The Role of Stochastic Forcing in Ensemble Forecasts of the 1997/98 El Niño

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ABSTRACT

The impact of stochastic intraseasonal variability on the onset of the 1997/98 El Niño was examined using a large ensemble of forecasts starting on 1 December 1996, produced using the Australian Bureau of Meteorology Predictive Ocean Atmosphere Model for Australia (POAMA) seasonal forecast coupled model. This coupled model has a reasonable simulation of El Niño and the Madden–Julian oscillation, so it provides an ideal framework for investigating the interaction between the MJO and El Niño. The experiment was designed so that the ensemble spread was simply a result of internal stochastic variability that is generated during the forecast. For the initial conditions used here, all forecasts led to warm El Niño-type conditions with the amplitude of the warming varying from 0.5° to 2.7°C in the Niño-3.4 region.

All forecasts developed an MJO event during the first 4 months, indicating that perhaps the background state favored MJO development. However, the details of the MJOs that developed during December 1996–March 1997 had a significant impact on the subsequent strength of the El Niño event. In particular, the forecasts with the initial MJOs that extended farther into the central Pacific, on average, led to a stronger El Niño, with the westerly winds in the western Pacific associated with the MJO leading the development of SST and thermocline anomalies in the central and eastern Pacific. These results imply a limit to the accuracy with which the strength of El Niño can be predicted because the details of individual MJO events matter. To represent realistic uncertainty, coupled models should be able to represent the MJO, including its propagation into the central Pacific so that forecasts produce sufficient ensemble spread.

1. Introduction

The role of the Madden–Julian oscillation (MJO) in the onset and development of El Niño is a much-debated issue. Results from analyses of observed and model data and from different methods show inconsistent results. On the one hand, based on the investigation of observed (reanalysis) data, it has been suggested that there is no significant relationship between interannual variations of the level of global MJO activity and sea surface temperature (SST) variation representing the El Niño–Southern Oscillation (ENSO; e.g., Salby and Hendon 1994; Hendon et al. 1999; Slingo et al. 1999; Kessler 2001). In addition, many dynamical coupled models do not have explicit MJO activity (e.g., Zebiak and Cane 1987; Latif et al. 1998; Chen et al. 2004) but are used successfully for prediction of ENSO events up to nine months in advance, suggesting that the MJO may not be necessary for ENSO development (e.g., Zhang et al. 2001).

On the other hand, several studies suggest a possible interaction between the MJO and ENSO and that the MJO may contribute importantly to the triggering, evolution, and strength of El Niño (e.g., Lau and Chan 1986, 1988; Kessler et al. 1995; Krishnamurti et al. 2000; Bergman et al. 2001; Zhang and Gottschalck 2002; Lengaigne et al. 2004). A robust lagged relationship between boreal spring MJO activity in the western Pacific and subsequent El Niño strength was recently reported by Hendon et al. (2007), whose results also suggest that the causal relationship stems from sustained westerly anomalies in the western Pacific that accompany enhanced MJO activity and that these westerly anomalies project efficiently onto the developing El Niño.

As an example, the onset of the 1997/98 El Niño was associated with several MJO events and has been the focus of several studies. McPhaden (1999), for instance, documented the MJO activity in the winter of 1996/97 that was associated with the onset of the 1997/98 El Niño. Several studies have indicated that the equatorial
westerly wind events (WWEs) associated with the MJO activity were the key for the interaction of the MJO with onset and development of this El Niño (e.g., Slingo 1998; van Oldenborgh 2000; Krishnamurti et al. 2000; Bergman et al. 2001; Lengaigne et al. 2003, 2004; Boulanger et al. 2004; Batstone and Hendon 2005). However, the WWE enhancement at the onset of ENSO depends, at least partly, on the slow evolution of the SST anomaly (SSTA) in the central Pacific and should thus not be considered purely “atmosphere only” stochastic noise (e.g., Yu et al. 2003; Batstone and Hendon 2005; Eisenman et al. 2005; Vecchi et al. 2006). Critically, however, most of the numerical models that succeeded in predicting the onset of the 1997/98 El Niño were unable to forecast its intensity until the WWE was incorporated (e.g., Barnston et al. 1999; Landsea and Knaff 2000; Vitart et al. 2003; Vecchi et al. 2006). The hypothesis, then, is that the MJO is relevant to the timing, initial growth, and strength of the developing El Niño but not necessary for the event itself (e.g., Kleeman and Moore 1997).

The aim of the present paper is to explore the influence of MJO events on the ensemble spread of coupled model forecasts of the 1997/98 El Niño. We use the Australian Bureau of Meteorology Predictive Ocean Atmosphere Model for Australia (POAMA) coupled seasonal prediction system, which exhibits skillful prediction of ENSO two–three seasons in advance (Alves et al. 2003). The coupled model component of the system has a good simulation of the ENSO, including its amplitude and phase locking to the seasonal cycle (Zhong et al. 2005). The POAMA model also simulates a reasonable MJO, including a realistic spectral peak at low zonal wavenumbers at eastward intraseasonal frequencies for lower-tropospheric zonal wind (however POAMA’s MJO occurs at a slightly lower frequency than observed) and a reasonable phasing, magnitude, and penetration of the eastward-propagating zonal wind and convection anomalies into the western Pacific (Zhang et al. 2006; Marshall et al. 2008). As will be shown, POAMA generates a relatively large spread of ENSO forecasts simply because of its internal variability and thus provides the ideal experimental framework for investigating the MJO–ENSO relationship. Unlike many previous studies (e.g., Roulston and Neelin 2000; Moore and Kleeman 1999; Zavala-Garay et al. 2005; Eisenman et al. 2005), we do not need to impose artificial MJO forcing in our methodology because the POAMA model naturally generates MJO activity.

A brief description of the experimental set up, the coupled model, and the ensemble generation method are described in section 2. Some basic features of ensemble results are examined in section 3, and the impact of the MJO on the spread of the ensemble is explored in detail in section 4. Finally, a discussion and summary are presented in section 5.

2. Description of the model and forecasts

a. Coupled model system

The first version of POAMA (POAMA-1; Alves et al. 2003) is the dynamical coupled model seasonal forecast system developed jointly by the Bureau of Meteorology Research Centre (BMRC) and Commonwealth Scientific and Industrial Research Organisation (CSIRO) Marine and Atmospheric Research (CMAR). The atmospheric component of POAMA-1 is the Australian Bureau of Meteorology unified atmospheric model (BAM 3.0d). It has a horizontal spectral resolution of T47 and 17 vertical levels. The ocean model component is the Australian Community Ocean Model version 2 (ACOM2). It was developed by CMAR, and is based on the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM) version 2. The grid spacing is 2° in the zonal direction. The meridional spacing is 0.5° within 8° of the equator, increasing gradually to 1.5° near the poles. There are 25 levels in the vertical, with 12 in the top 185 m. The maximum depth is 5000 m. Technical details of ACOM2 are given in Schiller et al. (2002).

The ocean and atmosphere models were coupled using the Ocean Atmosphere Sea Ice Soil (OASIS) coupling software (Valcke et al. 2000). No flux correction is applied to the exchanged fluxes between the oceanic and atmospheric models. [For more details about the POAMA model, see the POAMA Web site online at http://poama.bom.gov.au/ and Alves et al. (2003).]

A set of 180 nine-month hindcasts, one per month (started on the first day of each month) for the 15-yr time period of 1987–2001, has been used to generate the forecast model climatology and to assess the performance of POAMA-1 seasonal forecasts. Forecast anomalies used in this study are computed relative to this hindcast climatology, which is a function of start month and lead time. Strictly speaking, the year of the verification should be left out of the climatology but we have not done so as the results presented below are insensitive to the definition of the climatology.

Ocean initial conditions for the hindcasts were taken from an ocean data assimilation scheme that is based on the optimum interpolation (OI) technique described by Smith et al. (1991). Atmospheric initial conditions were taken from the appropriate date during a parallel Atmospheric Model Intercomparison Project (AMIP)-style integration of the atmosphere-only model forced with observed weekly SST from Reynolds and Smith (1994).
b. Experiments and ensemble generation method

The purpose of this study is to investigate the influence of intraseasonal variability on the spread of ensemble forecasts. We focus on forecasts for the onset of the 1997/98 El Niño and look at the role of stochastic forcing in determining the evolution and spread of 90 forecasts starting on 1 December 1996.

An ensemble generation method is designed for these experiments to reduce the effects of uncertainties in model initial conditions while at the same time to emphasize the uncertainties due to intraseasonal variability that develops during the forecast. Identical atmospheric and land initial conditions for all 90 forecasts were taken as the ensemble mean of the state on 1 December 1996 from 36 different simulations of the atmosphere-only model forced with observed SST (AMIP-style runs). The rationale of using the ensemble mean of the AMIP simulations for 1 December 1996 as the atmospheric initial condition was to remove as much MJO and other intraseasonal variability as possible from the initial state in order to emphasize the impact of the subsequent internally generated intraseasonal variability on the evolution of El Niño. However, this atmospheric initial condition results in initialization shock because the ensemble mean of the AMIP states is not a balanced solution of the atmospheric model equations. Furthermore, as a result of this shock, all individual members develop an MJO event in the first 30–45 days of the forecast. Nonetheless, the focus of this study is on the differences in behavior of the MJO events and subsequent evolution of El Niño in each member, so the deficiencies of this particular atmospheric initial condition are largely immaterial.

The ocean initial conditions were taken from the POAMA ocean data assimilation system valid for 1 December 1996 but included physically insignificant perturbations. These perturbations were included only for the purpose of introducing a small numerical difference to allow the stochastic part of the model, particularly the atmospheric model, to evolve differently. Small random SST perturbations, drawn from a uniform distribution of amplitude of 0.001°C and zero mean, were added to generate the 90 different initial ocean states. The magnitude of the SST perturbations is at least one order of magnitude smaller than those adopted by previous studies (e.g., Vialard et al. 2003).

3. Ensemble results

a. General ensemble characteristics

Figure 1 shows the evolution of the Niño-3.4 (5°S–5°N, 170°–120°W) SSTA for each of the 90 forecasts that were initialized on 1 December 1996. All forecasts in the ensemble show development of an El Niño. Comparison with the observed SSTA (dashed curve) shows that the forecasts were generally good, capturing both the timing and amplitude of the initial growth of the event. The spread of the ensemble, which provides some measure of the uncertainties attached to those forecasts, has a rapid growth during the first 4 months and is about 1.5°C at month 4. During months 5–9, the spread of the ensemble continues to increase, though more slowly, finally reaching about 2.2°C at 9 months. We will consider the role intraseasonal variability played in generating the spread of the ensemble forecasts.

Hovmöller plots of equatorially averaged (10°S–10°N) ensemble-mean anomalies of 10-m (surface) zonal wind (U10), SST, and depth of the 20°C isotherm (Z20) illustrate typical ENSO evolution (Figs. 2a–c). Anomalous surface westerlies (Fig. 2a) and the center of anomalous deep convection, as represented by negative outgoing longwave radiation (OLR) that we use as a proxy for deep convection (not shown), initially lies just to the west of the warmest SST anomalies (Fig. 2b) in the far western Pacific–eastern Indian Ocean and then slowly migrates eastward in association with the positive SST anomalies moving into the central-east Pacific during January–May 1997. Surface westerlies and convection is
FIG. 2. (a) Time–longitude diagram of U10 anomalies averaged over 10°S–10°N for the ensemble mean. The contour interval is 0.6 m s⁻¹, with shaded areas representing positive anomalies (westerly anomalies). (b) Time–longitude diagram of anomalies of the ensemble mean averaged over 5°S–5°N. The contour interval is 0.2°C, with shaded areas representing positive anomalies. (c) Time–longitude diagram of 20°C isothermal depth anomalies of the ensemble mean averaged over 5°S–5°N. The contour interval is 4 m, with shaded areas representing positive anomalies.
then sustained in the central-western Pacific through August 1997. Positive Z20 anomalies (Fig. 2c) also slowly propagate eastward from the western Pacific into the eastern Pacific. In addition, all ensemble members exhibit a similar intraseasonal behavior in the first few weeks of the forecasts (Fig. 2a), which results in the generation of a downwelling oceanic Kelvin wave (Fig. 2c) that arrives in the eastern Pacific at the end of January 1997, prior to the development of El Niño later in the year. This initial development of intraseasonal variability, as discussed in section 2, results partially from the spinup and initial adjustment resulting from the use of ensemble-mean atmospheric initial conditions.

b. Evolution of ensemble extremes

To understand the impact of the intraseasonal variability that develops during the forecasts, we investigated the processes that resulted in forecasts warmer and colder than the ensemble mean. For a clean separation, we grouped the 90-member ensemble into terciles based on the intensity of monthly mean Niño-3.4 indices: the 30 strongest, intermediate, and weakest cases representing strong, moderate, and weak El Niños, respectively.

The evolutions of the mean differences in U10, SST, and Z20 between the 30 strongest and 30 weakest forecasts after 9 months are shown in Figs. 2d–f, respectively. Significance of these differences is estimated using a t-test for the differences between two means (each formed with 30 samples). Differences significant at the 95% level are indicated by a heavy contour and shaded areas where appropriate in Figs. 2d–f. At the end of the forecast, the strongest cases show increased warming throughout the east Pacific and increased cooling in the west Pacific consistent with a stronger El Niño pattern (Fig. 2c). The differences in SST can be traced back as far as mid-February, when the central and east Pacific starts to warm and the west Pacific starts to cool in the 30 strongest cases, relative to the 30 weakest cases. The difference in U10 (Fig. 2d) shows that the 30 strongest cases develop stronger westerly winds in the west Pacific at the end of January/beginning of February, in comparison to the 30 weakest cases. These enhanced westerly winds, which are associated with the enhanced convection near the date line (not shown), lead to increased oceanic downwelling Kelvin waves during February–April (Fig. 2f). These results suggest that processes happening in the first few months of the forecasts precondition the longer-term development.

Table 1 shows the distribution of the 30 strongest and weakest cases at the end of the forecast (i.e., the end of August 1997) broken down according to the tercile of their monthly Niño-3.4 value from March to July 1997. Of the 30 forecasts in the upper tercile at the end of the 9 months, 17 are in the upper tercile for Niño-3.4 in March (i.e., at the end of 4 months), whereas only 1 is in the lower tercile in March. Similarly, of the 30 forecasts in the lower tercile at the end of 9 months, 15 are already in the lowest tercile in March and only 6 are in the upper tercile in March. These statistics indicate that the processes leading to enhanced (reduced) Niño-3.4 warming in the first 4 months are significantly related to enhanced (reduced) warming at 9 months, but there is not a one-to-one relationship.

4. Role of the MJO

a. Decomposition of the forecasts

The analysis in section 3b indicates that the state of the Niño-3.4 SST index at the end of the forecast relative to the ensemble mean can be traced back to the behavior after only 4 months. We focus now on development of the spread of the SST forecasts in the first 4 months and the role that the MJO plays. For the rest of this paper, we will refer to the upper tercile based on Niño-3.4 in March as the WARM set. Similarly, the middle tercile will be referred to as the NEUTRAL set and the lower tercile as the COLD set.

b. MJO propagation

Although all forecasts generate an MJO event in the first 2–3 months, the characteristics in each forecast differ significantly in strength, timing, phase speed, and extent of eastward penetration into the central Pacific. To extract the MJO activity from each ensemble member, we use multivariate empirical orthogonal function (EOF) analysis, similar to Wheeler and Hendon (2004, hereafter WH04), for the combined fields of equatorially averaged (10°N–10°S) OLR and U10. Anomalies of each field were calculated relative to the 90-member ensemble

Fig. 2. (Continued) anomalies. (d) As in (a), but for the difference in U10 (thin contours; interval 0.5 m s\(^{-1}\)) between the means of the 30 warmest and the 30 coolest cases. Bold contours indicate differences significant at better than the 95% significant level using a t test. Shaded areas represent westerly anomalies that pass the t test at the 95% significant level. (e) As in (d), but for SST averaged over 5°S–5°N. The contour interval is 0.2°C. Significance is indicated similar to (d). (f) As in (e), but for Z20. The contour interval is 2 m. The date line is indicated as a dashed line.
mean rather than to the hindcast climatology, thereby removing the mean interannual signal and leaving mainly the intraseasonal variability. No other temporal filtering was applied.

EOF analysis of the combined field results in a leading pair of EOFs, EOF1 and EOF2, which accounts for 11.42% and 9.89% of the explained daily variance of the combined fields, respectively. EOF1 (Fig. 3a) has negative OLR anomalies centered near 170°W over the central-western Pacific and positive OLR anomalies centered near 130°E over the eastern Indian Ocean–Maritime Continent. The corresponding maximum westerly (easterly) U10 anomalies are to the west of the center of the negative (positive) OLR anomalies. EOF2 (Fig. 3b) has its negative OLR anomalies peak near 160°E over the Maritime Continent–western Pacific and relatively weak positive OLR anomalies over the central-eastern Indian Ocean. The relationship between OLR and U10 pattern in EOF2 is similar to that in EOF1. The leading two EOFs can be taken together as a pair and together capture the eastward propagation of the MJO. We note that the two EOFs derived here are shifted approximately 1/4 cycle relative to the leading pair of EOFs shown in WH04. Moreover, the deepest convective center in the EOF1 (EOF2) in this paper (Fig. 3) shows a slightly eastward shift, compared to that in the corresponding EOF2 (EOF1) in WH04. Therefore, the approximate locations of the MJO cycle (Fig. 4) are different to that in WH04.

The principal component (PC) time series PC1 and PC2 show the peak correlation at a lag of about 15–18 days (not shown). The peak cross correlation is on average about 0.3 but is as high as 0.65 in some members. The maximum correlation coefficient between PC1 and PC2 is consistent with that in WH04 though the period of the model’s MJO is slightly longer than observed (e.g., Zhang et al. 2006; Marshall et al. 2008). The variation of the magnitude of the cross correlation of the PCs between each member is an indication that the level of MJO activity is highly variable across each ensemble member.

The behavior of the MJO in the forecasts is depicted in the two-dimensional phase space diagram defined by PC1 and PC2 for the WARM and COLD ensemble members (Figs. 4a,b). To remove high frequency noise, a 15-day running mean (which passes zero power after a 15-day period and approximately half power after a 30-day period) has been applied and each point represents the pentad-averaged location. The first 3 months are shown, except that the first 20 days of data are not plotted in order to eliminate any spinup effects.

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<td>30 weakest cases</td>
<td>6/9/15</td>
<td>5/6/19</td>
<td>1/8/21</td>
<td>0/6/24</td>
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Table 1. Distribution of the number of 30 strongest and weakest cases in August 1997 broken down according to the tercile of their monthly Niño-3.4 value from March to July 1997. The capital letters S, M, and W represent the strong, intermediate, and weak El Niño forecasts, respectively.

Fig. 3. Spatial structures of (a) EOF1 and (b) EOF2 of the combined OLR (solid) and U10 (dashed) ensemble-mean anomalies. Each field is averaged over the latitudes of 10°S–10°N and normalized by its global (all longitudes) variance before EOF analysis. The first 2 leading components account for 11.42% and 9.89% explained variance, respectively.
The anticlockwise circling traces the MJO activity that generally starts from the Indian Ocean (phase 2, marked as $S$) and then propagate eastward along the equator. The MJO activity in the WARM forecasts (Fig. 4a) systematically penetrates farther east into the central-western Pacific (phases 6 or 7, marked as *) in the first 3 months than in the COLD forecasts (Fig. 4b), in which the MJO appears to stall in the Maritime Continent region (phases 4 or 5). This result implies that in the first 3 months, the MJO events for the WARM forecasts propagate more systematically eastward and with relatively faster phase speed than for the COLD forecasts.

Aspects of Fig. 4 are summarized in Table 2. Of the 30 WARM forecasts at the end of March, the MJO propagates at least as far east as the western Pacific (phase 5) in 29 of them, whereas 9 reach phase 6 and 2 reach phase 7. In only one WARM forecast does MJO activity stall in the Maritime Continent (phase 4). With respect to the COLD forecasts, only 11 forecasts exhibit MJO propagation into the western Pacific (phase 5) and in only 1 case does the MJO activity reach phase 6. The propagation of the MJO in 16 COLD forecasts is stalled in phase 4, whereas in the other 2 cases the propagation is limited to phase 3.

The convective activity associated with the first MJO event in each set of forecasts originates in the eastern Indian Ocean rather than the western-central Indian Ocean, which is the usual origin for observed MJOs. This partially reflects the bias of the model’s MJO, which does not show enough amplitude in the western and central Pacific (e.g., cf. the structure of the MJO in Fig. 3 with the equivalent plot in WH04). We think this might also be linked to the ensemble-mean AMIP atmospheric initial conditions not being in balance with the oceanic initial condition, as subsequent MJO events can be traced farther back into the western Indian Ocean in accord with observations.

c. Structure of the intraseasonal variability

To better demonstrate differences in the eastward propagation of the initial MJO event in each of the

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<td>30 WARM cases</td>
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<td>30 COLD cases</td>
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Fig. 4. The phase space plots for the combined OLR and U10 anomalies, which are projected onto the leading two EOFs (see Fig. 3). Each point represents the 5-day averaged (pentad) location of two corresponding principal components (PC1 and PC2). The higher-frequency variability has been removed by a 15-day running-mean filter [e.g., the start point (S) and end point (*) of each forecast case represents the averaged location over days 21–25 and 86–90, respectively]. The dashed (solid) curves indicate the propagation of the MJO during days 21–60 (61–90). The approximate locations of the MJO cycle in the phase space (e.g., the central-western Pacific for phases 6 and 7) are labeled similarly to WH04, noting that we have transposed PC1 and PC2. The plots show 30 individual (a) WARM cases and (b) COLD cases.
WARM and COLD forecast sets, we construct MJO composites of various variables. The composites were constructed relative to when the first minimum of bandpass-filtered OLR, averaged over 10°S–10°N, passed over the base longitude (140°E) during the first 2 months of the forecast. This is referred to as day 0. The average date of these base points (day 0) of both WARM and COLD sets is around 8 January 1997 (approximately 38 days after the forecast initial date, 1 December 1996). The bandpass filter retained eastward-propagating wave-numbers 1–4 with periods of 20–120 days.

The lag time–longitude composite sections of the OLR and U10 anomalies are shown in Fig. 5. For both WARM and COLD forecasts, convective activity develops over the eastern Indian Ocean from days −40 to −10, although a little bit stronger for the COLD forecasts. After that, the convective anomalies propagate eastward with phase speed of around 4–5 m s\(^{-1}\) (see the dashed line with the arrow in Figs. 5a,b), which is characteristic of the MJO. Between days −40 and −30, both sets of forecasts show convection in the western Pacific, which subsequently starts to decay and is probably a result of the initial spinup.

Both sets show convection reaching the western Pacific from around day −10 onward. However, the convection in the WARM forecasts penetrates farther than the convection in the COLD forecasts. The convection in the WARM forecasts penetrates past the date line to the eastern edge of the warm pool at about 160°W, a location some previous studies have indicated as a sensitive region for the development of El Niño events. However, the convection associated with the COLD forecasts, while still propagating eastward in a similar manner over the eastern Indian Ocean and Maritime Continent, does not reach the date line, decaying at about 175°E. The eastward extension of the MJO event that occurs in the first few months of the WARM forecasts expands distinctly farther east than that for the COLD forecasts by about 25° longitude along the equator during the same period (e.g., from lag −10 to +25 days). It is also seen in Fig. 5 that the COLD forecasts maintain relatively stronger suppressed convection anomalies to the east of the convection, with a strong OLR gradient between the western and central Pacific throughout the composite period.

The convection patterns are associated with westerly wind anomalies to the west and easterly wind anomalies to the east, which propagate eastward with the convection (Figs. 5c,d). Similar to the convection anomalies, the westerly wind anomalies propagate as far as the date line in the WARM forecasts but only reach around 165°E in the COLD forecasts. Furthermore, the magnitude of the westerly anomalies is stronger in the WARM forecasts than in the COLD forecasts (maximum of around 3 m s\(^{-1}\) for the WARM forecasts compared to around 2 m s\(^{-1}\) for the COLD forecasts). The easterly wind anomalies to the east of the convection are weaker in the WARM forecasts than in the COLD forecasts. The longer fetch and stronger amplitude of the westerly wind anomalies to the west of the convection, together with the reduced easterly wind anomalies to the east of the convection are likely to have a significant impact on the ocean and this will be explored in the next section.

To explore the above results further, we investigate the two-dimensional spatial structure and evolution of MJO-related convective activity and surface wind anomalies captured by the composite analysis. The horizontal evolution of the composite OLR and U10 anomalies from a lag of −20 days to a lead of +20 days in 10-day intervals are illustrated in Fig. 6. Currently, few numerical models can produce the MJO with a spatial structure and propagation characteristics similar to the observed, even though many of them can produce dominant eastward propagation with reasonable zonal scales, periods, and phase speeds (e.g., Zhang et al. 2006). The composite OLR and U10 anomalies for the WARM forecasts (Fig. 6a), however, display similar spatial characteristics to those described in observational studies (e.g., Hendon and Salby 1994; WH04). This includes the evolution of the phasing of the westerly anomalies from being near quadrature with the convection in the eastern Indian Ocean to near in phase with the convection in the western Pacific. We also note that by day +20, westerly anomalies extend across the entire western Pacific. However, a model bias, whereby the convective anomaly tends to develop on both sides of the equator as the MJO propagates into the western Pacific (e.g., Zhang et al. 2006), is apparent.

In contrast, the center of negative OLR anomaly for the MJO in the COLD forecasts (Fig. 6b) does not penetrate as far to the east, although the spatial pattern and strength are roughly similar to the composite MJO of the WARM forecasts. Similarly, the westerly anomalies never extend past about 160°E, whereas strong easterly anomalies are maintained east of 160°E even through day +20. Hence, the initial MJO in the WARM cases leads to westerly anomalies across the entire western Pacific, whereas in the COLD cases the westerly phase of the MJO never makes it past about 160°E.

The analysis above demonstrates a clear relationship between the spread of the 90-member ensemble and the behavior of the MJO during the first 4 months of the forecast, particularly in relation to the eastward extension of convection into the central Pacific. So far, we have not explored cause and effect. We deal with this in
FIG. 5. (a) Lag time–longitude composite of the OLR anomalies for WARM cases in the first 4 months. The base point is selected by the minimum of the bandpass-filtered (periods of 20–120 days and eastward-propagating wavenumbers 1–4) OLR anomalies along 140°E in the first 2 months. The contour interval is 10 W m$^{-2}$, with shaded areas representing negative anomalies (convective activities). The date line is indicated as a dashed line. (b) As in (a), but for COLD cases. (c) As in (a), but for the U10 anomalies. The contour interval is 1 m s$^{-1}$, with shaded areas representing positive anomalies (westerly anomalies). (d) As in (c), but for COLD cases.
the next section by looking at the evolution of ocean SST and thermocline depth during the MJO events.

d. Ocean dynamical response

The downwelling oceanic Kelvin waves forced by the MJO-induced westerly wind over the western Pacific are commonly considered as the linkage through which the MJO impacts ENSO (e.g., Vecchi and Harrison 2000; Zhang and Gottschalck 2002). Anomalies in the 20°C isotherm depth (Z20) averaged over 5°S–5°N are used to represent the vertical change of the thermocline depth associated with the eastward propagation of Kelvin waves. Lag time–longitude sections of Z20 anomalies are constructed based on the same base point of bandpass-filtered OLR anomalies (see section 4b), but we display a lag of 0–80 days to see the impact of the MJO on the ocean (Fig. 7).

The westerly anomalies associated with the MJO (Figs. 5c,d) generate eastward-propagating downwelling Kelvin waves for both WARM and COLD forecasts (Figs. 7a,b), but their amplitude and eastward extension are rather different. The WARM forecasts produce a relatively stronger downwelling Kelvin wave; the amplitude is about 24 m for the WARM forecasts at lag 25 days compared to only about 18 m for the COLD forecasts at the same time. In the COLD forecasts, shallow thermocline anomalies develop in the east Pacific reaching up to −14 m at lag of 25–40 days, mainly because of easterly wind anomalies. The WARM forecasts also develop shallow thermocline anomalies, but only reach −10 m. Stronger downwelling thermocline anomalies develop earlier and farther east in the WARM forecasts compared to the COLD forecasts, mainly because of the downwelling Kelvin wave. By the end of the composite period (lag 80 days) the WARM forecasts show downwelling thermocline anomalies across the whole Pacific, with a maximum near 155°–175°W. However, the COLD forecasts still have upwelling anomalies in the far east Pacific and the maximum downwelling anomalies have only reached around 170°W. This is because the westerly winds penetrate farther into the central Pacific in the WARM forecasts, as discussed earlier.

The downwelling Kelvin wave suppresses equatorial upwelling of relatively cold water, which results in positive SSTAs over the east Pacific (e.g., McPhaden and Yu 1999; Lengaigne et al. 2002; Zhang and Gottschalck 2002). In addition, westerly anomalies lead to eastward surface current anomalies, which in turn can lead to

Fig. 6. (a) Lag time–horizontal composite of OLR (thin contours; interval 14 W m⁻²) and U10 vector anomalies for WARM cases corresponding to Figs. 5a,c. Shaded areas represent negative OLR anomalies (convective activity) that pass the t test at the 95% significant level (bold contours). Black arrows illustrate wind vector anomalies that are statistically significant at the 95% level for the t test and the largest vector is shown on the bottom (m s⁻¹). The lag time interval is 10 days from a lag of −20 to 20 days. The date line is indicated as a dashed line. (b) As in (a), but for COLD cases corresponding to Figs. 5b,d.
relative advection of warm-pool water eastward and warming of the SSTAs in the central Pacific (e.g., Bergman et al. 2001). The time–longitude composite section of the SSTA corresponding to the first MJO event for WARM and COLD forecasts is shown in Fig. 8. The positive SSTA for the WARM forecasts (Fig. 8a), which appears at the central-eastern Pacific (175°E–150°W) after a lag of +55 days, exhibits a maximum greater than 0.75°C. For the COLD forecasts (Fig. 8b) the positive SSTAs are distinctly weaker, reaching only 0.45°C at the end of the composite period and their eastward extension is also restricted to west of the date line. Interestingly, at lags of 0–30 days, the COLD forecasts produce warmer SSTAs in the west Pacific and cooler SSTAs in the east Pacific. This leads to a stronger east–west gradient in the COLD forecasts than the WARM forecasts, thus supporting stronger easterly anomalies in the central Pacific. The COLD anomalies maintain a relatively strong east–west SSTA gradient, with the center of the gradient propagating slowly eastward during the composite period. The WARM forecast, however, shows a rapid weakening of the east–west SSTA gradient from a lag of 30 days as the central and eastern Pacific warm up.

e. Cause and effect

It is often difficult to separate cause and effect; do SST changes impact MJOs or do MJOs lead to SST changes? Figures 5 and 8 show that in the west Pacific, the first MJO during each WARM forecast occurs over relatively colder water than in the COLD forecasts (see the lag of 0–30 days), suggesting that the first MJO in the forecasts is independent of the SST but has an impact on the subsequent SST evolution. The lead–lag relationship is more clearly illustrated by the differences in the composites of the WARM and COLD forecasts. In Fig. 9, we superimpose the difference of the composite zonal wind anomalies (WARM – COLD) on top of the

Fig. 7. (a) As in Fig. 5a, but for the lead time–longitude composite of 20°C isotherm depth anomalies meridionally averaged over 5°S–5°N for the WARM cases, noting that we do not show those composite results of Z20 and SSTA before the base point (day 0) in Figs. 7–9. Contour interval is 2 m with shaded areas representing positive anomalies. (b) As in (a), but for the COLD cases.
differences in the composite Z20 anomalies and the composite SSTAs. Anomalies in Fig. 9 are only displayed where the differences are significant at the 95% level. On the one hand, increased westerly wind anomalies, between 130°E and the date line around lag 0 in the WARM forecasts compared to the COLD forecasts, lead the development of positive thermocline anomalies associated with the downwelling Kelvin waves described previously. On the other hand, there is little difference in the SSTAs before a lead of 210 days. Significant positive SST anomalies (≥0.2°C) warm anomalies only develop in the central Pacific after a lag of 10 days and therefore after the MJO-related westerly wind anomalies start to develop. These results suggest that it is the MJO activity, particularly its greater eastward extension in the WARM forecasts, that leads to increased westerly anomalies, which in turn leads to positive thermocline anomalies and SSTA warming. These results support the mechanism identified by Kessler et al. (1995) and Bergman et al. (2001).

5. Summary

We have examined the impact of stochastic intra-seasonal variability on forecasts of the onset of the 1997/98 El Niño. A 90-member, 9-month forecast ensemble starting on the 1 December 1996 was generated. The experiment was designed so that the ensemble spread was simply a result of internal stochastic variability. All forecasts lead to warm El Niño-type conditions though the amplitude of the warming varied from 0.5° to 2.7°C in the Niño-3.4 region. The evolution of El Niño in the first 4 months had a strong impact on the longer-term evolution; in particular, forecasts that showed strongest SST warming in the central-east Pacific at 9 months, on average, were also the warmest in the east-central Pacific at the end of the first 4 months. Forecasts were separated into terciles based on the Niño-3.4 SSTA at month 4 and the role of intraseasonal variability in the upper WARM tercile and lower COLD tercile was investigated.
All forecasts generated an MJO event in the first few months of the forecasts. Analyses of the propagation and strength of the MJO showed that while both WARM and COLD forecasts were associated with similar MJOs, those in the WARM forecasts propagated farther into the central Pacific.

A composite analysis of the first MJO in each forecast was used to explore the difference in the characteristics of the MJOs in the WARM and COLD forecasts and their impact on the coupled model evolution. This showed that the MJO convective activity in the WARM forecasts expanded farther east, propagating from the eastern Indian Ocean to the eastern edge of the warm pool (~160°W), whereas convective activity associated with the MJOs in the COLD forecasts stalled around 150°E. The surface westerly anomalies associated with the MJO for the WARM forecasts also shifted farther east into the central Pacific (near the date line). Thus, as anticipated, the longer zonal fetch of westerly winds in the WARM forecasts leads to relatively stronger downwelling Kelvin waves and stronger SSTA warming over the central-eastern Pacific. More evidence regarding this issue is provided by forming composites relative to the eastward penetration of the first MJO event (not shown). Those forecasts that demonstrate the greatest eastward penetration are also those forecasts that warm the most. This analysis of the role of the MJO is in good agreement with previous studies (e.g., Kessler et al. 1995; Bergman et al. 2001) that a longer zonal fetch of westerly wind forcing tends to produce a stronger downwelling Kelvin wave warming up the central Pacific, reducing the east–west SST gradient, and allowing the westerly anomalies to shift farther east. Moreover, Marshall et al. (2009), using the same POAMA model in stand-alone climate mode, compared the composite behavior of MJO events during the development of strong El Niños to weak El Niños and found that the strong El Niños were associated with a more zonally expansive region of the MJO during the onset phase of the El Niño. Their results further support our conclusions.

Our results indicate that there is a clear separation of cause and effect: little difference in SST in the west Pacific was apparent between the composites from the WARM and COLD forecasts prior to the MJO reaching the west Pacific. The development of westerly winds in the central Pacific associated with the MJO surface westerlies led to the development of the SSTAs and the thermocline anomalies. In the COLD forecasts, the MJO stalled in the west Pacific, the convection and surface westerlies did not reach the date line, and a strong east–west anomalous SST gradient was maintained across the central Pacific. However, for the WARM forecasts, the MJO and associated westerlies reached the date line, leading to SSTA warming and thermocline deepening in the central Pacific and significant reduction in the anomalous east–west SST gradient across the central Pacific, thereby leading to warmer forecasts at the end of the nine months.

All 90 forecasts developed El Niño conditions, probably because of positive heat-content anomalies across the equatorial Pacific in the ocean initial condition from 1 December 1996. This implies that the MJO or stochastic intraseasonal variability may not be necessary for El Niño development, at least in this case. That said,
all forecasts did develop an MJO event during the first 4 months, indicating that perhaps the background oceanic state and initial atmospheric condition favored MJO development. However, the results show that the stochastic behavior or details of the MJOs that developed during December 1996–March 1997 in the forecasts did have a significant impact on the subsequent strength of the event. In particular, the forecasts with MJOs that extended farther into the central Pacific early in the forecast, on average, lead to a stronger El Niño.

We cannot dismiss the possibility that some of the variation of MJO activity is tied to the state of the coupled system (e.g., Yu et al. 2003; Batstone and Hendon 2005; Eisenman et al. 2005; Vecchi et al. 2006; Hendon et al. 2007). Hence, some of the variation of MJO activity and its subsequent impact on the evolution of El Niño may be predictable. However, our results suggest a limit in our ability to forecast the exact strength of El Niño because the details of individual MJO events matter. Furthermore, these results imply that for coupled model forecasts to produce sufficient ensemble spread that represents realistic uncertainty, the models must not only have a good simulation of the MJO but also be able to capture its propagation into the central Pacific. Coupled models with deficient representations of the MJO or biases in the mean state that affect the propagation of the MJO into the central Pacific, such as the common problem of an enhanced cold tongue, may lead to skewed ensemble distributions.

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