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Large scale dynamics and MJO forcing of ENSO variability

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[1] A simple two-predictor regression model is developed to estimate the relative influence of large-scale low frequency ocean-atmosphere dynamics and high frequency atmospheric forcing on peak sea surface temperature (SST) anomalies associated with El Niño/Southern Oscillation (ENSO) variations for the period 1980–2005. One predictor is equatorial warm water volume (WWV), which is an index for the role that upper ocean heat content plays in regulating ENSO variability. The other predictor characterizes high frequency atmospheric forcing in the western Pacific linked to the Madden-Julian Oscillation (MJO). The two-predictor model accounts for about 60–65% of peak Nino3.4 SST anomaly variance at 2–3 season lead times and suggests about equal influence (on average) of low frequency dynamical processes and the MJO on peak ENSO SST anomalies over the past 25 years. The implications of these results for ENSO prediction are discussed.
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1. Introduction

[2] Year-to-year variability associated with the El Niño/Southern Oscillation (ENSO) is governed by large-scale ocean dynamics and coupled ocean-atmosphere interactions [*Schopf and Suarez*, 1988; *Battisti and Hirst*, 1989]. These processes result in alternating periods of anomalously warm El Niño conditions and cold La Niña conditions every 2–7 years. No two El Niño or La Niña events are exactly alike though, and the cycle between warm and cold phase ENSO conditions exhibits considerable irregularity in amplitude, duration, temporal evolution, and spatial structure.

[3] One factor contributing to this irregularity is high frequency atmospheric forcing in the form of episodic weather events, which is a prominent feature of most El Niños over the past 50 years. According to theory, the precise role of this high frequency forcing depends on ENSO stability characteristics, which are governed by large-scale background wind and ocean temperature conditions [*Fedorov and Philander*, 2000]. If ENSO results from instability of the coupled ocean-atmosphere system, high frequency forcing will introduce differences in the details of individual warm and cold events whose evolution is largely determined by the growth of the instabilities. If on the other hand ENSO is a damped or stable oscillator, high frequency forcing will be essential for initiating and sustaining ENSO events [*Moore and Kleeman*, 1999]. Stability characteristics of ENSO are nonstationary depending on the slow changes in background conditions in the tropical Pacific, though *Fedorov and Philander* [2000] estimate the system was near neutral stability during the 1980s and 1990s.

[4] A number of modeling and observational studies have addressed the role of episodic westerly wind forcing, often treated as stochastic noise, on the ENSO cycle [e.g., *Blanke et al.*, 1997; *Kessler and Kleeman*, 2000; *Bergman et al.*, 2001]. The most recent of these studies have indicated that background SSTs associated with developing ENSO events can modulate this high frequency atmospheric forcing, suggesting that it may be partially deterministic in character [*Batstone and Hendon*, 2005; *Eisenman et al.*, 2005; *Vecchi et al.*, 2006]. To complement and extend these recent studies, we use an empirically based regression analysis to assess the role of large-scale low frequency dynamics, high frequency MJO forcing, and their relationship to one another in the development of peak Nino3.4 SST anomalies for the period 1980–2005.

2. Indices and Model Formulation

[5] We develop a simple two-predictor regression model for the Nino3.4 SST index, which is a commonly used measure to determine the state of ENSO. One predictor is equatorial warm water volume (WWV) integrated across the basin, which embodies the essential physics of the large-scale variability associated with ENSO [*Meinen and McPhaden*, 2000]. The other predictor is an index of Madden-Julian Oscillation (MJO) variability, which represents one of the most significant sources of high frequency forcing on the ENSO cycle [*Moore and Kleeman*, 1999; *Hendon et al.*, 2006]. Though a variety of other phenomena contribute to weather variations that can influence ENSO, including cold air outbreaks from mid-latitudes [*Yu and Reinecker*, 1998], tropical cyclones [*Keen*, 1982], and generic westerly wind bursts [*Vecchi and Harrison*, 2000], variability associated with the MJO accounts for about half the variance in surface winds on intraseasonal time scales in the region of the western Pacific warm pool where the evolution of ENSO is particularly sensitive to episodic forcing [*Hendon et al.*, 2006].

[6] Monthly anomalies of Nino3.4 SST, WWV, and the MJO surface wind index for the western Pacific (MJO_{wpac}) over the period 1980–2005 are shown in Figure 1. In each case, anomalies are generated by removing the respective mean seasonal cycles. The Nino3.4 SST index, which is averaged over 5°N–5°S, 120°–170°W, is taken from the U.S. Weather Service Climate Prediction Center web page (http://www.cpc.ncep.noaa.gov/). WWV is computed as the volume integral of water above the 20°C isotherm

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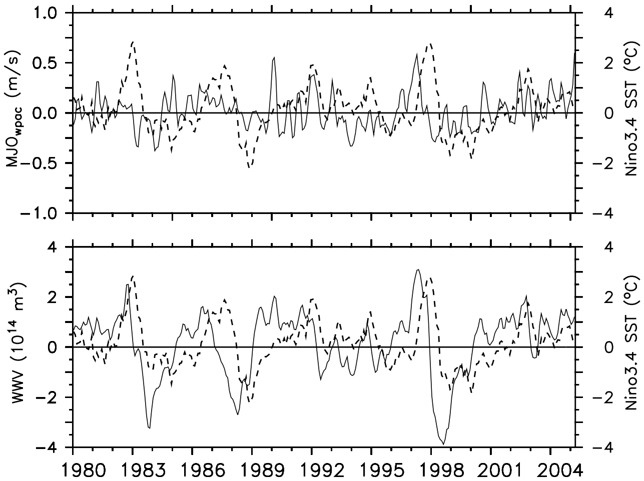

Figure 1. Monthly anomalies of Nino3.4 SST (dashed lines), WWV, and MJO_{wpac}.

between 5°N–5°S, 120°E–80°W inferred from an analysis of mooring, Argo, and expendable bathythermograph data. The WWV index is a measure of the recharge and discharge of heat into the equatorial band, which is an important determinant of variability associated with the ENSO cycle [e.g., *Jin*, 1997]. It is also a leading predictor of Nino3.4 SST and is especially valuable for constraining predictions from the early part of the calendar year at 2–3 season lead times because, unlike SST, it exhibits no spring persistence barrier [*McPhaden*, 2003; *Clarke and Van Gorder*, 2003].

[7] The index of MJO activity (MJO_{wpac}) is computed as in *Hendon et al.* [2006] using the NCEP/NCAR reanalysis [*Kalnay et al.*, 1996]. Daily time series of surface zonal winds are band pass filtered at periods of 30–95 days. We retain global wavenumbers 1–5 (eastward propagating only) as in *Hendon et al.* [2006] and earlier studies [e.g., *Wheeler and Weickmann*, 2001]. The standard deviation is then computed over 90 day running periods and averaged in the western Pacific over 5°N–5°S, 120°E–180°, after which the mean seasonal cycle is removed. *Zhang and Gottschalck* [2002] developed a similar MJO index and related it to SST variability along the equator. However, their index was based on the oceanic Kelvin-wave response to MJO forcing rather than zonal wind variability associated with the MJO itself.

[8] The zero lag all season correlation between Nino3.4 SST and MJO_{wpac} is only 0.15, which might lead to the conclusion that there is no consistent relationship between ENSO and the MJO (arrived at in several previous studies as described by *Hendon et al.* [2006]). However, *Hendon et al.* [2006] identified a robust lagged relationship between MJO activity in late boreal spring and ENSO development in the subsequent autumn/winter. They showed for example that enhanced MJO activity in the western Pacific during spring is associated with an eastward expansion of the western Pacific warm pool and enhanced westerly surface zonal wind anomalies, which are well known precursors to development of El Niño [e.g., *Kessler et al.*, 1995; *McPhaden*, 1999].

[9] Lagged correlations as a function of lead-time and calendar month for a) Nino3.4 SST and WWV and b) Nino3.4 SST and MJO_{wpac} are shown in Figure 2. WWV in February–April (FMA) is highly correlated (>0.7) with Nino3.4 at lead times of 8–12 months. Thus peak ENSO SST anomalies at the end of the calendar year are well predicted by late boreal winter/spring WWV anomalies as discussed previously by *McPhaden* [2003] and *Clarke and Van Gorder* [2003]. Similarly, MJO activity in surface winds in the western Pacific in April–June (AMJ) is highly correlated with peak ENSO SST anomalies 6–10 months later. *Moore and Kleeman* [1999] and *Hendon et al.* [2006] identify boreal spring as the time when growth of ENSO SST anomalies will be most sensitive to MJO forcing because that is a transitional time of year associated with the seasonal re-establishment of the equatorial cold tongue in the central and eastern Pacific. According to these arguments, if MJO activity is elevated, seasonal development of the cold tongue can be arrested, whereas if MJO activity is reduced, cold tongue development can be accelerated.

[10] We use these lagged relationships illustrated in Figure 2 to form a simple regression model for estimation of Nino3.4 SST anomalies at or near the peak of the ENSO event, namely, either the December–February (DJF) or October–December (OND) seasons, based on antecedent WWV and MJO_{wpac} conditions:

$$\text{Nino3.4 SST(DJF)} = \alpha \text{WWV(FMA)} + \beta \text{MJO}_{wpac}\text{(AMJ)} + \varepsilon, \quad (1)$$

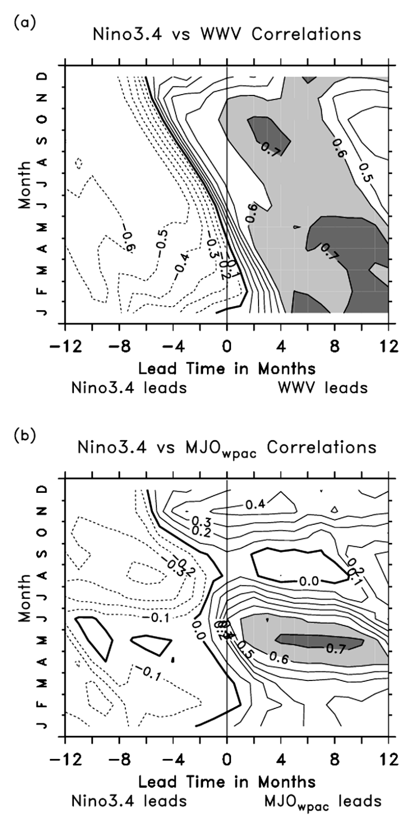

Figure 2. (a) Correlation between WWV and Nino3.4 SST anomalies as a function of lead-time and month. (b) Correlation between MJO_{wpac} and Nino3.4 SST anomalies as a function of lead-time and month. Correlations ≥0.6 are shaded.

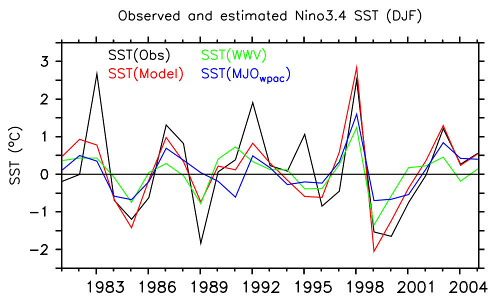

Figure 3. Observed DJF Nino3.4 SST (black line) and Nino3.4 SST estimated from (1) (red line). Also shown is Nino3.4 estimated separately from αWWV in FMA (green line) and βMJO_{wpac} in AMJ (blue line).

where small residual means have been first removed from the time series and variability unaccounted for by the regression is represented by ε We test the applicability of the regression model in a cross-validated fashion. That is, α and β are computed for each target season/year through linear least-squares fitting applied only to data from other years in the record. As we have 25 full DJF, FMA, and AMJ seasons available between 1980 and 2005, we are thus are able to compute α, β, and the predicted SST a total of 25 times, building each regression model from an independent set of 24 years of data.

[11] Determination of the statistical significance of the results requires an estimation of the effective time between independent samples. Various methods are available to estimate such integral time scales. These methods, as summarized by *Davis* [1976] and *Kessler et al.* [1996], lead to values ranging from 6 months to 2 years for each of NINO3.4 SST, WWV and MJO_{wpac}. We use the method preferred by *Kessler et al.* [1996], that is, the integral of the autocorrelation function to the first zero crossing. This choice results in an integral time scale of approximately 1 year, which is in the mid-range of values for all three variables.

3. Results

[12] Results of the cross-validated regression analysis show that this simple two predictor model accounts for about 65% of the observed peak season (DJF) Nino3.4 SST anomaly for the period 1980–2005 with correlation between estimated and observed Nino3.4 SST of 0.8 (Figure 3, Table 1). The model seems to perform better after 1995, before which estimated Nino3.4 amplitudes tend to be lower than observed (Figure 3). The apparent improvement in model performance after 1994 may be due to the completion in December 1994 of the Tropical Atmosphere Ocean (TAO) array which since the mid-1990s has provided the bulk of upper ocean thermal field data within 5° of the equator in the Pacific [*Smith and Meyers*, 1996] and the majority of the in situ equatorial Pacific surface wind data for assimilation into NCEP atmospheric analyses [*McPhaden et al.*, 1998].

[13] On average, the relative contributions of the WWV and MJO_{wpac} indices to model skill are comparable, with correlations between Nino3.4 SST and αWWV of 0.71 and between Nino3.4 and βMJO_{wpac} of 0.72, respectively (Table 1). The contributions of these two predictors varies for individual events with, for example, WWV being more important for the 1998–99 La Niña and MJO_{wpac} accounting for almost all the variance of the weak 2004–05 El Niño-like event (Figure 3). Both predictors on the other hand missed the 1994–95 DJF warm anomaly for reasons that are not entirely clear.

[14] Results were essentially the same using October–December instead of December–February Nino3.4 SST as the predictand (Table 1). We also repeated these calculations using Nino3 instead of Nino3.4. WWV and MJO_{wpac} variations contributed roughly equally to estimated Nino3 SST variations and together accounted for about 55–60% of the observed Nino3 SST variance. We likewise redid the analysis using a quadratic form of the WWV predictor, that is, $\alpha_1 WWV + \alpha_2 WWV^2$, to account for potential asymmetries in the relationship between positive and negative values of WWV and ENSO cold tongue SST anomalies [*Meinen and McPhaden*, 2000]. However, taking these asymmetries into account did not lead to statistically significant differences in the regression analysis results for either Nino3 or Nino3.4 SST. Collectively, these various tests indicate a robust relationship between both WWV and MJO_{wpac} variability early in the calendar year and subsequent peak season ENSO SST anomalies in the eastern and central Pacific.

[15] It has been argued from both theoretical and modeling studies that the impact of MJO variability on ENSO is felt primarily through residual low frequency westerly surface wind anomalies (i.e., the "low frequency tail" of the MJO spectrum that results from either the episodic

Table 1. Regression Results for Nino3.4 Estimates in October–December (OND) and December–February (DJF) Based on Equation (1)[a]

	Obs. Correlated With Est. SST	Obs. Correlated With αWWV (FMA)	Obs. Correlated With βMJO_{wpac} (AMJ)	
Nino3.4 (DJF)	0.80 (0.63, 0.90)	0.71 (0.49, 0.85)	0.72 (0.50, 0.85)	
Nino3.4 (OND)	0.77 (0.58, 0.88)	0.67 (0.42, 0.82)	0.73 (0.52, 0.86)	
	α, °C per 10^{14} m^3	β, °C per m s^{-1}	RMS SST Diff., °C	Std Dev Nino3.4 SST, °C
Nino3.4 (DJF)	0.43 ± 0.19	3.41 ± 1.52	0.71	1.17
Nino3.4 (OND)	0.41 ± 0.20	3.57 ± 1.56	0.75	1.18

[a]Cross-validated correlations between observed Nino3.4 and Nino3.4 SST estimated from (1) for October–December (OND) and December–February (DJF); correlations between observed Nino3.4 SST and that predicted from αWWV (FMA) and βMJO_{wpac} (AMJ) separately; estimated regression coefficients α and β; root-mean square (RMS) difference between estimated and observed Nino3.4 SST; and the standard deviation of Nino3.4 SST for the season indicated. Coefficients α and β represent the average of the regression coefficients from each of the 25 years; likewise the 90% confidence limits for these coefficients represent the average of the 90% confidence limits for each of the 25 years.

occurrence of the MJO or an induced coupled response to it) rather than through the details of the episodic forcing itself [*Roulston and Neelin*, 2000; *Lengaigne et al.*, 2004; *Zavala-Garay et al.*, 2005; *Batstone and Hendon*, 2005; *Hendon et al.*, 2006]. To test this idea, we developed an alternative version of (1) in which βMJO_{wpac} (AMJ) was replaced by βU_{wpac} (AMJ) where U_{wpac} (AMJ) is the seasonal mean (AMJ) surface zonal wind anomaly in the region 5°N–5°S, 120°E–180°. This seasonal zonal wind anomaly is highly correlated with MJO_{wpac} in AMJ (correlation coefficient of $\sim$0.75) and more strongly correlated at this time of year than at any other [*Hendon et al.*, 2006]. Using this version of the model, cross-validated predictions of Nino3.4 are virtually identical to those shown in Figure 3 with a correlation between predicted and observed Nino3.4 SST of 0.83 (90% confidence limits of 0.68 and 0.91).

[16] These results support the idea that westerly anomalies associated with the low frequency tail of the MJO spectrum are primarily responsible for the impact of intraseasonal variability on ENSO. However, one reviewer suggested an alternative interpretation of the relationship between the MJO and seasonally averaged zonal winds, namely mean westerly winds favor the development of MJO activity [*Zhang*, 2005]. During the AMJ season though, mean zonal winds in our western Pacific index region are almost always easterly even when anomalies are westerly because of the underlying seasonal cycle. Thus, while the low frequency tail of MJO variability may not be the only source of AMJ westerly anomalies in the western Pacific, it appears to be a significant contributor to those anomalies.

4. Discussion and Conclusions

[17] In this study we have developed a simple two-predictor regression model to estimate the relative influence of large-scale, low frequency, ocean-atmosphere dynamics and high frequency atmospheric forcing on peak ENSO SST anomalies over the period 1980–2005. One predictor is WWV, which is an index for the role that upper ocean heat content plays in regulating ENSO variability. The other predictor, MJO_{wpac}, characterizes episodic high frequency zonal wind forcing during boreal spring in the western Pacific associated with the MJO. These are not the only predictors one could use to characterize the relative contributions of large-scale low frequency dynamics and shorter time scale weather variability on the ENSO cycle, nor do they account for all the observed variance in ENSO SST anomalies in the eastern and central equatorial Pacific. However, our simple model shows that together these two indices alone account for 60–65% of peak Nino3.4 SST anomaly variance at 2–3 season lead times. Moreover, the model suggests about equal influence (on average) of high frequency wind forcing associated with the MJO and low frequency large-scale dynamics on ENSO variability over the past 25 years.

[18] Our model (1) with βMJO_{wpac} (AMJ) replaced by βU_{wpac} (AMJ) is analogous to the statistical prediction model developed by *Clarke and Van Gorder* [2003], though our interpretation of the western Pacific seasonal surface wind anomalies in terms of the low frequency spectral tail of MJO forcing is different. Also, the high spring season correlation between MJO_{wpac} and U_{wpac} weakens later in the calendar year, suggesting that large scale surface zonal wind anomalies that subsequently develop during the evolution of El Niño and La Niña events result primarily from positive feedbacks with SST anomalies. In this sense, our model is a special case of the *Clarke and Van Gorder* [2003] model in that we do not consider the full evolution of the ENSO cycle, but just the long lead-time WWV and MJO influences from the first half of the calendar year on later peak cold tongue ENSO SST anomalies. For this reason, and since our MJO index requires time-filtering to construct (which is problematic for real-time application), it may be more practical to use the mean zonal wind rather than MJO_{wpac} in empirical ENSO forecast schemes.

[19] It is evident from Figure 3 that our two indices are correlated with one another. More generally, individual monthly means of WWV and MJO_{wpac} are significantly correlated (0.5–0.6) at 0 $\pm$ 1–3 month lags during the months of November to June. There are two possible interpretations of this correlation, one of which has been mentioned already, namely that the background state modulates high frequency forcing associated with the MJO. Another interpretation is that WWV is influenced by the high frequency component of the wind forcing. Kelvin waves have basin scale zonal wavelengths [*Zhang and Gottschalk*, 2002] and as MJO variance increases during the onset of El Niño, thermocline depth variations associated with the envelope of wind-forced Kelvin waves will project onto increases in WWV. These two possibilities are not mutually exclusive and it is likely that both contribute to the correlation.

[20] Our results suggest that ENSO forecasts from the early part of the year, whose skill is strongly influenced by the presence of precursory subsurface thermal field anomalies [*Latif et al.*, 1998; *Clarke and Van Gorder*, 2003], will regularly miss a percentage of the peak ENSO SST anomalies unless the statistics of the MJO can be predicted in the AMJ season. Peak SST anomalies were in fact consistently underestimated from early in the calendar year during the 1997–98 El Niño [*McPhaden*, 1999; *Barnston et al.*, 1999], the 2002–03 El Niño [*McPhaden*, 2004], and the most recent weak warm El Niño-like event in 2004–05. In each case, underestimates could be related to the effects of unanticipated boreal spring episodic westerly wind forcing associated with the MJO. In the case of the weak 2004–05 El Niño-like warming, predictions made by most organizations from early 2004 called for neutral conditions to prevail in the tropical Pacific, since early calendar year WWV precursors were largely absent. It was not until after the occurrence of eastern and central Pacific SST warming associated with a westerly wind event in June 2004 that forecasters began to consistently predict the subsequent development of an El Niño [*Lyon and Barnston*, 2005].

[21] Finally, in view of record length limitations, it is not possible to unambiguously assert from statistical considerations alone that two-predictor empirical models involving western Pacific zonal winds and WWV are superior to one predictor models for estimating peak Nino3.4 SST (cf. *Kug et al.*, 2005). Specifically, our two-predictor model cannot, with 90% confidence, outperform a single predictor model using either MJO_{wpac} or WWV. It would require about 80 degrees of freedom (equivalent to about 80 years of data) to discriminate between correlations of 0.80 from

0.72 for DJF Nino3.4 SST (Table 1) at this level of confidence. However, these two correlations are significantly different from one another with 60% confidence. We would desire a higher level of confidence in reporting our results, but we would have to wait another 55 years (assuming stationarity of the time series) to distinguish statistically between the one and two predictor models if we were to adopt a standard of 90%. The choice of confidence limits is ultimately arbitrary in any case, whether they are 90%, 95%, or 99% [*Nicholls*, 2001]. Thus, despite the relatively low statistical confidence of our results by conventional standards, we feel that the physical concepts discussed in this paper warrant highlighting as a stimulus to further modeling and theoretical work.

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