response to different stochastic forcings: (a) $\tau_p$, (b) $\tau_{MJO}$, and (c) $\tau_{NMJO}$. Fields shown are time–longitude sections of SSTA along the equator in °C.
are contained in $\tau_{\text{MJO}}$ and can also produce some variability, although small, in this model (Fig. 7c). The largest anomalies produced by $\tau_{\text{MJO}}$ develop in the 1982/83 warm event but are also present in other years. Thus other sources of SF besides the MJO also contribute to the SST variability, and the influence of the MJO varies from event to event. It can also be seen that the response is highly linear most of the time: the response to $\tau_n$ nearly equals to the sum of the responses to $\tau_{\text{MJO}}$ and $\tau_{\text{NMJO}}$. The model nonlinearity is however also present, especially around 1996. A model integration forced by SF from the last 50 yr shows that most of the time the system can be described by the corresponding tangent linear model (ZG03). Since the MJO produces the largest variability, we analyze next in more detail the characteristics that make the $\tau_{\text{MJO}}$ estimate an optimal stochastic forcing for this model.

1) Importance of the low-frequency tail of $\tau_{\text{MJO}}$

The way the model amplifies the $\tau_{\text{MJO}}$ perturbations is fully appreciated by analyzing how the thermocline anomalies in the ocean feedback via atmosphere-ocean interactions. An example of the mechanism is shown in Fig. 8. The $\tau_{\text{MJO}}$ forcing for the period January 1997–June 2003 is shown in Fig. 8a (shades) along with its 90-day running mean (contours). To emphasize the large scale of the external forcing, all the longitudes (0°–360°) are shown, with the zonal domain of the ocean model and the climatological warm-pool edge (180°) indicated with dashed and dotted lines, respectively.

We first show in Fig. 8b how the ocean responds to the MJO forcing (i.e., no coupling with the atmosphere was allowed in this run). It can be seen that the effect of $\tau_{\text{MJO}}$ on the ocean component is to force Kelvin waves in the western Pacific of either sign. These intraseasonal oceanic Kelvin waves forced by the MJO have been observed to produce SST anomalies in the eastern Pacific that tend to modify the zonal SST gradient (Zhang 2001). The consequent weakening or strengthening of the climatological trade winds will produce additional thermocline and SST perturbations that feed back to the atmosphere. Figure 8c shows these coupled zonal wind anomalies (contours) along with the SST anomalies resulting from both the externally imposed $\tau_{\text{MJO}}$ and the internally produced coupled contributions. It can be seen that the largest coupled wind anomalies are initiated after a sequence of several Kelvin waves of the same sign have been forced by $\tau_{\text{MJO}}$ as evidenced by its 90-day running mean. The cumulative effect of the forced Kelvin waves is to persist the (small) SST anomalies in the eastern Pacific just enough for the coupled mechanism to develop into a mature ENSO. This can be corroborated by subjecting the model to $\tau_{\text{MJO}}'$ forcing, which is shown in Fig. 9. Consider the cases highlighted above, where the sequence of events is such that wind anomalies of the same sign persist for some period of time. A consequence of the filtering process implied in the estimation of $\tau_{\text{MJO}}'$ is that such sequences of persistent wind anomalies are broken down into easterly and westerly anomalies (cf. Figs. 8a and 9a). The enforced alternation of downwelling and upwelling Kelvin waves (Fig. 9b) does produce coupled wind anomalies (evident in the line contours around the date line in Fig. 9c), but the alternation of upwelling/downwelling Kelvin waves tends to cancel out any growing SST anomalies (the contour and color scale in Fig. 9c is one-fifth of that in Fig. 8c).

2) Importance of the spatial structure of $\tau_{\text{MJO}}$

Figure 7 shows that $\tau_{\text{NMJO}}$, which is mainly composed of variability with wavenumbers larger than 2, does not produce much ENSO variability. As mentioned before, not all the low-frequency variability in the residual ($\tau_n$) is contained in the $\tau_{\text{MJO}}$ estimate, only that of large zonal extent. Figure 10 compares the power at ENSO frequencies along the equator (5°S–5°N latitudinal mean) for $\tau_{\text{MJO}}$ and $\tau_{\text{NMJO}}$. Both pseudostresses have comparable power at low frequencies, and similar results were obtained when including other latitudinal bands. Therefore, the large zonal extent of the MJO seems to be optimal for forcing ENSOs in the model.

Using the ideas of generalized stability theory introduced by Farrell and Ioannou (1996a,b), ZG03 have shown that about 71% of the expected variance in this model can be explained by considering just two EOFs of stochastic forcing in the equatorial Pacific. These two EOFs account for just 28% of the $\tau_n$ variance in the equatorial Pacific. Because their analysis of stochastic forcing is limited to the equatorial Pacific, which is the domain of the model, it is possible to quantify the expected variance in the model solution by a weighted sum of the projections of the EOFs of stochastic forcing onto the so-called stochastic optimals. These stochastic optimals are the spatial patterns that the stochastic forcing must possess in order to maximize the stochastically induced variance in the model. However, it was difficult from their analysis to make a clear connection with the MJO since the spatial domain used in that study was limited to the equatorial Pacific.

As noted in the previous section, most of the model
Fig. 8. (a) Time–longitude section of the meridional mean (15°S–15°N) of $\tau_{MJO}$ for the period Jan 1997–Jun 2003. Shades show the daily anomalies, and the contours show its 90-day running mean. The boundaries of the ocean model and the date line are indicated with dashed and dotted lines, respectively. (b) Stand-alone ocean equatorial thermocline depth (Z20) forced by $\tau_{MJO}$, and (c) coupled model SST (shades) and zonal stress anomalies (contours) forced by $\tau_{MJO}$. Units are (a) m$^2$ s$^{-2}$, (b) m, and (c) m$^2$ s$^{-2}$ and °C.
Fig. 9. Same as in Fig. 7, but for $\tau_{MJO}$. 
variability is excited by the cumulative effect of stochastically induced Kelvin waves. Stochastically induced Rossby waves are also generated, but their amplitude is much smaller, having then a small contribution to the thermocline variability. A detailed description of the covariance structure of the forcing of equatorial ocean waves by SF can be found in ZG03.

At a fixed longitude, the amplitude of the forced Kelvin wave \( K \) is proportional to the projection of the given stress onto the oceanic Kelvin wave structure, that is,

\[
K(x) = \int_{-15^\circ}^{15^\circ} \sigma e^{-1/2(y/R)^2} dy,
\]

(2)

where \( R = (c/\beta)^{1/2} \) is the equatorial Rossby radius (about 3° for \( c = 2.4 \text{ m s}^{-1} \)), and \( y \) is latitude. The zonal structure of \( K \) for the first two stochastic optimals
(\(K_{\text{opt}}\)) of this model is shown in ZG03. The leading stochastic optimal (which explains 60% of the stochastically forced variance) is of large zonal extent with larger weights over the western Pacific and is the optimal excitation for the model ENSO mode.

Relevant to the linear scenario discussed here is the low frequency of the projection of \(K\) onto the corresponding \(K_{\text{opt}}\).

\[
p_{K}(t) = \int K(x, t)K_{\text{opt}}(x)\,dx,
\]

where the integral is performed over the zonal domain of the ocean model. Therefore one would expect a large model response if the low-frequency tail of \(p_{K}\) is large. Figure 10 shows the Niño-3 index (mean temperature in the domain 5°S–5°N, 150°–90°W) for the period January 1980–June 2003 when the coupled model is forced with \(\tau_{\text{MJO}}\) (Fig. 10b) and with \(\tau_{\text{NMJO}}\) (Fig. 10c). Also included in these figures is the 90-day running mean of \(p_{K}\) as a proxy of its low-frequency tail (the amplitudes have been rescaled to aid comparison with the Niño-3 index). It can be seen that even though the amount of energy at low frequencies in \(\tau_{\text{MJO}}\) and \(\tau_{\text{NMJO}}\) is about the same (Fig. 10a), \(\tau_{\text{MJO}}\) has a generally larger projection on the leading stochastic optimal. Therefore this forcing is expected to produce a larger response as is the case. The linearity of the response to \(\tau_{\text{MJO}}\) was corroborated by forcing the model with the 90-day running mean of \(\tau_{\text{MJO}}\), which is also shown in Fig. 10b (dashed line). It should be noted however that what is observed in Figs. 10b and 10c is not the direct response of the ocean to the externally imposed atmospheric forcing. The time taken by the Kelvin waves to cross the Pacific is about 3 months, and therefore the part directly forced by \(\tau_{\text{MJO}}\) is felt in the eastern Pacific after about this time. The lag in the response observed in Figs. 10b and 10c fluctuates between 6 and 12 months, reflecting the fact that the observed SSTs are the result of both stochastic and coupled atmospheric forcings as is shown in Fig. 8c. When the Kelvin waves become coupled to the atmosphere in the central and eastern Pacific, they slow down considerably (Hirst 1986).

The magnitude of the projection of \(\tau_{\text{MJO}}\) and \(\tau_{\text{NMJO}}\) onto the second stochastic optimal, which explains about 15% of the stochastically induced variance, was found to be about one-tenth of that for the first stochastic optimal and therefore does not contribute substantially to the modeled variability.

4. Summary

In this work we have analyzed the possibility that the tropical Pacific coupled ocean–atmosphere system linearly amplifies perturbations produced by the leading mode of intraseasonal variability, the Madden–Julian oscillation (MJO). This required an estimate of the low-frequency tail of the MJO. Using 23 yr of NCEP–NCAR reanalyses and Reynolds SST, we have identified and removed the ENSO-related wind anomalies. The resulting wind residual was then used to produce two estimates of surface stress associated with the MJO. The first estimate (\(\tau_{\text{MJO}}\)) was based on the intraseasonally bandpassed wind residual. In the second estimate (\(\tau_{\text{NMJO}}\)), we isolated the MJO signals with the same spatial structures as those of the first estimate but including its low-frequency variability. The two estimates are very similar in terms of their time evolution, distribution and size of variance, etc. We have found that the spatial structure that dominates the intraseasonal band also dominates the low-frequency band of the anomalies that are independent of ENSO SST. We define the non-MJO contribution (\(\tau_{\text{NMJO}}\)) as that part of the residual with westward variance, eastward or steady variance with time scales shorter than 30 days, and all the variance with global wavenumbers higher than 2.

The low-frequency contribution of the MJO is a mode of variability not included in most coupled models used to date. We therefore studied its effect on ENSO variability using a coupled model of intermediate complexity. It was found that this “MJO-like forcing” can explain a large fraction of the interannual variability in an asymptotically stable version of the model. The modeled interaction between \(\tau_{\text{MJO}}\) and ENSO takes place via linear dynamics: The largest coupled wind anomalies are initiated after a sequence of several Kelvin waves of the same sign have been forced by \(\tau_{\text{MJO}}\). Coupled processes are identified as important in amplifying these perturbations. The cumulative effect of the stochastically forced Kelvin waves is to persist the (small) SST anomalies in the eastern Pacific just long enough for the coupled mechanism to develop into a mature ENSO. The amount of energy in \(\tau_{\text{MJO}}\) is only a modest fraction of that contained in the residual stress. However, a large fraction of the modeled interannual variability is excited by the \(\tau_{\text{MJO}}\) estimate.

Relevant to the linear scenario is the low-frequency content of the different forcings. Even though the amount of energy at low frequencies in \(\tau_{\text{MJO}}\) and \(\tau_{\text{NMJO}}\) is about the same, \(\tau_{\text{MJO}}\) has a larger projection on the leading stochastic optimal (i.e., the spatial pattern of stochastic forcing preferred by the model), explaining why a larger fraction of the modeled variance is linearly excited by the \(\tau_{\text{MJO}}\) estimate.

It could certainly be possible that what we call “the low-frequency tail of the MJO” is actually another phenomenon having the same spatial structure as that of...
the intraseasonal MJO ($\tau_{\text{MJO}}$). If that were the case, we would expect that low-frequency component to occur irrespectively of the MJO. We have found however that this low-frequency variability is synchronized with anomalies of seasonal MJO activity. To demonstrate this, we have compared two indices, one measuring the interannual anomalies in seasonal activity of the intraseasonal MJO ($\tau_{\text{MJO}}$), and the other measuring the low-frequency tail of $\tau_{\text{MJO}}$. The former is measured by computing the running standard deviation, $\overline{\sigma}$, within a 3-month running window of $\tau_{\text{MJO}}$ averaged over the equatorial western Pacific ($5^\circ S$–$5^\circ N$, $130^\circ E$–$180^\circ E$). When the annual cycle of $\overline{\sigma}$ is removed, the resulting time series, $\Delta \overline{\sigma}$, represents interannual anomalies in seasonal activity (i.e., it measures above or below normal MJO activity). Note that this definition is based on the intraseasonal MJO, and therefore the low-frequency variance is not included in this index. The second index, measuring the low-frequency variability of $\tau_{\text{MJO}}$, is defined as the running mean using the same time window and spatial domain and is denoted as $\overline{\tau_{\text{MJO}}}$. The time–evolution of these two indices (Fig. 11) suggests that indeed the low-frequency variability observed in the $\overline{\tau_{\text{MJO}}}$ estimate is due to anomalously high or low activity of the intraseasonal MJO. The correlation between the two time series is 0.6, which is significant at the 95% level. Similar time series with significant correlations were obtained when the spatial domain considered was extended in latitude and longitude and the length of the running window was varied up to 12 months.

Our experiments suggest that it is this low-frequency tail that is important for realistic ENSO simulations, in agreement with stochastic linear theories of ENSO (e.g., Penland 1996; Moore and Kleeman 1999a; ZG03). This result is also consistent with the work of Roulston and Neelin (2000), who conclude that impacts of the MJO are largely restricted to the low-frequency tail rather than the intraseasonal MJO. However in that work the “MJO” forcing was constructed from the EOFs of the residual monthly FSU stresses with its temporal variability modeled as first-order autoregressive processes and then artificially enhancing the power at the 30–60-day band. Therefore, in addition to the potential problems associated with the use of monthly values to characterize the SF, that forcing included spatial structures not associated with the MJO, and the temporal evolution was somewhat artificial. In the present work we have considered daily anomalies in the equatorial belt ($15^\circ S$–$15^\circ N$). This allowed us to clearly isolate the spatial and temporal structure of the MJO. We have shown here that a large fraction of the low-frequency tail of the residual stress can be associated with the MJO. This distinction is important since the MJO seems to possess the optimal spatial pattern to perturb the coupled system. In theory, this low-frequency variability should be reproduced by global models. However, an evaluation of several model simulations of the MJO shows that if a model can approximate the observed space–time spectral characteristics of the MJO, it is extremely rare that the simulated MJO has the correct structure, spatial distribution, and seasonal cycle (Zhang 2005, manuscript submitted to Rev. Geophys.). As emphasized here, it is not just the low-frequency variability but also the spatial structure that allows the MJO to linearly influence ENSO.

The obvious question is then, what produces such low-frequency variability of the MJO. This is a question that cannot be answered in the present study and is a topic of intense research. Roulston and Neelin (2000) speculate that the observed reddening in the residual stress (not just MJO), referred to as “climate noise,” is likely due to SST outside the tropical Pacific. Indeed, Vimont et al. (2003) have identified a “seasonal footprinting mechanism” in which part of the summer variability in the tropical Pacific can be attributed to mid-latitude SST anomalies forced by the overlying atmosphere during the previous winter.

It is also well known that the centers of activity of the MJO and WWBs exhibit an east–west migration as the warm-pool edge meanders in response to ENSO phases (Kessler 2001; Yu et al. 2003) and that such meandering is influenced by the MJO itself (e.g., Picaut and Delcroix 1995; Vialard et al. 2001; McPhaden 2004). More-

![Fig. 11. Time evolution of interannual anomalies in seasonal activity of the intraseasonal MJO ($\Delta \overline{\sigma}(\tau_{\text{MJO}})$), and low-frequency evolution of the MJO ($\overline{\tau_{\text{MJO}}}$) over the western equatorial Pacific ($5^\circ S$–$5^\circ N$, $130^\circ E$–$180^\circ E$). Vertical dotted lines indicate 1 Jan of each year.](image-url)
over, this meandering of the warm-pool edge helps to set up favorable conditions for tropical cyclone development associated with midlatitude cold surges (Yu and Rienecker 1998). These tropical cyclones are however not included in our MJO estimate (see section 2c).

In addition to these external factors, it is also possible that the low-frequency variability of the MJO arises from its irregularity (i.e., its departures from being a perfect oscillation). Between the possible sources of such irregularity are the marked asymmetry between the surface winds (with westerlies dominating over easterlies), and variations in the timing between events, amplitude, and duration. For example, Yano et al. (2004) have shown that a large fraction of the observed low-frequency variability of the surface zonal stress at 160°E can be explained by the occurrence of pulse-like westerly wind events (i.e., lasting several weeks), characteristics of the MJO. This is a simple consequence of the well-known fact that a pulse-like event has a wide spectral contribution. We have shown here that not considering this asymmetry (i.e., by bandpassing the anomalies) can produce very different results.

The large zonal extent is an important factor that differentiates the MJO from other sources of stochastic forcing. However, since the shape of the stochastic optimals is model dependent, these results should be corroborated using other models. Also, the potential exists for other processes to be important such as nonlinear interactions and heat flux anomalies associated with the MJO. This requires the use of more complicated ocean physics such as that considered by oceanic global circulation models. This is work in progress to be reported later.

Acknowledgments. This work was supported by a grant from the NOAA Office of Global Programs through the Cooperative Institute for Marine and Atmospheric Studies (CIMAS). We thank three anonymous reviewers for their valuable comments that led to significant improvement of the original manuscript. The NCEP–NCAR reanalysis data were obtained from the NOAA–CERES Climate Diagnostics Center.

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