Modeling the Mechanisms of Linear and Nonlinear ENSO Responses to the Pacific Meridional Mode

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ABSTRACT

Interactions between the Pacific meridional mode (PMM) and El Niño–Southern Oscillation (ENSO) are investigated using the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM) and an intermediate coupled model (ICM). The two models are configured so that the CESM simulates the PMM but not ENSO, and the ICM simulates ENSO but not the PMM, allowing for a clean separation between the PMM evolution and the subsequent ENSO response. An ensemble of CESM simulations is run with an imposed surface heat flux associated with the North Pacific Oscillation (NPO) generating a sea surface temperature (SST) and wind response representative of the PMM. The PMM wind is then applied as a forcing to the ICM to simulate the ENSO response. The positive (negative) ensemble-mean PMM wind forcing results in a warm (cold) ENSO event although the responses are not symmetric (warm ENSO events are larger in amplitude than cold ENSO events), and large variability between ensemble members suggests that any individual ENSO event is strongly influenced by natural variability contained within the CESM simulations. Sensitivity experiments show that 1) direct forcing of Kelvin waves by PMM winds dominates the ENSO response, 2) seasonality of PMM forcing and ENSO growth rates influences the resulting ENSO amplitude, 3) ocean dynamics within the ICM dominate the ENSO asymmetry, and 4) the nonlinear relationship between PMM wind anomalies and surface wind stress may enhance the La Niña response to negative PMM variations. Implications for ENSO variability are discussed.

1. Introduction

The dominant mode of variability in the tropical Pacific is the El Niño–Southern Oscillation (ENSO; Wallace et al. 1998) phenomenon. While all ENSO events evolve differently, warm ENSO (El Niño) events are generally characterized by warm sea surface temperature (SST) anomalies in the central and eastern Pacific, a reduction in the climatological zonal trade wind strength in the central tropical Pacific, and a deepening of the thermocline in the central and eastern tropical Pacific (Rasmusson and Carpenter 1982; McPhaden et al. 1998; Wallace et al. 1998). Although cold ENSO (La Niña) events are generally characterized by the opposite structure, warm and cold ENSO events are not symmetric opposites.1 In particular, cold ENSO events tend to be weaker in amplitude overall, and tend to have relatively larger amplitude in the central Pacific compared to the eastern Pacific (Larkin and Harrison 2002). It is generally thought that ocean dynamics play a critical role in the evolution of ENSO events through their role in coupled ocean–atmosphere interactions in the tropical Pacific (Neelin et al. 1998), although variability with ENSO-like characteristics can be simulated without ocean dynamics (Clement et al. 2011).

The second most dominant statistical mode of coupled ocean–atmosphere variability in the tropical Pacific is the Pacific meridional mode (PMM; Chiang and Vimont 2004). The positive phase of the PMM is characterized...
by a northward meridional SST gradient around the location of the climatological intertropical convergence zone (ITCZ), a corresponding northward shift of the climatological ITCZ, and meridional winds that blow toward warmer water (the negative phase contains the opposite structure). Unlike ENSO, the PMM does not require ocean dynamics for its existence and instead evolves via a positive feedback between wind-induced evaporation and SST—the so-called WES feedback (Xie and Philander 1994; Chang et al. 1997; Vimont 2010). In short, positive SST anomalies drive an atmospheric circulation that includes a reduction in the climatological trade wind strength, which reduces the climatological evaporative heat flux over the original positive SST anomalies.

It has been shown that the PMM and ENSO are related through the so-called seasonal footprinting mechanism (SFM) (Vimont et al. 2003b; Chang et al. 2007; Alexander et al. 2010; Park et al. 2013). The SFM operates as follows: midlatitude atmospheric variability associated with the atmospheric North Pacific Oscillation (NPO) (Rogers 1981; Linkin and Nigam 2008) generates positive SST anomalies in the eastern subtropical Pacific through a relaxation of the climatological trade winds. These SST anomalies have a strong projection on the spatial structure of the PMM. The SST anomalies persist beyond the time scale of NPO variations, and subsequently propagate equatorward and westward via air–sea feedbacks (Liu and Xie 1994; Vimont et al. 2009; Vimont 2010) in a manner very similar to the PMM. The SST anomalies generate an atmospheric circulation that can lead to the development of an ENSO event during the ensuing months (Vimont et al. 2003a,b; Chang et al. 2007; Alexander et al. 2010). While there is some discussion in the literature about the equivalence of the SFM and the PMM, for the purposes of this paper we assume that the two are intimately related, as the dynamical processes that govern PMM evolution provide the conduit by which midlatitude atmospheric variability can impact ENSO through the SFM.

Despite evidence for the PMM’s influence on ENSO, the specific mechanisms that link the PMM to ENSO are not clear. Observational (Vimont et al. 2003b) and model (Vimont et al. 2003a; Alexander et al. 2010) analyses suggest that the link occurs through the direct forcing of equatorial Kelvin waves by wind stress anomalies associated with the PMM, which lead to a deepening of the thermocline in the eastern Pacific. In a different view, recent studies suggest that changes in the trade winds induce heat content anomalies in the equatorial region that lead to ENSO variation through “trade-wind charging” (Anderson et al. 2013; Anderson and Perez 2015) or through a “discharge–recharge” paradigm (Jin 1997; Anderson and Maloney 2006). Specifics of how the PMM interacts with ENSO are important due to the seasonal timing of the PMM and the seasonality of the ENSO growth rate. The PMM’s influence on ENSO is likely to be most prominent if that influence occurs at a time when ENSO growth rates are largest (the seasonal ENSO growth rate is related to a potential instability due to equatorial upwelling in regions of vertical thermocline perturbations; Battisti 1988). This possibility, however, has not been explored.

Many recent studies have documented the role of the extratropics on the spatial characteristics of ENSO since two types of ENSO events have been defined. This may have large implications on global climate since these ENSO events, termed central Pacific (CP) and eastern Pacific (EP), produce significantly different teleconnection patterns and climatic impacts (Larkin and Harrison 2005; Kumar et al. 2006; Ashok et al. 2007; Weng et al. 2007, 2009; Kao and Yu 2009). While any individual ENSO event may contain characteristics of both types of event (Kug et al. 2009; Takahashi et al. 2011; Vimont et al. 2014), extratropical atmospheric forcing, such as the NPO, has been suggested to influence the development of CP ENSO events (Alexander et al. 2010; Yu and Kim 2011; Kim et al. 2012; Vimont et al. 2014). Additionally, Lin et al. (2015) show that the PMM is important to the generation of CP ENSO events through the SFM, although preexisting thermocline conditions may influence this relationship (Capotondi and Sardeshmukh 2015). Despite the connection between the PMM and CP ENSO events, the main goal of this study is not to directly simulate the PMM’s role in ENSO diversity, but rather to better understand mechanisms by which the PMM interacts with ENSO variability in general.

Identifying the specific processes and mechanisms through which extratropical atmospheric forcing, especially the NPO, influences the development of ENSO events requires further attention. Understanding the interactions between the PMM that contribute to the evolution of ENSO events can guide predictability studies of ENSO events. For example, Larson and Kirtman (2014) show the PMM is an important precursor to ENSO although it is not, yet, a reliable predictor. In this study we use a combination of two models containing complementary sets of physics in order to separate and characterize the interactions between the extratropics and ENSO development. The National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM) in “slab ocean” configuration contains the necessary air–sea interactions to
generate a response representative of the PMM to a surface heat flux associated with the North Pacific Oscillation, but is unable to fully simulate ENSO due to the lack of ocean dynamics. An intermediate coupled model (ICM) (Battisti 1988), which contains the necessary ocean dynamics to simulate ENSO variability but not the PMM (due to the lack of wind-induced evaporation), is used to simulate the ENSO response to the atmospheric forcing generated by CESM. It is important to note that the ICM used is unable to simulate ENSO diversity with respect to CP and EP events, which is a caveat for this study. However, here we are interested in using the ICM to isolate the mechanisms by which PMM variability influences ENSO-like variability in general. The thermodynamic air–sea interactions contained in CESM and the ocean dynamics in the intermediate coupled model cleanly separate the two primary mechanisms responsible for linking PMM and ENSO events, allowing us to identify the different processes responsible for the simulated ENSO response.

This paper is organized as follows. The CESM model description, NPO forcing, and PMM response to the NPO forcing are described in section 2. Section 3 describes the ICM and presents the ENSO response of the ICM to the PMM atmospheric forcing (generated by the CESM). Section 4 identifies the nonlinear sources generating asymmetry between warm and cold ENSO responses, and section 5 summarizes and discusses the results.

2. CESM1.2

The PMM response to the NPO forcing is simulated using the National Center for Atmospheric Research Community Earth System Model. This study simulates the PMM as opposed to using the observed PMM in order to cleanly isolate PMM dynamics from ENSO dynamics. Isolating the PMM from ENSO is difficult in the observed record since the PMM evolution is linked to the subsequent evolution of ENSO, and hence lagged inferences about the PMM evolution would be “contaminated” by ENSO mechanisms such as the Bjerknes feedback and thermocline influences. This section describes the experimental setup, NPO heat flux forcing, and PMM response to the NPO forcing.

a. Model description and experimental setup

Experiments utilizing the NCAR CESM are performed to determine the response from a surface heat flux forcing associated with the NPO. The model is run with approximately a 2° × 2° horizontal grid (f19_g16 resolution), 30-layer vertical resolution, Community Atmosphere Model version 5 (CAM5) physics, prescribed modal aerosols, and coupled to a slab ocean model (SOM). This configuration contains the ocean–atmosphere thermodynamic coupling necessary to simulate PMM variability through the SFM (Vimont et al. 2001); however, the SOM is unable to simulate ENSO variability since it contains no ocean dynamics. The SOM calculates the change in mixed layer temperature as

\[
\rho c_p h_{\text{mix}} \frac{dT_{\text{mix}}}{dt} = F_{\text{net}} - Q_{\text{flx}},
\]

where \(\rho\) is the density of ocean water, \(c_p\) is the specific heat, \(h_{\text{mix}}\) is the depth of the mixed layer, \(T_{\text{mix}}\) is the temperature of the mixed layer (assumed to be the same as the surface temperature), \(F_{\text{net}}\) is the net surface heat flux, and \(Q_{\text{flx}}\) is a prescribed seasonally varying term that represents the horizontal and vertical oceanic heat flux divergence in the mixed layer column.

We repeat the model experiments that were performed in Vimont et al. (2009, hereafter VAF2009) in which the CESM is forced by surface heat flux anomalies associated with the NPO (described below), except that here the observed NPO heat flux and an updated version of the atmospheric model (CAM5) are used. Results are only briefly described here; for further analysis, see VAF2009.

The NPO net surface heat flux anomalies are calculated as follows. Empirical orthogonal function (EOF) analysis is applied to intraseasonal sea level pressure (SLP) anomalies in the North Pacific (20°–90°N, 110°E–70°W) using the NCEP reanalysis between 1948 and 2002. The intraseasonal filter is constructed by removing the boreal winter [November–March (NDJFM)] mean for each year from the corresponding monthly values during that year, effectively removing interannual variability. The intraseasonal filter is applied so that the calculation of the net surface heat flux associated with the NPO will be less influenced by the resulting change in SST. Since the change in SST will be calculated directly by the model, it should not be incorporated into the externally applied heat flux forcing. The NPO is defined as the second EOF of boreal winter (NDJFM) intraseasonal sea level pressure anomalies. The NPO is defined using NCEP reanalysis data since the second EOF of a long control CESM simulation (with climatological SST) does not represent the NPO pattern well (not shown). The NPO surface heat flux is estimated by regressing the intraseasonally filtered net surface heat flux (calculated from NCEP reanalysis) onto the standardized second principal component of SLP. We define the positive phase of the NPO (+NPO) as positive SLP anomalies over Alaska and the North Pacific and negative SLP anomalies over the subtropical Pacific.
Figure 1 shows the sea level pressure and net surface heat flux (shading) associated with the positive phase of the NPO (defined as the second EOF of intraseasonal SLP in the North Pacific). The SLP contour interval is 0.5 mb (std dev)\(^{-1}\). Solid (dashed) contours correspond to positive (negative) values. The zero contour has been omitted. Downward heat fluxes are defined as positive. The negative phase of the NPO is the opposite polarity.

The seasonal progression of the ensemble-mean low-level wind and sea surface temperature response of the CESM to positive and negative NPO forcing is shown in Fig. 2 (left and right columns, respectively) and largely matches the results found in VAF2009. The +NPO forcing generates an anomalously warm SST structure extending southwestwards from the eastern subtropical Pacific toward the central tropical Pacific. This SST structure, characteristic of the PMM, propagates farther into the tropics and reaches its peak magnitude during the boreal spring (MAM) following the wintertime NPO forcing.

To show the relative timing of the ensemble-mean PMM response to the NPO forcing, we plot the time evolution of the applied NPO forcing and ensemble-mean PMM index (normalized by their peak amplitudes) in Fig. 3. The NPO index shows the constant forcing applied to the CESM during boreal winter (NDJFM). Linear ramp-up and ramp-down of the forcing are evident from October–November and March–April, respectively. The PMM index is calculated by projecting the SST pattern simulated by the CESM onto the spatial pattern of the PMM [defined by applying maximum covariance analysis to non-ENSO tropical SST and 10-m winds over the time period 1950–2005; see Chiang and Vimont (2004) for details]. The resulting ensemble-mean time series in Fig. 3 is normalized by its maximum amplitude. The simulated PMM SST time series peaks in April (year 1) immediately after the NPO forcing is shut off, consistent with a simple integration of the applied forcing. We note the simulated wind time series peaks around the same time or slightly later than the SST response.

Because of thermodynamic coupling, the PMM SST anomalies persist into late summer and fall (JJA–SON), long after the NPO forcing has been removed (Figs. 2e–j). They slowly decay through the following winter (DJF). As in VAF2009, westerly wind anomalies in the western and central tropical Pacific are generated during the boreal summer (JJA) and fall (SON) due to the persistent warm SST anomalies (Figs. 2e,g). The –NPO forcing generates a wind response that is approximately a symmetric opposite to the +NPO simulation. Cold SST anomalies extend southwestward from the eastern subtropical Pacific and propagate into the tropics through spring (MAM), where they persist into the summer and fall (JJA–SON). The cold SST anomalies in the tropics initiate easterly wind anomalies in the western and central tropical Pacific during summer.
(JJA) and fall (SON). The asymmetry of the ensemble mean CESM response to positive and negative NPO heat flux was investigated and was found to be small compared to the symmetric response (results not shown). The symmetry of the atmospheric response will be addressed in the context of ENSO variability in section 3b.

In the slab ocean model the evolution of SST can only be influenced by changes in the surface heat flux [Eq. (1)], which also largely matches the response found in VAF2009. Here, we only compare the response to the +NPO forcing results to the results obtained by VAF2009 since the response to the −NPO is largely the opposite sign. The CESM shows a decrease in downwelling shortwave radiation and an increase in precipitation in the western and central tropical Pacific during the spring (MAM) following the positive NPO forcing. The response during the summer (JJA) is similar, although shifted slightly northward. The latent heat flux response to the +NPO, similar to the results in VAF2009, shows decreased upward surface latent heat fluxes in the western tropical Pacific that expand to the central Pacific subtropics through the summer, whereas increased upward latent heat fluxes are located in the

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**FIG. 2.** Seasonally averaged ensemble-mean CESM SST (shaded) and low-level wind (vectors) anomalies due to (left) positive and (right) negative NPO surface heat flux forcing. Values are only plotted where they exceed the 95% confidence level based on a two-tailed univariate (or bivariate in the case of vectors) t test.
The PMM wind anomalies from the CESM simulations during boreal winter (NDJFM). The PMM index is calculated by projecting the ensemble-mean SST pattern simulated by the CESM onto the spatial pattern of the PMM defined by Chiang and Vimont (2004) and represents the ensemble-mean temporal evolution of the PMM forcing for the ICM. The PMM index is normalized by the maximum amplitude. The ENSO index (Niño-3.4 index) represents the ENSO response to the 20 month ensemble-mean PMM wind forcing.

Fig. 3. Timeline showing the relative occurrence of the NPO heat flux forcing (solid line), the PMM (dashed line), and the resulting ENSO event (stars); see text for details. The NPO index shows constant maximum forcing applied to the 20-month CESM simulations during boreal winter (NDJFM). The PMM index is calculated by projecting the ensemble-mean SST pattern simulated by the CESM onto the spatial pattern of the PMM defined by Chiang and Vimont (2004) and represents the ensemble-mean temporal evolution of the PMM forcing for the ICM. The PMM index is normalized by the maximum amplitude. The ENSO index (Niño-3.4 index) represents the ENSO response to the 20 month ensemble-mean PMM wind forcing.

3. Intermediate coupled model

The PMM wind anomalies from the CESM simulations, as described in the previous section, are used as an external forcing to an intermediate coupled model that contains the necessary physics to simulate ENSO, but not PMM variations. In this section, we present the results from the ICM experiments.

a. Model description and experimental setup

The ICM used in this study is the Battisti (1988) version of the Zebiak and Cane (1987) intermediate coupled model. Realistic changes were made to three key model parameters (Thompson and Battisti 2001): the model’s ocean damping rate (changed from 2.5 yr⁻¹ to 1 yr⁻¹), the surface drag coefficient (changed from 3.2 × 10⁻³ to 2.0 × 10⁻³), and Rossby wave reflection efficiency (reduced from 100% to 80%). The primary result of these parameter alterations is to produce a linearly stable ENSO regime in which the model requires external forcing to generate ENSO variability.

The ICM simulates the ENSO response to the PMM atmospheric forcing (as described in the previous section). This ICM contains the necessary ocean dynamics to simulate ENSO variability but does not contain the necessary air–sea thermodynamic interaction (i.e., the WES feedback mechanism) needed to simulate the PMM. The two-layer atmosphere is governed by linear shallow water equations on a beta plane and is forced by atmospheric heating anomalies that result from wind convergence and evaporation from SST anomalies (Zebiak and Cane 1987; Battisti 1988; Gill 1980). The rectangular, reduced gravity ocean model (30°–30°S, 124°E–80°W) contains an upper layer governed by linear shallow water wave dynamics and a motionless deep layer. The upper layer includes a fixed depth (50 m) surface mixed layer that determines SST through a temperature tendency equation (Battisti 1988). The fully coupled ICM internally calculates the anomalous pseudowind stress vector on the ocean due to external forcing and coupling as follows:

\[ \tau = u |u| - u_m |u_m|, \]  

(2)

where \( |u| \) is the amplitude of the wind vector \( u \), and the total wind \( u \) is the sum of the mean wind \( (u_m) \) (this includes the annual cycle), an externally applied wind forcing if applicable \( (u_e) \), and the internal ICM coupled response \( (u_i) \):

\[ u_i = u_m + u_e + u_e . \]

(3)

To determine the ENSO response to the PMM, we force the ICM with CESM-simulated wind response to the NPO forcing. The full 20 months (October through May) of CESM-simulated wind anomalies, which capture the growth and decay of the PMM, are applied to the ICM to determine the ENSO response to a realistic PMM pattern and evolution. We apply the CESM forcing \( (u_e) \) as either the anomalous ensemble-mean wind (as in Fig. 2), or as each of the 40 individual ensemble member wind anomalies from both the +NPO and −NPO forced CESM experiments. The wind anomalies are calculated as the difference between the ensemble mean (or individual member) and the mean of the CESM control simulation in order to retain nonlinearities in the wind forcing. Hereafter, the imposed CESM wind will be referred to as the external “PMM forcing” for the ICM, with the response to the +NPO (−NPO) forced CESM experiments referred to as +PMM (−PMM) forcing.

Four versions of the ICM are run for each external forcing: a fully coupled model, an uncoupled ocean-only model, a model in which only oceanic Rossby waves respond to the external forcing (RW model), and a model in which only oceanic Kelvin waves respond to the external forcing (KW model). The fully coupled model allows for full air–sea interactions and atmospheric feedback to influence the oceanic response.
uncoupled model removes all atmospheric feedback to the ocean (\(u_c\) is set to zero), thus allowing the ocean to only respond to the external forcing. The RW model and KW model allow the external forcing (\(u_f\)) to excite only oceanic Rossby or Kelvin waves, respectively, while permitting the coupled wind anomalies (\(u_c\)) to interact fully with the ocean model. Note that in the Battisti (1988) ICM the shallow water ocean model state is projected into Kelvin and Rossby wave space and the state is evolved via wave characteristics. Thus, in the RW model \(u_f\) is set to zero in Eq. (3) for the Kelvin wave forcing but retained for the Rossby wave forcing and vice versa for the KW model. Note that \(u_c\) is retained for both routines in both models so that once an ENSO event is initiated, it can evolve via both Kelvin and Rossby wave propagation. An additional set of sensitivity experiments is run to investigate the source of asymmetry in the ENSO response to PMM forcing, which will be described in section 4.

A wide range of sensitivity experiments are run using various combinations of the PMM external forcing and ICM physics. These experiments (numbered 1 through 10) are summarized in Table 1. Readers should refer to Table 1 for all references to experiments. For all experiments the phrases “symmetric” and “asymmetric” refer to the degree to which the model response to positive and negative forcing is similar in amplitude but with reverse polarity (symmetric) versus not-similar even with reverse polarity (asymmetric).

### Table 1. Summary of intermediate coupled model experiments.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Forcing</th>
<th>Result</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Full ICM</td>
<td>± Ensemble-mean PMM wind</td>
<td>+PMM (\Rightarrow) warm ENSO, −PMM (\Rightarrow) cold ENSO</td>
<td>Greater amplitude response to +PMM</td>
</tr>
<tr>
<td>2) Uncoupled</td>
<td>As in Exp. 1</td>
<td>As in Exp. 1</td>
<td>Weaker amplitude response than in Exp. 1</td>
</tr>
<tr>
<td>3) Full ICM: Ensemble</td>
<td>± PMM wind of individual ensemble members</td>
<td>Ensemble mean as in Exp. 1. Individual runs vary widely</td>
<td>As in Exp. 1, but role of natural variability is important</td>
</tr>
<tr>
<td>4) RW model: Forcing only projects onto Rossby modes</td>
<td>As in Exp. 1</td>
<td>+PMM (\Rightarrow) no ENSO, −PMM (\Rightarrow) warm ENSO</td>
<td>PMM forced Rossby waves do not strongly influence fully coupled response</td>
</tr>
<tr>
<td>5) KW model: Forcing only projects onto Kelvin modes</td>
<td>As in Exp. 1</td>
<td>As in Exp. 1, but greater amplitude</td>
<td>PMM forced Kelvin waves dominate fully coupled response</td>
</tr>
<tr>
<td>6) Seasonally shifted forcing: Coupled and uncoupled ICM</td>
<td>+PMM wind shifted by 0, 3, 6, or 9 months</td>
<td>All produce warm ENSO events, but largest amplitude for lag 0 and lag +9 months</td>
<td>Seasonality of ENSO growth rate influences amplitude of response</td>
</tr>
<tr>
<td>7) Linear wind</td>
<td>+PMM wind, −1* (+PMM wind)</td>
<td>As in Exp. 1</td>
<td>As in Exp. 1</td>
</tr>
<tr>
<td>8) Linear stress</td>
<td>+PMM stress, −1* (+PMM stress)</td>
<td>As in Exp. 1</td>
<td>As in Exp. 1</td>
</tr>
<tr>
<td>9) Linear wind and thermocline</td>
<td>As in Exp. 7</td>
<td>+PMM event has lesser amplitude than −PMM event</td>
<td>When thermocline effects removed, other nonlinearities generate larger cold ENSO events</td>
</tr>
<tr>
<td>10) Linear stress and thermocline</td>
<td>As in Exp. 8</td>
<td>ENSO response is nearly linear</td>
<td>Nonlinear ENSO response, as in Exp. 1, is dominated by thermocline</td>
</tr>
</tbody>
</table>

**b. ENSO response to PMM forcing**

The fully coupled (experiment 1; Table 1) ICM Niño-3.4 index (N34) response to the positive and negative ensemble-mean external forcing is shown as black solid lines in Figs. 4a and 4b, respectively. Note that the fully coupled ENSO response (black solid line, Fig. 4a) is normalized by its peak value and plotted in Fig. 3 to illustrate the sequencing of the NPO, PMM, and ENSO in our experimental setup. Although the PMM is nearly symmetric (Fig. 2), the ENSO response generated by this forcing is not (Fig. 4). The sources of asymmetry will be discussed in detail in section 4. The fully coupled ICM N34 index response to the ensemble-mean wind forcing (black) indicates that positive (negative) PMM forcing generates a large warm (cold) ENSO event in boreal fall, approximately one year after the NPO forcing is applied. The N34 index exceeds 1°C during the warm ENSO event. The cold ENSO event, however, is about half the magnitude of the warm event with the N34 index reaching only −0.5°C. The fully coupled ICM forced with −PMM also generates a warm ENSO event of similar magnitude to the initial cold ENSO event, in the
year following the initial cold event. This rebound event of the opposite sign is not seen in the PMM forced simulations, which simply decay throughout the year following the peak of the initial event.

The uncoupled ICM simulations (green lines, Fig. 4) show the uncoupled oceanic N34 index response to only the PMM forcing since the Bjerknes feedback is removed (experiment 2). The externally forced response is about half the amplitude of the fully coupled response and decays to zero earlier than the fully coupled ENSO events. The reduced magnitude of the response is especially apparent in the +PMM forced simulations (Fig. 4a). In contrast to the coupled case, the uncoupled response to the −PMM (Fig. 4b) does not generate a warm rebound event following the initial cold event, suggesting that the warm rebound is not generated by the PMM forcing but rather is a result of the coupled response.

The anomalous SST and thermocline structure of the fully coupled ENSO response due to the ensemble-mean +PMM and −PMM forcing is shown in Fig. 5. Since we are primarily interested in the mechanisms responsible for ENSO generation, we only show the seasons leading up to the peak of the ENSO event. The fully coupled ICM shows a warm (cold) ENSO event peaking in SON(1), approximately one year after the positive (negative) NPO forcing is initially applied. However, the magnitude of the cold SST anomalies is weaker than the warm event and the largest cold SST anomalies are concentrated in the eastern Pacific, whereas the warm event SST anomalies extend farther west.

The seasonal evolution of the anomalous thermocline response, relative to climatological thermocline depth specified in the model, to the PMM forcing is shown as black contours in Fig. 5. As early as MAM(1), thermocline anomalies emerge in the northwestern equatorial Pacific, between 0° and 15°N, and about 120° and 180°E. These anomalies appear prior to substantial SST anomalies, so are likely a direct Rossby wave response to the forcing. By JJA(1) the eastern equatorial Pacific experiences a deepened (shoaled) thermocline for the +PMM (−PMM) forced simulations. Despite the asymmetric response of eastern equatorial Pacific SST by JJA(1) the thermocline anomalies are largely symmetric until SON(1) and beyond.

To determine the robustness of the response to the ensemble-mean PMM forcing we analyze the spread of the N34 index responses across the 40 ensemble members to +PMM and −PMM wind forcing for the fully coupled ICM (experiment 3). The black lines in Fig. 6 show the response of the ICM to the ensemble-mean PMM wind forcing (as in Fig. 4). Although the ENSO response to the ensemble-mean PMM forcing clearly suggests the importance of the extratropical atmospheric forcing to ENSO generation, the large spread of the individual ensemble responses indicates the ICM is highly sensitive to perturbations in the external PMM wind forcing. To estimate the spread of the ENSO responses we calculate the month containing the largest standard deviation about the mean of all individual ensemble members for both the +PMM and −PMM forced simulations, which are 1.07°C during October and 0.87°C during November, respectively. The conclusions are not sensitive to the month used in the spread calculation. Because of the large spread among ensemble responses, especially for the +PMM case, we can conclude that natural variability, which is captured by the unforced, internal variability contained within the CESM, likely has a large influence on the formation of any given ENSO event due to its impact on the simulated PMM wind structure. Still, the average response of all individual ICM ensemble members (not shown) is

![Fig. 4: The Niño-3.4 index response of the ICM to the ensemble-mean PMM wind forcing. Top (bottom) figure shows the response to the positive (negative) PMM wind forcing. The Niño-3.4 index responses of the fully coupled model (black), uncoupled model (green), Rossby wave model (red), and Kelvin wave model (blue) are shown for each forcing. Timing of the boreal winter NPO is indicated with gray shading.](https://journals.ametsoc.org/doi/abs/10.1175/JCLI-D-16-0090.1)
very similar to the ICM response to the ensemble-mean PMM forcing (black lines), and there is a clear tendency for individual +PMM forcings to result in warm ENSO events (red lines, Fig. 6a), while individual −PMM forcings result in cold ENSO events (blue lines, Fig. 6b).

c. Rossby wave and Kelvin wave responses

The effectiveness of PMM wind anomalies in exciting Rossby and Kelvin waves is important because it provides guidance for what physical processes to monitor for understanding PMM–ENSO relationships. Also, both the PMM wind anomalies and ENSO growth rates experience strong seasonality. As a result, the timing of the SST response to PMM forcing (which will appear in the eastern Pacific much more rapidly for Kelvin wave forcing than for Rossby wave forcing) may depend on details of how the forcing excites an equatorial response.

The SST response in the RW model (experiment 4) illustrates the evolution of the ICM when the external forcing is only allowed to excite equatorial Rossby waves. The RW model does not generate an ENSO event for the positive forcing case (red curve, Fig. 4a). However, the negative forcing excites a large persistent warm ENSO event (red curve, Fig. 4b). The seasonal SST and thermocline response in the RW model is shown in Fig. 7. During the early seasons [MAM(1) and JJA(1)] both +PMM and −PMM forcing produce thermocline anomalies in the northwestern tropical Pacific, as in experiment 1. Early in the season, these thermocline anomalies have nearly symmetric structures.
between the positive and negative forced experiments. However, the SST response in the eastern Pacific is very different for the two simulations. In particular, the $^1$PMM forcing produces very weak negative SST anomalies in the eastern equatorial Pacific whereas the $^2$PMM forcing produces larger and more widespread positive SST anomalies from JJA(1) through DJF(2); this discrepancy is discussed within the context of thermocline–SST feedbacks in section 4. The larger and more widespread anomalies in the negative forcing simulation are more likely to excite coupled feedbacks during JJA(1) and SON(1), at which time ENSO growth is most pronounced. In either case, the response to Rossby wave forcing tends to counteract the fully coupled response (Fig. 4).

The KW model simulations (experiment 5, blue curves in Fig. 4) indicate that Kelvin wave excitation from the $^+PMM$ ($^-PMM$) forcing dominates the generation of the warm (cold) ENSO events. The seasonally averaged SST and thermocline structures of the externally forced Kelvin wave response are shown in Fig. 8. As with the RW model, the thermocline response is relatively symmetric during MAM(1) and JJA(1), and starts to diverge after SON(1), especially in the western equatorial Pacific. The associated SST anomalies in the eastern equatorial Pacific have larger amplitude for the $^+PMM$ than for the $^-PMM$ forcing. In the western Pacific, thermocline anomalies for the KW model simulations are far more confined to the equator than the structures seen in experiments 1 and 3, especially during SON(1). Western Pacific thermocline anomalies in the KW model experiments can only be caused by the internally generated coupled wind anomalies in the KW model (as $u_f$ does not force Rossby waves in the KW model). The meridionally broader structure in the fully coupled experiments (experiments 1 and 3) suggests that PMM wind anomalies ($u_f$) force higher-order Rossby waves, which are less effective (via smaller reflection coefficients) at generating Kelvin waves that ultimately generate the ENSO response in the eastern equatorial Pacific.

d. Seasonality

The role of the PMM in generating ENSO depends both on the direct response to PMM forcing and to the ENSO growth rate, the latter of which is largest during boreal summer and fall (Battisti 1988). To investigate the dependence of the ENSO response to the seasonality of the PMM, we ran the ICM with the positive PMM wind forcing applied with a 0-month (identical to experiment 1), 3-month, 6-month, and 9-month lag (experiment 6). The original (lag 0) PMM forcing refers to the observed CESM timing of the PMM, where the PMM amplitude peaks in boreal spring (MAM) (Fig. 3 shows the timing of the original PMM forcing). For the lag +3 month run the same sequence of CESM-generated PMM wind forcing is applied to the ICM, but it is applied 3 months after the CESM timing [i.e. the PMM peaks during summer (JJA)]. We lag the wind forcing rather than rerunning CESM with lagged NPO forcing so that we can use an identical wind forcing in the ICM to cleanly separate the effect of seasonality in the ENSO growth rate from seasonality of the PMM. In a similar way, the lag +6 month run refers to the PMM peaking in fall (SON), and the lag +9 month run refers to the peak of the PMM occurring in winter (DJF).

Figure 9 shows the N34 index responses of the uncoupled (top) and fully coupled ICM (bottom) to the lagged $^+PMM$ forcing. The solid line indicates the ENSO response to the original PMM forcing (lag 0), the dashed line shows the response to the PMM forcing applied at lag +3 month, the dotted line shows the
response to lag +6 month PMM forcing, and the dot-dashed line shows the response to lag +9 month PMM forcing. The uncoupled ICM simulations (Fig. 9a) all result in similar magnitude ENSO events with N34 index peaking at 0.5°C. However, the N34 index peaks in January for the lag +3 month forcing, in March for the lag +6 month forcing, and in July for the lag +9 month forcing. The shift in the timing of the uncoupled ENSO event is consistent with the lag of the applied PMM forcing and indicates that the direct response does not depend strongly on seasonality of the forcing. Kelvin and Rossby wave model simulations with lagged forcing also support this result (not shown).

The fully coupled simulations show the importance of the seasonality of the ENSO growth rate due to coupled interactions with the seasonally varying mean state (Fig. 9b). The lag +9 month and lag 0 simulations indicate the strongest ENSO response occurs when the Kelvin wave signal reaches the eastern Pacific in boreal summer and fall, respectively, corresponding to the timing of large potential instability for ENSO (Battisti 1988). In terms of thermocline–SST relationships, this large instability corresponds to boreal summer and fall, when zonal SST gradients are largest across the Pacific, and the thermocline in the eastern equatorial Pacific is nearest to the surface (Zhu et al. 2015). The weakest ENSO response occurs when the Kelvin wave signal arrives in the eastern Pacific during boreal spring (lag +6 month simulation; Fig. 9b), which corresponds to a minimum in oceanic instability and ENSO growth potential. At that time, mean zonal wind and mean zonal SST gradients are weak, and the thermocline is farthest

![Figure 7](image-url)
from the surface in the eastern Pacific, reducing potential ENSO growth (Zhu et al. 2015). While the lagged PMM forcing represents a physically unrealistic situation, since the PMM has an observed seasonality, the results are instructive since they demonstrate that the timing of the PMM as it occurs in nature is best suited to generate ENSO growth.

4. Sources of asymmetry

This section identifies the sources of ENSO asymmetry simulated by the ICM. The four potential sources producing the asymmetry within our experiments include asymmetry in the PMM forcing; the atmospheric heating calculation within the ICM, which includes an iterative convergence feedback and nonlinear dependence on regions of mean state moisture convergence (Zebiak and Cane 1987); the nonlinear calculation of “pseudostress” from wind within the ICM [Eq. (2)]; and oceanic dynamics, primarily the effect of upwelling on an anomalously shallow or deep thermocline. In this section we systematically remove each of these nonlinear processes to investigate the influence of each on the asymmetric ENSO response to the PMM forcing. We note that of the various sources of asymmetry, the atmospheric heating calculation and the relationship between subsurface temperature and thermocline depth are very dependent on parameterizations within the ICM, while the wind and pseudostress calculations are not (i.e., asymmetric results due to asymmetric wind forcing or nonlinear pseudostress calculations may be more generally applicable in nature). This will be addressed in more detail in section 5.
First, the role of asymmetry in the ensemble-mean CESM simulated PMM wind is examined. In calculating the wind stress [Eq. (2)], the externally applied wind forcing $\mathbf{u}_f$ was replaced with either $+1$ times the positive PMM wind anomalies (identical to the positive forcing in experiment 1) or $-1$ times the positive PMM wind anomalies. Thus, the fully coupled ICM is forced with linear opposite PMM wind, referred to as the positive and negative forcing (experiment 7). The resulting N34 indices due to the positive and negative wind forcing are plotted as solid and dashed black lines, respectively, in Fig. 10a and are nearly indistinguishable from the corresponding indices in Fig. 4. This is consistent with the approximately symmetric PMM wind response to NPO forcing, as seen in Fig. 2, and indicates that any slight asymmetry of the PMM wind is not responsible for the asymmetric ENSO responses. For consistency, experiments 7, 8, 9, and 10 will continue to use $+1$ ($-1$) times the positive PMM wind anomalies for “positive” (“negative”) forcing, respectively.

Next, the role of nonlinearity in the pseudostress calculation [Eq. (2)] is examined by using linearly opposite pseudostress to force the ICM (experiment 8). To estimate a “linear” component of the pseudostress, the pseudostress anomalies ($\mathbf{\tau}'$) are calculated separately for the forcing ($\mathbf{u}_f$) and the coupled ($\mathbf{u}_c$) wind anomalies via

$$
\mathbf{\tau}'_f = (\mathbf{u}_m + \mathbf{u}_f) | \mathbf{u}_m + \mathbf{u}_f| - | \mathbf{u}_m|, \quad (4)
$$

$$
\mathbf{\tau}'_c = (\mathbf{u}_m + \mathbf{u}_c) | \mathbf{u}_m + \mathbf{u}_c| - | \mathbf{u}_m|, \quad (5)
$$

$$
\mathbf{\tau}' = \mathbf{\tau}'_f + \mathbf{\tau}'_c, \quad (6)
$$

where $\mathbf{\tau}'$ is the total pseudostress anomaly applied in the model. First, the $+\text{PMM}$ wind forcing is used to calculate $\mathbf{\tau}'_f$ and the ICM is run by replacing Eq. (2) with Eq. (6); results are plotted as a solid blue line in Fig. 10a. Next, the ICM is rerun with the same $\mathbf{\tau}'_f$ calculated from the $+\text{PMM}$ wind [i.e., Eq. (4) is the same] but $\mathbf{\tau}'_c$ in Eq. (6) is multiplied by $-1$ to simulate a negative external forcing. Hence, the resulting $\mathbf{\tau}'_f$ in Eq. (6) is the exact
linear opposite of the “positive” case. The resulting N34 evolution (plotted as blue dashed line in Fig. 10a) remains very similar to the fully coupled case (experiment 1, black line in Fig. 4b). The similarity of the N34 evolution from experiment 7 (black lines Fig. 10a) and experiment 8 (blue lines Fig. 10a) indicates the nonlinear pseudostress calculation plays a minor role in the asymmetric ENSO response in this case.

Nonlinearity in the atmospheric heating is implicitly addressed in the uncoupled simulations (experiment 2; green lines in Fig. 4), which show that warm ENSO events produced by +PMM wind anomalies have larger amplitude than the cold ENSO events produced by the −PMM forcing. We repeated the uncoupled experiment with linearized wind (as in experiment 7) or linearized pseudostress (as in experiment 8) and found that the results did not differ substantially from the original uncoupled case (experiment 2). Results from the linearized uncoupled experiments are not shown because of similarity with the original uncoupled simulations. We also determined that the iterative quality of the heating calculation from the model had little influence on the results. As such, nonlinearity in the atmospheric heating calculation is not responsible for the asymmetric ENSO response to the PMM.

Finally, to show that the asymmetric ENSO response is largely a result of the relationship between thermocline depth and anomalous vertical temperature gradients, the subsurface temperature dependence on thermocline depth was made symmetric around the mean thermocline depth (experiments 9 and 10). The model parameterizes the subsurface temperature anomaly \( T_{\text{sub}} \) in terms of the upper-layer height anomaly \( h \), or thermocline depth anomaly, as follows (Battisti 1988):

\[
T_{\text{sub}} = \Theta(h) \tanh[\lambda(h_{\text{bar}} + 1.5|h|)] - \tanh[\lambda h_{\text{bar}}],
\]

where parameters \( \Theta = 28 \text{ K} \) and \( \lambda^{-1} = 80 \text{ m} \) for \( h > 0 \), and \( \Theta = -40 \text{ K} \) and \( \lambda^{-1} = 33 \text{ m} \) for \( h \leq 0 \), and \( h_{\text{bar}} \) is the mean thermocline depth along the equator as a function of longitude (solid line; Fig. 11). For the linearized thermocline experiments the parameters are changed to \( \Theta = 28 \text{ K} \) and \( \lambda^{-1} = 80 \text{ m} \) for \( h > 0 \), and \( \Theta = -28 \text{ K} \) and \( \lambda^{-1} = 80 \text{ m} \) for \( h \leq 0 \), which results in a symmetric subsurface temperature structure as a function of the thermocline depth perturbation (dashed line; Fig. 11). The original asymmetry between positive and negative thermocline anomalies is related to the vertical temperature structure of the ocean and equatorial upwelling, and is dependent on the mean thermocline depth \( h_{\text{eq}} \); the SST tendency in locations where the mean thermocline is near the surface (e.g., the eastern equatorial Pacific) is more sensitive to changes in thermocline depth (via mean upwelling) than regions where the mean thermocline is far from the surface.

Figure 10b shows the same simulations as in Fig. 10a, except using the symmetric subsurface temperature \( T_{\text{sub}} \) structure as described above. The fully coupled N34 index response to the linear PMM wind forcing (solid and dashed black lines respectively; experiment 9) indicates that the asymmetry in the \( T_{\text{sub}} \) parameterization is largely responsible for the asymmetric N34 response. However, with symmetric \( T_{\text{sub}} \) parameterization, the response to −PMM wind anomalies has larger amplitude than the response to +PMM wind anomalies. When the nonlinearity in the wind stress calculation is also removed (experiment 10; blue lines in Fig. 10b), the ENSO response to the PMM forcing is nearly symmetric. These results show that in the absence of nonlinearity arising from \( T_{\text{sub}} \) asymmetry (e.g., where the thermocline–SST feedbacks are weak) the wind stress calculations and other nonlinearities tend to produce stronger cold ENSO events. Implications of this finding will be discussed in section 5.

5. Conclusions

A set of ensemble simulations using the NCAR Community Earth System Model (CESM) and the Battisti (1988) version of the Zebiak and Cane (1987) intermediate coupled model (ICM) were run to examine the ENSO response to the Pacific meridional mode (PMM). These two models were chosen due to the complementary physics each model contains. The CESM (run with CAM5 physics coupled to a slab ocean model) contains the atmosphere–ocean coupling
required to simulate the PMM, but does not contain ocean dynamics to simulate ENSO. The ICM contains the necessary ocean dynamics to simulate ENSO, but is unable to simulate the PMM due to lack of wind-induced evaporation. The combination of the CESM and the ICM used in this study cleanly separates the interactions between the PMM and ENSO, thus allowing us to identify the dominant mechanisms through which the PMM interacts with ENSO.

We find the CESM generates a response representative of the PMM when forced with the surface heat flux associated with the North Pacific Oscillation (NPO), consistent with VAF2009. The positive (negative) NPO forcing initiates a persistent warm (cold) SST anomaly that propagates into the tropical Pacific. This SST anomaly subsequently generates westerly (easterly) wind anomalies in the western and central equatorial Pacific. We find that the symmetric response of the CESM to the NPO forcing is much larger than the asymmetric response, suggesting that the PMM is essentially linear.

While our CESM results support the known connection between the NPO and PMM variability, the results from the ICM experiments reflect the importance of extratropical atmospheric forcing in the development of ENSO events through the PMM. The fully coupled ICM produces a warm (cold) ENSO event that peaks in the boreal fall (SON) when forced with the +PMM (−PMM) ensemble-mean wind anomalies. Although this model tends to simulate ENSO events that peak a couple of months early, it is a useful tool for diagnosing the connection between the PMM and ENSO. Previous studies also show the connection between PMM variability and ENSO variability (Vimont et al. 2003b; Chang et al. 2007; Zhang et al. 2009; Larson and Kirtman 2013; Lin et al. 2015; Vimont et al. 2014); here, the specific mechanisms through which PMM variability excites ENSO development are explored. The sensitivity experiments performed using the ICM in this study show that oceanic Kelvin waves are the dominant mechanism responsible for ENSO development due to PMM atmospheric forcing. This is especially true for the response to the positive forcing. Although Kelvin waves dominate the initial ENSO response to both positive and negative PMM atmospheric forcing, we find that externally excited oceanic Rossby waves are only significant for the ENSO response in the tropical Pacific in the −PMM forced case. Furthermore, we show the importance of the seasonality of the PMM on ENSO generation through interactions with Kelvin and Rossby wave mechanisms. The seasonality of the CESM simulated PMM, which peaks in boreal spring, supports ENSO growth through Kelvin wave propagation. These results support the importance of Kelvin wave propagation in ENSO development found in previous studies (Vimont et al. 2003a,b; Alexander et al. 2010).

The large spread between individual ensemble members indicates the importance of natural variability in the PMM–ENSO relationship. Natural variability, captured by the unforced, internal variability of the CESM, is especially important to the structure of simulated PMM wind anomalies. Variations in the initial conditions and random noise result in a large range of ENSO responses between ensemble members. Because of this large spread, we conclude that natural variability has important implications for the predictability of ENSO through the PMM. Although the ensemble shows a clear bias toward warm (cold) ENSO events when forced with positive (negative) PMM winds, not all PMM simulations generate an El Niño event, supporting the results found in Chang et al. (2007). Other studies show similar a relationship between the PMM and ENSO. Larson and Kirtman (2015) also find a warm ENSO bias in response to the positive PMM. The large spread between ensemble members may indicate low skill for ENSO predictability using solely the PMM. Similarly, Larson and Kirtman (2014) show little skill in ENSO forecasts using the PMM, motivating further research to better understand the role of natural variability and stochastic forcing in ENSO development.

We also found that the ENSO response simulated by the ICM to positive and negative PMM wind anomalies is asymmetric, despite the ICM being in a linearly stable regime. Several sensitivity experiments utilizing the ICM demonstrated that ocean dynamics within the ICM dominate the ENSO asymmetry. Most notably, the nonlinear parameterization of subthermocline temperature in the ICM dominates the asymmetry in the ENSO response to PMM forcing. After “symmetrizing” (i.e., making symmetric around the mean thermocline depth) the subthermocline temperature parameterization as a function of thermocline depth, we find that the ICM produces stronger negative ENSO events due to wind stress calculations and other nonlinearities.

The nonlinearities of the ICM are understood and known to be strongly dependent on the thermocline parameters. However, the results obtained after these thermocline nonlinearities are removed may shed some light on observed ENSO behaviors. The N34 results from simulations containing symmetric thermocline–SST feedbacks indicate the importance of the wind stress effects that were previously masked by the thermocline dynamics. In these simulations, the ICM generates larger-magnitude cold ENSO events, indicating the importance of the wind stress on ENSO variability in the absence of strong thermocline variations. Although the ICM is incapable of simulating ENSO diversity,
these results may have implications for ENSO characteristics that may be masked by the large nonlinear thermocline behaviors. For example, events that rely less on thermocline–SST feedbacks, such as central Pacific (CP) ENSO events (Kao and Yu 2009), may be more influenced by atmospheric forcing. We do, in fact, find stronger cold ENSO events in the absence of thermocline–temperature asymmetry largely due to wind stress, which is consistent with the negative skewness of CP ENSO events (Kao and Yu 2009). We reiterate, though, that a more complete analysis of how the PMM influences ENSO diversity should be conducted with a model that better simulates such ENSO diversity.

We acknowledge the use of an unrealistic intermediate coupled model with known biases in this study. Despite these limitations, results herein do contribute to our understanding of PMM–ENSO interactions. In particular, the results highlight the importance of directly forced Kelvin waves in the PMM’s influence on ENSO initiation, and the potential importance of nonlinearity in wind stress and thermocline–SST interactions in producing nonlinear ENSO responses. Furthermore, the results from individual model simulations show that natural variability and stochastic forcing may play a very strong role in ENSO initiation and development. Future work to identify stochastic forcing structures that are either favorable or detrimental for ENSO development with respect to the NPO/PMM initiation mechanism, including development of ENSO diversity, which is not fully addressed herein, would be beneficial.

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