Collaborative Project: Pacific Decadal Variability and Central Pacific Warming El Niño in a Changing Climate

Lead PI and PD: E. Di Lorenzo (GaTech)
Co-PIs: K. Cobb (GaTech), N. Schneider (U. Hawaii), B. Anderson (Boston U.), D. Vimont (U. Wisconsin-Madison), M. Alexander (NOAA/ESRL)

University of Wisconsin – Madison Contribution:

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PI: Dr. Daniel J. Vimont
Contact Information: University of Wisconsin – Madison
Atmospheric and Oceanic Sciences Department
1225 W. Dayton St.
Madison, WI 53705
Phone Number: 608-263-3420
Fax Number: 608-262-0166
Email: dvimont@wisc.edu

Applicant/Institution: University of Wisconsin – Madison
Research and Sponsored Programs
21 N. Park Street, Suite 6401
Madison, WI 53715-1218
Phone: 608-262-3822; FAX 608-262-5111

Work Accomplished:

This project funded two efforts at understanding the interactions between Central Pacific ENSO events, the mid-latitude atmosphere, and decadal variability in the Pacific. The first was an investigation of conditions that lead to Central Pacific (CP) and East Pacific (EP) ENSO events through the use of linear inverse modeling with defined norms. The second effort was a modeling study that combined output from the National Center for Atmospheric Research (NCAR) Community Atmospheric Model (CAM4) with the Battisti (1988) intermediate coupled model. The intent of the second activity was to investigate the relationship between the atmospheric North Pacific Oscillation (NPO), the Pacific Meridional Mode (PMM), and ENSO. These two activities are described herein.

In addition to the findings below, this project contributed to the training of two graduate students (K. Watkins, M.S. 2012, and E. Thomas, current student). The first activity resulted in the following publication:

Vimont, D. J., Alexander, M. A., & Newman, M. (2014). Optimal growth of Central and East Pacific ENSO events, (May), 1–8. doi:10.1002/2014GL059997.1.

The second activity (above) is being incorporated into another publication that is in progress.

Investigation of CP and EP ENSO variability using Linear Inverse Modeling

The first project aimed to understand large-scale precursors to CP and EP ENSO events, which could lend insight into critical processes responsible for interactions between CP ENSO events and decadal variability over the Pacific. Results are presented in a recently accepted manuscript, Vimont et al. (2014) and summarized herein.

Linear Inverse Modeling (LIM) is used to investigate optimal initial conditions that grow into

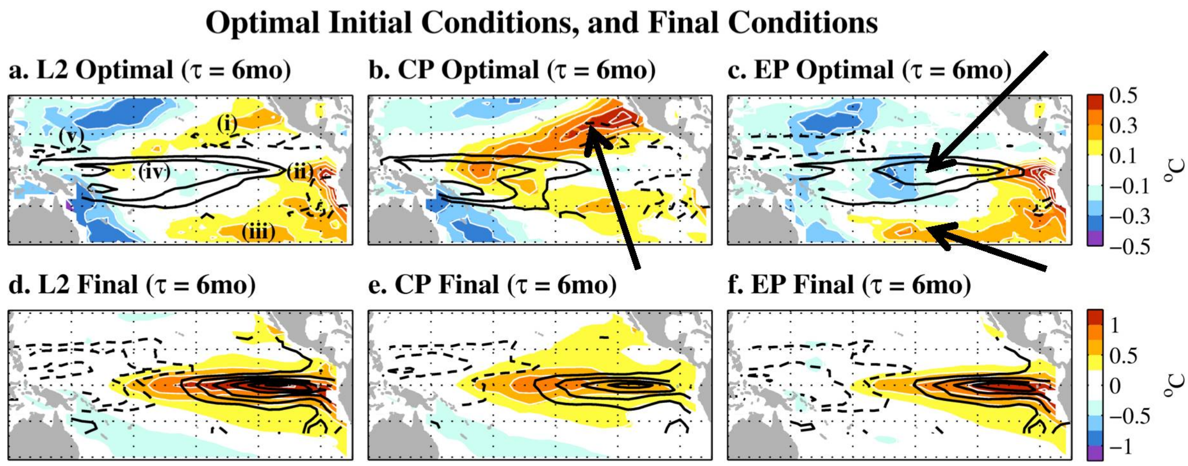

Figure 1: τ= 6mo optimal initial (a, b, c) and associated final (d, e, f) structures calculated under the L2 (a, d), CP (b, e) and EP (c, f) norms. SST is shaded [contour 0.1°C in (a, b, c) and contour 0.25°C in (d, e, f)] and Z20 variations are contoured in black [contour 1.6m in (a, b, c) and contour 4m in (d, e, f); solid contours denote positive values, dashed contours denote negative values, and the zero contour has been omitted]. Note that the contour interval for the final condition is 2.5 times the contour interval for the optimals.

CP or EP ENSO events. A full description of the analysis method and results are presented in Vimont et al. (2014). We use LIM with defined CP and EP norms to specifically identify optimal initial conditions that grow into CP or EP ENSO events. Shown in Fig. 1 are the optimal 6mo initial conditions (top row) and final conditions (bottom row) from the SST / thermocline LIM used in Vimont et al. (2014). Of note is the finding that CP initial conditions (panel b) involve large subtropical SST anomalies related to the NPO/PMM (highlighted by the large black arrow), while EP initial conditions involve westerly equatorial zonal winds, deepened equatorial thermocline across the Pacific, and a zonally elongated band of SST anomalies in the southern Pacific, along about 20°S (also highlighted by the large black arrows in panel c). These results confirm the hypotheses in the original proposal that CP events are triggered via the Seasonal Footprinting Mechanism [Vimont et al. (2001; 2003a, b)].

Further investigation of the role of the NPO in generating CP events can be inferred through looking at the atmospheric structures that generate the optimal pattern in Fig. 1b. Recall that linear inverse modeling assumes the following mathematical formulation for the "state" $\boldsymbol{x}$ of the system:

$$\frac{d\mathbf{x}}{dt} = \mathsf{L}\mathbf{x} + \xi. \tag{1}$$

The state of the system, in this case, is the SST and thermocline variability over the tropical Pacific (the domain in Fig. 1), as represented by the first 9 EOFs of SST and the first 3 EOFs of thermocline variability. The model in (1) indicates that the state of the system $\boldsymbol{x}$ evolves via deterministic dynamics (**L**) as well as stochastic forcing ($\boldsymbol{\xi}$). An estimate of the actual stochastic forcing can be obtained as a residual between the time derivative of the state of the system [on the left hand side of (1)] and the deterministic evolution of the state [the first term on the right hand side of (1), **Lx**]. We calculate that stochastic forcing, and project the result onto the optimal structure shown in Fig. 1b. The result is a time series of stochastic forcing that would tend to generate optimal initial conditions that lead to a CP event.

We evaluate the spatial structure of atmospheric variability associated with the CP optimal initial condition in Fig. 2. Fig. 2(a) shows the CP optimal structure, and the regression of SLP

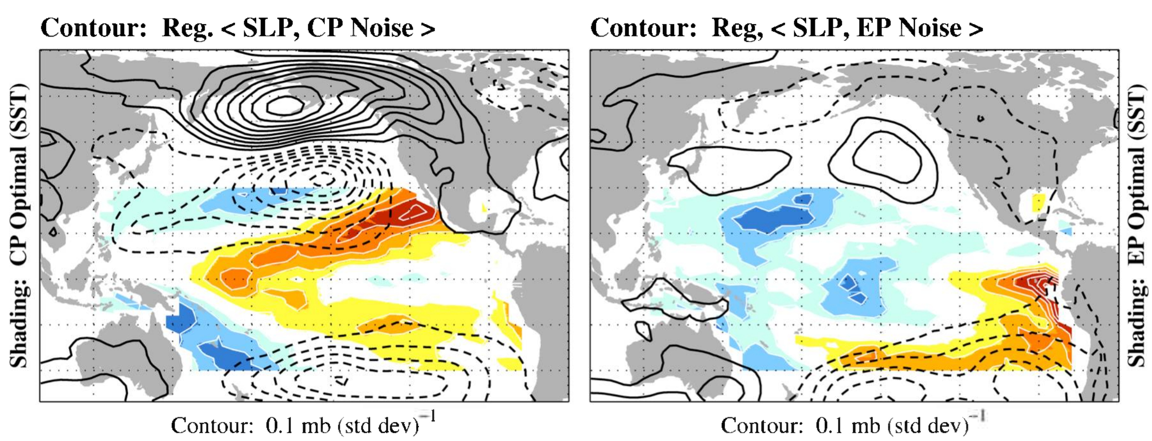

Figure 2: Atmospheric structures that generate the (left) CP ENSO optimal and (right) EP ENSO optimal. SST is shaded (units are arbitrary) and contours denote SLP anomalies (contour 0.1 mb (std dev)$^{-1}$). The left figure indicates that CP optimals are generated by the atmospheric North Pacific Oscillation, which confirms a link between the NPO and CP ENSO variability.

onto the noise forcing time series associated with that optimal structure. The SLP regression map bears a strong resemblance to the atmospheric North Pacific Oscillation [Rogers (1981), Linkin and Nigam, 2008]. This confirms a hypothesized link between the atmospheric North Pacific Oscillation and Central Pacific ENSO events. No similar structure is seen as a precursor to EP events, however additional analysis (not shown) indicates that zonal wind variations in the tropical Pacific may be responsible for generating EP events. This is a fundamentally different structure and phenomenon that is shown in Fig. 2a).

Results from the Linear Inverse Modeling are written up in Vimont et al. (2014). Note that the analysis of stochastic forcing (Fig. 2) is not included in that publication.

Connections between the Mid-latitudes and Tropics: Dynamical Modeling

Results from previous studies, and from the LIM (above) suggest that the mid-latitude atmosphere can influence tropical ENSO variability through the Seasonal Footprinting Mechanism [Vimont et al. (2001; 2003a, b)]. But, how does the signal of mid-latitude variability propagate to the equatorial Pacific? Vimont (2010) shows that thermodynamically coupled structures tend to propagate equatorward through the Wind-Evaporation-SST (WES) feedback. This mechanism can explain how subtropical SST anomalies can generate equatorial wind stress anomalies, but cannot explain the subsequent evolution of ENSO because ENSO relies on coupled interactions between the atmosphere, surface ocean, and thermocline (the so-called Bjerknes feedback).

We investigated the dynamical pathways by which the atmospheric NPO can influence ENSO using two different models. The two models we use are the National Center for Atmospheric Research Community Atmospheric Model coupled to a slab ocean model

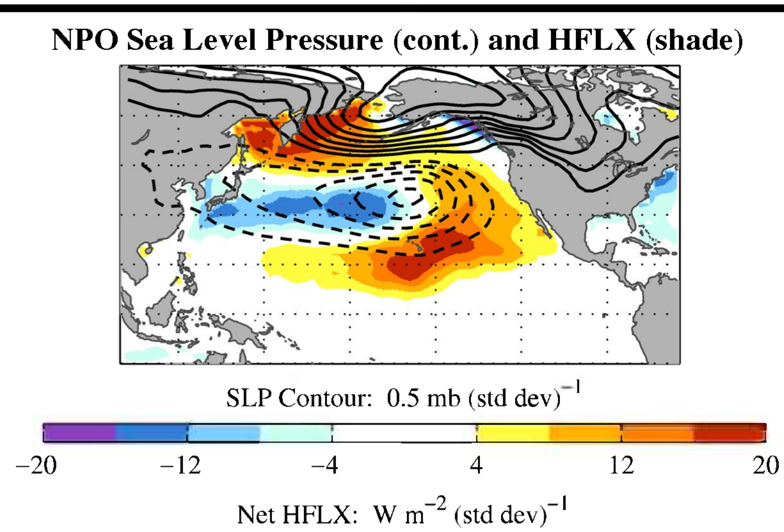

Figure 3: Spatial structure of the atmospheric NPO (contours), and associated surface heat flux (shading). The surface heat flux shown is used as a forcing to the CAM+SOM model from November through March. In April, the heat flux is shut off and the coupled CAM+SOM is allowed to evolve on its own. The response in CAM+SOM is shown in Fig. 4.

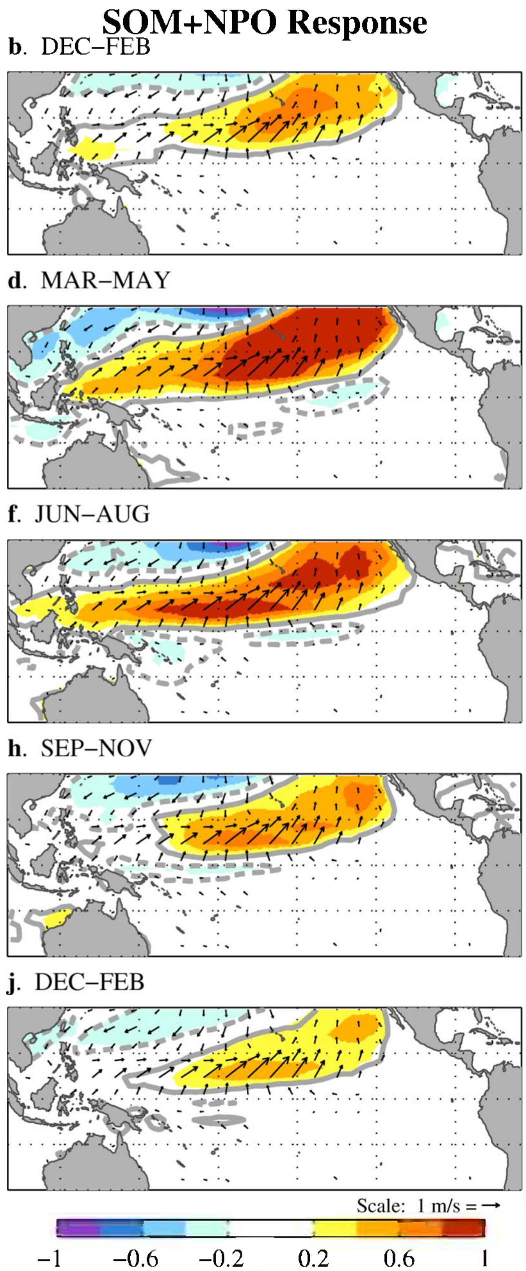

Figure 4: Response of CAM+SOM to NPO surface heat flux forcing associate with the NPO (Fig. 3). Forcing is applied from November through March, then the model is allowed to evolve on its own (panels f – j). Shown is SST (shading) and surface winds (scale at bottom). Contours denote a response that is significant at the 95% level using a t-test, and vectors are only shown where statistically significant using a bivariate t-test.

(CAM+SOM), and the Battisti (1988) version of the Zebiak and Cane (1987) intermediate coupled model (B88 ICM)[1]. The CAM+SOM contains physics necessary for meridional mode variations but not ENSO, while the B88 ICM contains physics necessary for ENSO but not meridional mode variations. A combination of these two models allows direct investigation of how the PMM and ENSO interact. The first set of experiments was documented in Vimont et al. (2010), and involves running an ensemble of simulations using CAM+SOM in which the coupled model is forced by surface heat flux variations (Fig. 3) associated with the NPO throughout boreal winter, then allowed to evolve on its own for another 12mo. This simulates the thermodynamically coupled response of the atmosphere / ocean system to the NPO/PMM, but cannot simulate ENSO (the slab ocean model does not contain the necessary physics for generating ENSO variability). The response of the CAM+SOM is shown in Fig. 4, and is consistent with results from Vimont et al. (2010). In particular, the model produces surface wind anomalies in the equatorial region shortly after the forcing is turned on (panels b-d) (note that the response in Fig. 4 does *not* include any of the original wind anomalies associated with the NPO). Even after the forcing is shut off, the model continues to produce surface wind anomalies through the following boreal winter (panel j). This demonstrates that the atmospheric NPO (contours in Fig. 3) can generate PMM-related SST anomalies (shading in Fig. 4) that generate surface wind anomalies in the tropics.

To investigate the response of ENSO to NPO / PMM – generated surface wind

[1] Model parameters in the B88 ICM are adjusted to more realistic values that result in a linearly stable ENSO mode, as in Thompson and Battisti (2001).

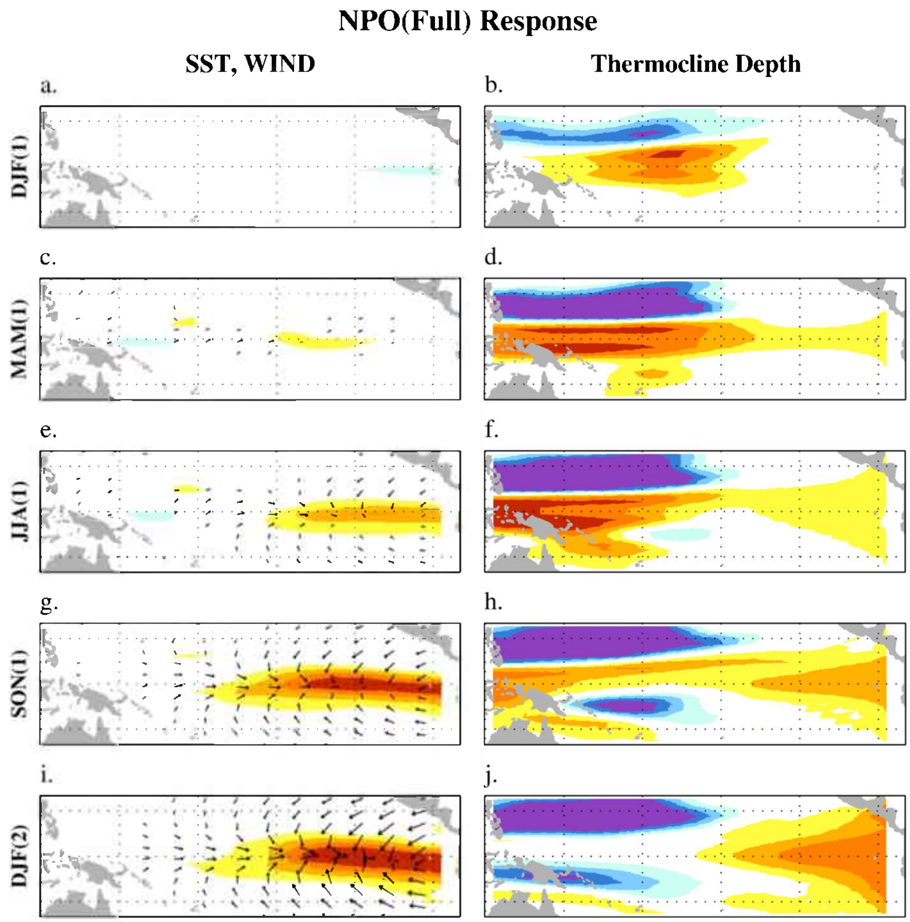

Figure 5: SST and surface wind (left) and thermocline depth (right) response of the B88 ICM to surface wind forcing from CAM+SOM (not shown; see text for details). The figure shows the response of ENSO to wind variations associated with the PMM's response to the NPO. The NPO forcing is applied to CAM+SOM from November of year 0 (not shown) through March of year 1 (the beginning of panel 3). The resulting wind stress is applied to the B88 ICM from November of year 0 (not shown) through March of year 2. The B88 ICM response includes the development of an ENSO event during the DJF *after* the NPO forcing is applied (one year later).

anomalies, the surface wind variations from the CAM+SOM experiment (Fig. 4) were applied as an external forcing to the B88 ICM. Results from the B88 ICM are shown in Fig. 5. Note that each panel of Fig. 5 shows the response of the B88 ICM to surface wind anomalies from a corresponding panel of Fig. 4. The B88 ICM generates thermocline anomalies very soon after the surface wind anomalies are imposed (the response is evident as early as DJF after the forcing is imposed; Fig 5b). The B88 ICM starts warming at the surface relatively soon after the forcing is imposed, but does not warm substantially until boreal Fall (Fig. 5g), well after the NPO forcing has been shut off in CAM+SOM (the wind anomalies from CAM+SOM are still forcing the B88 ICM, though). By late summer and early fall, thermocline anomalies in the far eastern equatorial Pacific have developed with some amplitude.

A more detailed analysis of the evolution of the tropical Pacific is shown via the evolution of the Niño 3.4 index (N34) in Fig. 6. The N34 index for the SST evolution shown in Fig. 5 is shown as a thick black line in Fig. 6. The B88 ICM indicates that the PMM generates an ENSO event during the boreal fall following the NPO forcing, with amplitude of about 0.5°C. How does the PMM generate this ENSO event? We answer this question by altering the physics of the B88 ICM response. First, we investigate how the model would respond if there were no coupling (i.e. the Bjerknes feedback were shut off). This is shown in the thin, dashed black line with circles in Fig. 6. There is some warming by boreal fall, but it is only half the amplitude of the full coupled response, and dies off by early spring after the ENSO event. Hence, not surprisingly, coupling is important for the dynamical response to the NPO/PMM – related surface wind anomalies.

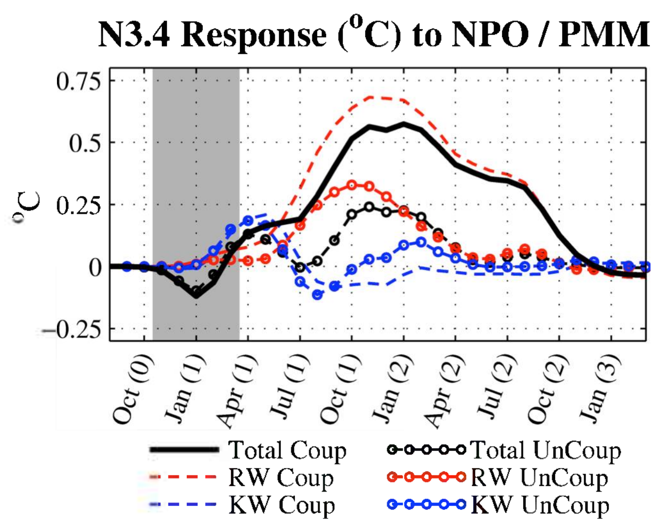

Figure 6: Niño 3.4 index response of the Battisti (1988) ICM to atmospheric forcing from the PMM, as simulated by the CAM+SOM (see text). Shown are the full coupled model response (thick black line), the coupled ICM response to Kelvin wave forcing only (dashed blue line), the coupled ICM response to Rossby wave forcing only (red dashed line), the uncoupled ocean response to Kelvin wave forcing only (dashed blue line with circles), and the uncoupled ocean response to Rossby wave forcing only (dashed red line with circles).

How does the equatorial ocean respond to the surface wind forcing? We answer this question by allowing the imposed surface wind anomalies to generate *either* equatorial oceanic Kelvin waves only (blue curves in Fig. 6) or equatorial oceanic Rossby waves only (red curves in Fig. 6). When the model responds with Kelvin waves only (and the model is still coupled; thin dashed blue curve in Fig. 6), there is very little response to the imposed forcing! In contrast, when the model responds with Rossby waves only (and the model is still coupled; thin dashed red curve in Fig. 6), the model reproduces the full warming (and produces a bit more warming). Analysis of the uncoupled response to Kelvin-only or Rossby-only forcing (blue and red curves with circles, respectively) shows the reason why Rossby waves are so important. The Kelvin-only uncoupled response is quick, but is weak and dies off by boreal summer following the NPO. In contrast, the Rossby-only uncoupled response starts to amplify in boreal summer following the NPO, and persists through the following boreal winter. This allows plenty of time for coupled feedbacks to amplify the Rossby-only response. The larger Rossby response may also be due to timing: the equatorial Pacific is more sensitive to perturbations during boreal Fall than boreal Spring. More research is needed to determine whether the seasonality of the response is important. Results from this analysis are being prepared for publication.

Concluding Remarks

This project successfully demonstrated a pathway by which the atmospheric NPO can influence tropical Pacific ENSO variability. First, it was shown that the atmospheric NPO is most effective at generating Central Pacific ENSO events, as hypothesized in the original proposal. This was accomplished via a novel method of designing and utilizing a specific norm for use with linear inverse modeling. The results highlighted the role of the NPO specifically in generating CP ENSO events, using purely observational data.

How does the connection between the NPO and CP ENSO events evolve? The second task in this research explicitly simulated this connection by taking surface forcing from the NPO, applying it to the CAM+SOM model that includes physics necessary for the PMM, but cannot simulate ENSO. The resulting simulation provided surface wind anomalies associated with the PMM (the NPO -> PMM connection). Next, the surface wind anomalies were used to force the B88 Intermediate Coupled Model (ICM) that includes ENSO physics, but does not contain thermodynamic feedbacks necessary for the PMM (hence the B88 ICM forced simulations contain the PMM -> ENSO connection). As hypothesized, the tropical Pacific responds by producing an ENSO event. Interestingly, the ENSO event evolves via excitation of equatorial oceanic Rossby waves, not by generation of equatorial oceanic Kelvin waves. This is important because it provides clues into time scales of interaction between the NPO, the PMM, and ENSO.

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