On the joint role of subtropical atmospheric variability and equatorial subsurface heat content anomalies in initiating the onset of ENSO events

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Abstract. Previous research has shown that seasonal mean variations in both the subtropical/extra-tropical sea-level pressures over the central North Pacific and the subsurface heat content anomalies in the western equatorial Pacific are significantly related to the state of the El Niño/Southern Oscillation (ENSO) 12-18 months later. Here we find that positive (negative) subsurface temperature anomalies in the western equatorial Pacific during boreal-summer/fall, followed by negative (positive) anomalies in the sea-level pressure fields over the subtropical central North Pacific during boreal-winter, tend to result in positive (negative) mature ENSO events 12-15 months later (i.e. during the following boreal winter). When the intervening sea-level pressure anomalies are of the same sign as the preceding heat-content anomalies, the correlation between the heat-content anomalies and the following boreal-winter ENSO state disappears. There is still some relation between the boreal-winter sea-level pressure anomalies and the ENSO state the following year when the two precursor patterns are of the same sign, however the correlation is smaller and the ENSO events tend to be weaker. Additional analysis indicates that the two precursor fields are related to one another, however the sea-level pressure variations contain more unique information about, and provide better predictability of, the state of the following ENSO system than do the heat content anomalies.
1. Introduction

The El Niño/Southern Oscillation (ENSO) phenomenon is a coupled mode of variability of the ocean/atmosphere system in the equatorial Pacific characterized by a relative warming/cooling of the sea-surface temperatures (SSTs) over the eastern equatorial Pacific and a change in the zonal sea-level pressure gradient across the basin (Philander, 1985). While the time-scale for this oscillation is typically considered to be 3-6 years, it has been found that there is a boreal spring “predictability barrier” preventing the use of equatorial Pacific SSTs themselves to predict mature ENSO events (which typically occur during boreal winter - Larkin and Harrison, 2002) at time-scales longer than approximately 9 months (Torrence and Webster, 1998). However, theoretical and observational results suggest that preceding temperature anomalies in the subsurface equatorial Pacific can be present up to a year before mature ENSO events (Jin, 1997; Li, 1997; Meinen and McPhaden, 2000; Meinen and McPhaden, 2001; McPhaden, 2003) and that these anomalies can initiate the onset of overlying SST anomalies following the boreal spring “predictability barrier” (McPhaden, 2003). In addition, precursor fields in the subtropical/extra-tropical atmospheric fields also appear to overcome this “predictability barrier” and are related to the initiation of ENSO events that develop over the course of the following 9-12 months (Kidson, 1975; Trenberth, 1976; Reiter, 1978; Rasmusson and Carpenter, 1981; van Loon and Shea, 1985; Barnett, 1985; Trenberth and Shea, 1987; van Loon and Shea, 1987; Lysne et al., 1997; Gu and Philander, 1997; Li, 1997; Barnett et al., 1999; Pierce et al., 2000; Wang, 2001; Vimont et al., 2001; Vimont et al., 2003; Anderson, 2003; Anderson, 2004).

Here we will examine the joint relation of the seasonal-mean tropical/extra-tropical atmosphere structure and tropical Pacific heat content anomalies to the onset of boreal-winter
ENSO events. In Section 2, the various datasets used in this study are described. The relation between tropical Pacific sea-surface temperature anomalies and antecedent atmospheric and oceanic variability is examined in Section 3. Findings are summarized and briefly discussed in Section 4.

2. Data

The principal atmospheric dataset used in this investigation is the reanalysis product from the National Centers for Environmental Prediction (NCEP – see acknowledgements). Details about this dataset, including its physics, dynamics, and numerical and computational methods, are discussed in Kalnay et al. (1996) and Kistler et al., (2001). For this paper, we focus on the monthly-mean sea-level pressures because they contain significant precursor information regarding the development of large-scale SST anomalies in the equatorial Pacific (Kidson, 1975; Trenberth, 1976; van Loon, 1984; van Loon and Shea, 1985; Barnett, 1985; van Loon and Shea, 1987; Trenberth and Shea, 1987; B. Wang et al., 1999; Chan and Xu, 2000; Wang, 2001; Larkin and Harrison, 2002; Vimont et al., 2003; Anderson, 2003). These fields are represented at 2.5-degree resolution in both the meridional and zonal direction, encompassing a total of 144x73 grid points. We choose to use the reanalysis product owing to its fairly long continuous record (1948-2003 here) and its systematic treatment of observational and numerical data over this entire period.

In addition to the reanalyzed atmospheric fields, this study will also examine the time-evolution of the seasonal-mean subsurface ocean state. For this field we archive the observational analysis from the Joint Environmental Data Analysis Center (JEDAC), which uses an optimal interpolation procedure to produce 5x2.5 degree gridded values of temperature, mixed-layer depth and heat content (vertically-integrated from the surface to 400m – White,
1995; White, Personal Communication) at monthly time-scales from 1955-2003. Here we focus on the heat content anomalies (also referred to as heat storage anomalies) because these have also been shown to be significantly related to the onset and development of large-scale SST anomalies in the equatorial Pacific (e.g. McPhaden, 2003). While estimates of tropical oceanic heat content anomalies prior to about 1980 are generally reduced compared with the period following the 1980’s (Chepurin and Carton, 1999), we find that results using a sub-set of data during this latter time period are qualitatively (and quantitatively) the same as those derived from the full time-series. Here we use the full time-series in order to increase the robustness of the results.

To capture variability in the equatorial Pacific sea-surface temperature field, we archive the NINO3.4 index provided by NOAA’s Climate Diagnostics Center (CDC - see acknowledgments for data availability). This index is defined as the area-average SST anomalies between 5N-5S and 170-120W. The version used here is derived from the Reanalyzed SST fields described in Hurrell and Trenberth (1999).

Throughout this paper, results will be based upon statistical relationships between anomalous values of various fields. For the heat-content fields (Fig.2) we tested the significance of these results explicitly by performing a bootstrap analysis following the methodology of Ebisuzaki (1997). This method retains the autocorrelation structure of the grid-point and index time-series, as well as their possible non-Gaussian distributions. In addition, it retains any spatially-correlated features the grid-point fields might have. We find that except in very localized regions, correlation values of |r|=0.35 are above the 95% confidence interval (not shown), which we set as the confidence interval for this figure.

3. Results
Previous research has shown that there exists a precursor mode of boreal-winter sea-level pressure (SLP) variability in the central tropical/extra-tropical North Pacific that precedes variations in the January-March El Niño/Southern Oscillation (ENSO) by approximately 12-15 months (Barnett, 1985; Trenberth and Shea, 1987; Chan and Xu, 2000; Vimont et al., 2003; Anderson, 2003). This mode of SLP variability can be captured by a sea-level pressure index (SLPI) centered on 140-175W and 10-25N (Anderson, 2003). The SLPI is constructed by calculating the monthly grid-point sea-level pressure anomalies with respect to the climatological value for the given month and then normalizing these grid-point anomalies by their interannual standard deviation. The normalized monthly anomalies are then area-averaged over 140-175W and 10-25N to arrive at monthly values for the SLPI. While the index itself is defined for the tropical central North Pacific region, it is significantly correlated with additional SLP anomalies over the subtropical and extra-tropical North Pacific (Anderson, 2003; Anderson, 2004) and as such we refer to it as a “subtropical” index in order to differentiate it from the tropically-centered SLP anomalies typically associated with the ENSO system.

The seasonal-mean value of the SLPI from November-March is shown in Figure 1 and indicates a significant negative correlation (r=-0.61) with the boreal-winter (January-March) ENSO state 12-15 months later (note that the concurrent correlation between the boreal-winter SLPI and NINO3.4 index is r=0.05). Previous observational and modeling studies suggest seasonal mean variations in the boreal-winter sea-level pressure fields in the SLPI region can influence the initiation and development of the ENSO system by modifying the wind-stress fields over the central tropical Pacific, which in turn are related to concurrent changes in the underlying subsurface temperature structure of the equatorial/tropical Pacific (Anderson, 2004; Anderson and Maloney, 2005). These same subsurface temperature anomalies are conducive to
initiating overlying surface temperature anomalies the following spring, which then develop into mature ENSO events by the following boreal winter (Jin, 1997; Meinen and McPhaden, 2000).

Figure 2 shows the evolution of this subsurface temperature structure, as represented by the 3-month mean heat storage anomalies from 0-400m, regressed against the seasonal-mean Nov.-Mar. SLPI, starting 9 months before the Nov.-Mar. SLPI period and progressing through the period concurrent with the SLPI. Here we multiply the heat storage regression values by -1 in order to present the related heat-storage evolution preceding positive ENSO events (i.e. El Niños). During the preceding boreal summer there are significant equatorial anomalies over the western Pacific indicative of anomalously warm subsurface temperatures in this region (there are no corresponding anomalies in the overlying equatorial SSTs – not shown). During the boreal fall and early boreal winter, these heat content anomalies migrate eastward towards the central Pacific and intensify during the late boreal winter concurrent with the SLPI. However, significant warm-water SST anomalies do not appear in the central and eastern equatorial Pacific until the following May-July period (Anderson, 2003)

These results indicate that there exist precursor subsurface temperature signatures prior to periods of anomalous sea-level pressures in the SLPI region. In turn, these results suggest that the ability of the SLPI anomalies during a given winter to initiate ENSO events may be dependent upon the existence of such subsurface temperature signatures, similar to what has been found for westerly-wind bursts (Moore and Kleeman, 1999; Perigaud and Cassou, 2000; Federov, 2002).

To test this hypothesis, we calculated a western-Pacific heat storage index (HST index) by area-averaging the heat-storage anomalies over the region 160-180E and 5N-5S (the core of the positive correlations seen in Figure 2d, e). We then calculated the 5-month mean for all 5-month
periods preceding the Nov.-Mar. SLPI and selected the period with the highest overall correlation with the SLPI. The selected period, from June-October, has a correlation with the seasonal mean Nov.-Mar. SLPI of $r=-0.49$. Not unexpectedly, the Jun.-Oct. HST index is also significantly correlated with the Jan.-Mar. NINO3.4 index 18 months later ($r=0.57$).

However, when the Jun.-Oct. HST index and the following Nov.-Mar. SLPI are conditioned upon whether they have the same or opposite sign (Figure 3), it is found that the boreal-summer/fall western Pacific HST index is much more strongly correlated with the Jan.-Mar. NINO3.4 index 18 months later if the HST index has the opposite sign as the intervening SLPI compared with years in which it has the same sign ($r=0.69$ and $r=0.06$ respectively). The correlation of the boreal-winter SLPI with the Jan.-Mar. NINO3.4 index one year later is also higher if the SLPI is preceded by anomalous western equatorial heat content anomalies of the opposite sign ($r=-0.72$) compared with years in which the heat content anomalies are of the same sign ($r=-0.37$). In addition to having a weaker correlation with the preceding SLPI (which is only significant at the 90% level for the 17 events shown here), the strength of the NINO3.4 index (represented by its interannual variance) is also weaker following years in which the SLPI and western Pacific HST index have the same sign as compared with years in which they have the opposite sign ($\sigma_{NINO3.4}^2 = 0.40$ compared with $\sigma_{NINO3.4}^2 = 1.45$; for the full time-series $\sigma_{NINO3.4}^2 = 1.0$ by definition).

To test whether the difference in the opposite-sign and same-sign composite SLPI/NINO3.4 and HST/NINO3.4 correlations described above are simply an artifact of the compositing procedure itself, we created and selected 1000 randomly-generated 47-year time-series for each index such that they had approximately the same overall correlations as observed [i.e., $r(\text{HST/NINO3.4})=0.57\pm/0.025$, $r(\text{SLPI/NINO3.4})=-0.61\pm/0.025$, and $r(\text{HST/SLPI})=-0.49\pm/0.025$].
0.025]. For each of the 1000 sets of time-series, we performed the same compositing procedure based upon the sign of the artificial SLPI and HST time-series and then correlated the composit ed time-series with the composit ed artificial NINO3.4 time-series. We find that the mean correlation values for the opposite-sign and same-sign composites are nearly identical (the difference between the two is \( \Delta r = 0.0019 \) and 0.0064 for the mean SLPI/NINO3.4 and HST/NINO3.4 correlations respectively). In addition, the differences between the *observed* opposite-sign/same-sign composite correlations are greater than 95% of the differences derived from the artificial time-series. Finally, the opposite-sign composite correlations for the observed SLPI/NINO3.4 and HST/NINO3.4 time-series (\( r = -0.72 \) and \( r = 0.69 \) respectively) are greater than 95% of those derived from the artificial time-series while the same-sign composite correlation for the observed HST/NINO3.4 time-series (\( r = 0.06 \)) is less than 95% of those derived from the artificial time-series. The same holds for the means, difference of means, and distributions of the composit ed artificial NINO3.4 variances. As such we argue that the compositing procedure itself is not likely (at the 95% confidence level) to have spuriously generated the observed opposite-sign/same-sign correlations or variances discussed above.

Overall it appears that the relation of either the HST index or the SLPI to the development of mature ENSO events is a function of the status of the other index; in addition the two indices are significantly correlated with one another, making it difficult to quantify the relation of the ENSO system to each index separately. One way to do so is to perform a “Granger Causality” test (Granger, 1969). This test quantifies the difference in predictability provided by the two indices by comparing the residual sum of squares between the actual and predicted values of the Jan.-Mar. NINO3.4 using two separate estimates. In the first estimate, both the Jun.-Oct. HST index and Nov.-Mar. SLPI, along with the previous year’s Jan.-Mar. NINO3.4 value, are used as
optimally-weighted linear predictors to estimate the following year’s Jan.-Mar. NINO3.4 value (termed the “unrestricted” prediction). In the second, one of the predictors is removed (the SLPI for instance) and a second estimate is made by optimally weighting the remaining predictors; this estimate is termed the “restricted” prediction. The skill of prediction (in a least squares sense) from the restricted model will always be less than that from the unrestricted model. The difference in the residual sum of squares from the two models quantifies the amount of non-redundant variance that is contributed by the predictor that is removed from the restricted model. Performing this analysis indicates that 19% of the variance of the boreal-winter NINO3.4 is contributed by the SLPI alone. In contrast, only 8% of the variance is uniquely contributed by the western Pacific HST index. Both of these values are significant at the 95% confidence interval with the SLPI-related causality significant at the 99% confidence interval. (Interestingly, the NINO3.4 index from the previous winter only explains an additional 0.1% of the variance in the following value of the NINO3.4.)

Similar results are obtained when the predicted time-series are generated from out-of-sample forecasts (in which a yearly value from each time-series is removed, the linear-statistical model is trained using the remaining values from the time-series, and then a forecast is made for the out-of-sample year). In addition, the out-of-sample cross-validation procedure indicates that when both the HST index and SLPI predictors are used, the out-of-sample forecast of the NINO3.4 index correlates with observed values at $r=0.69$. When we use only the SLPI as the sole predictor, the correlation drops to $r=0.64$. However, when we use only the HST index as the sole predictor, the correlation drops further to $r=0.55$. Overall, these results suggest that the western Pacific HST index does not provide much additional predictability regarding the state of the ENSO system beyond that associated with the SLPI; in addition, they suggest that
atmospheric variations in the SLPI region may have a greater influence upon the state of the
ENSO system compared with variations in the subsurface heat content anomalies over the
western Pacific.

4. Summary

The relationship between interannual variations in the subtropical atmospheric patterns,
equatorial Pacific heat content anomalies, and eastern equatorial Pacific SSTs is investigated
here. Previous results indicate that the initiation and onset of ENSO events in the tropical Pacific
can be partly associated with both seasonal-mean variations in the central subtropical/extra-
tropical Pacific sea-level pressure fields as well as in the subsurface heat content anomalies over
the western equatorial Pacific. Here we find that years in which there are: 1) positive (negative)
boreal-summer/fall subsurface temperature anomalies in the western equatorial Pacific and; 2)
negative (positive) anomalies in the boreal-winter sea-level pressure fields over the subtropical
central North Pacific, tend to result in positive (negative) mature ENSO events during the
following boreal winter (i.e., 12-18 months later). When the intervening sea-level pressure
anomalies are of the same sign as the preceding heat-content anomalies, the correlation of the
heat-content anomalies with the following ENSO state disappears; there is still some relation
between the boreal-winter sea-level pressure anomalies and the ENSO state the following year,
however the correlation is smaller and the ENSO events tend to be weaker under these scenarios.

Here we briefly discuss one possible physical interpretation for these statistical findings.
Results presented above suggest that for periods with pre-existing subsurface warm water
anomalies in the equatorial western Pacific, there is a much higher probability that these are
followed by positive ENSO events if, during the intervening boreal-winter period, there is a
decrease in the sea-level pressure fields over the central subtropical Pacific. This decrease in
subtropical sea-level pressures in turn is related to a decrease in the trade winds over the tropical North Pacific (Chan and Xu, 2000; Anderson, 2003; Anderson, 2004). As described in Li (1997), the curl of these wind-stress anomalies over the central Pacific (c.f. Fig.10a,b from Anderson, 2003 but reversed) can produce a convergence of meridional Sverdrup transport into the central equatorial Pacific and a subsequent increase in the basin-average thermocline depths. This process can allow for the eastward translation of the western equatorial heat content anomalies into the central and eastern Pacific (Li, 1997) as seen in Figure 2, where they can effectively modify the overlying sea-surface temperature fields. Conversely, during years in which there is an increase in the sea-level pressure fields over the central subtropical Pacific during the intervening boreal winter, the subsequent Sverdrup discharge might limit the eastward translation of the pre-existing heat-content anomalies in the western Pacific and hence the initiation of ENSO events. Further diagnosis of these interactions within climate simulation models is currently underway.

It is important to note that we do not argue that the initiation of ENSO events is solely related to variability in the preceding equatorial Pacific heat-content fields and subtropical Pacific sea-level pressure fields. A multi-variate linear regression of the two indices against the NINO3.4 index gives a correlation of $|r|=0.72$, which indicates that only about half of the variance of the ENSO system can be captured by these two indices.

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provided by NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA, from their Web site at: http://www.cdc.noaa.gov/ClimateIndices/index.html.
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Figure Legends

Fig. 1 Time-series of boreal-winter (November-March) seasonal mean SLPI anomalies, defined as the area-average grid-point sea-level pressure anomalies over 140-175W and 10-25N. (circles); time-series is shifted forward 12 months (i.e. the 1948 value is plotted in 1949). Also shown is the time-series of the January-March NINO3.4 index for the current year (solid line). Both time-series normalized by their respective interannual standard deviations.

Fig. 2 Three-month mean heat storage (HST) anomalies from 0-400m regressed against the seasonal mean SLPI anomalies for November-March. Initials give the respective three-month period; Numeral indicates lead/lag relation to January period of SLPI (-1: Year prior to SLPI; 0: Year concurrent with SLPI). All regression values are multiplied by -1 to represent the evolution preceding positive ENSO events. Contour interval is 15x10^7 Ws/m^2; minimum contour is +/- 15x10^7 Ws/m^2. Thick line represents regions with correlations greater than |r|=0.35 (approximately the 95% confidence limit using bootstrapping methodology – see text). (a) Jan-Mar HST preceding the SLPI; (b) Mar-May HST; (c) May-Jul HST; (d) Jul-Sep HST; (e) Sep-Nov HST; (f) Nov-Jan HST concurrent with the SLPI; (g) Jan-Mar HST; (h) Mar-May HST

Fig. 3 (a) Scatter-plot of the seasonal-mean November-March SLPI with the January-March NINO3.4 index the following year, plotted only for those years in which the SLPI has the opposite sign as the heat storage (HST) anomalies averaged over the region 160-180E, 5N-5S during the previous June-October period. (b) Same as (a) except for only those years in which the SLPI has the same sign as the preceding heat storage anomalies. (c) Scatter-plot of the heat storage anomalies averaged over the region 160-180E, 5N-5S during June-October with the
January-March NINO3.4 index 18 months later, plotted only for those years in which the following November-March SLPI has the opposite sign as heat storage anomalies. (d) Same as (c) except for only those years in which the SLPI has the same sign as heat storage anomalies.
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