On the Instability of the Tropical Western Pacific Warm Pool During the Boreal Winter and Spring

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A source of instability in the Tropical Western Pacific (TWP) warm pool is shown to be due to sea surface elevation variations caused by changes in the zonal sea-surface temperature (SST) gradient and the changes in the Pacific Ocean basin length in relation to the warm pool latitudinal location. The variation of the sea-surface elevation is measured by using the thermocline depth response calculated from a two-layer ocean. The warm pool is shown to be barely at equilibrium during the boreal late winter and early spring by comparing the measured thermocline at 110°W, 0°E with the calculated thermocline depth. Based on this analysis, a failure or reversal of the climatological zonal winds are apparently not a necessary precursor for the instability of the warm pool and initiation of a warm event. A warm event can be initiated by an increase in the size of the warm pool and/or an increase in zonal SST differences during the boreal winter/spring. This mechanism could be an alternate mechanism for El-Nino Southern Oscillation (ENSO) dynamics to that postulated by Bjerknes (1969).

Introduction

ENSO system dynamics of the equatorial Pacific postulates that the gradient of SST along the equator is the cause of the Walker circulation (Bjerknes 1969). Coupling of the ocean and atmosphere is considered to occur from the SST gradient forcing the zonal wind (Cane et al. 1997, Wallace et al. 1998). The precursor to an El-Nino event is thought to result from the collapse of the easterly zonal winds, the cause of which is unclear, but most likely to occur during the boreal winter. This report postulates that the movement of the warm pool (Yan et al. 1997), and variation in the Pacific Ocean basin length at warm pool centroid, and zonal SST differences across the Pacific, causes the equatorial Pacific ocean to be barely at equilibrium during the boreal late winter and early spring.

Data

SST 1° x 1° resolution data from the Hadley Met Office, UK, (Rayner et al. 1996) was used to calculate the TWP warm pool centroid, average temperature and area east of 120° with SST > 28.0° C (after Yan, et al. 1997). Across from the warm pool centroid location, the average temperature in the eastern Pacific was calculated for a 20° section stretching westward from the coast of South America and extending ±5° toward the north and south. A climatology of these values was also created from 1982 to 1996 for "normal years" that excluded warm event years and the months after April in the year prior to a warm event.

A zonal wind index of surface winds from 8°N - 8°S, 150°E - 140° was obtained from the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) (personal communication T. Mitchell). Outgoing longwave radiation (OLR) data are monthly anomalies of 1970 to 1995 climatology for 160°E - 160°W from the Climate Prediction Center (personal communication J. Janowiak). A climatology of OLR and wind index was created for "normal years."

The Tropical Ocean Global Atmosphere-Tropical Atmospheric Ocean (TOGA-TAO) (McPhaden 1995) buoy at 110°W, 0°N has the longest record of data, with daily temperature and depth data extending from 1981 to 1996. The daily data was averaged to monthly values, and a linear interpolation of the temperature was done to obtain the 20° C depth. The monthly 20° C thermocline depth (depths of 1 m, 15 m, 25 m, 35 m, 40 m, 50 m, 60 m, 75 m, 100 m, 120 m, 140 m, 150 m) was interpolated wherever the gap was only a month. Buoy data commenced from February 1980, and the January 1980 data was calculated by examining the trend in thermocline depth for the months of January to March in 1981 and comparing it with that of February 1980 to March 1980. Monthly averaging was then done to obtain a "normal year" climatology of the 20° C isotherm depth at 110°W, 0°N.

Results and Discussion

In Figure 1a the annual march of the SST difference shows little correspondence to the annual march of the wind index in Figure 1b. In Figure 1a, a climatalogy of OLR and wind

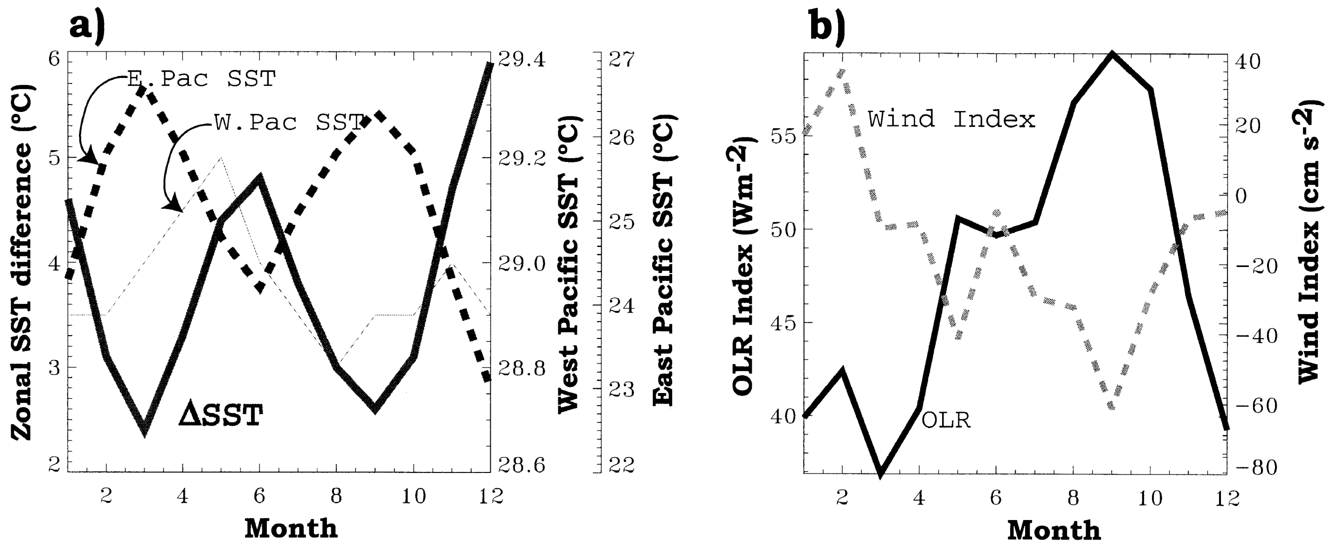

Figure 1. a) Climatology for "normal" years between 1982-1996 of SST for western Pacific (thin line) and eastern Pacific (dotted line) and the difference (solid line). b) Climatology for "normal years" of wind index for 8°N-8°S, 150°E-140° (up to 1992) and OLR for 160°E-160°W.

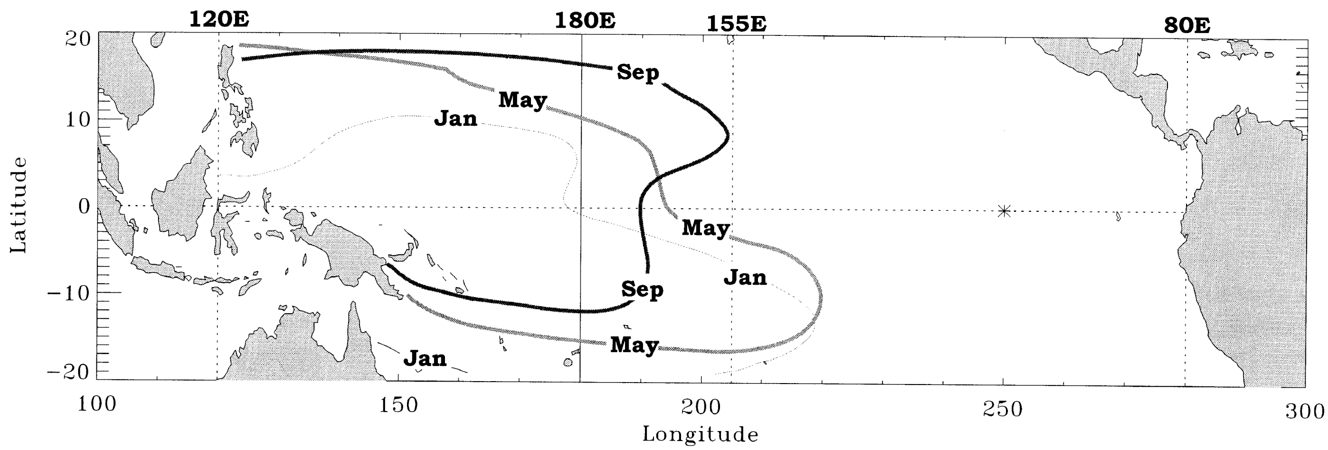

Figure 2. Contour of 28° C SST east of 120°E for January, May, and September of "normal year" climatology.

index has been created for "normal years." It can be seen in Figure 2 that during the "normal years," the monthly march of OLR precedes that of the wind index by about a month and that they are highly negatively correlated. However, as seen in Figure 1a, SST gradients are caused mainly by changes in the eastern equatorial Pacific. For example, consider eastern SSTs that are colder than normal with no changes in the western Pacific temperature. This would be an anomalous SST difference, but would not be reflected in the TWP temperature anomaly. The data indicates that ocean and atmosphere coupling by SST differences in forcing zonal wind patterns is tenuous (Figures 1a, b).

The warm pool system moves with a sun synchronous cycle, with the warm pool being north of the equator during the boreal summer and autumn (Figure 2) (Yan et al. 1997). The 28° C SST contours for January, May, and September in Figure 3 are from a monthly "normal year" average from 1970 to 1996 of the SST data. In the boreal winter and spring, the warm pool is south of the equator, as seen in the "normal year" climatalogy (Figure 2 and Figure 3a). When the warm pool is north of the equator, the centroid is further toward the west as seen by the 28° C isotherm in Figure 2, and zonal centroid location in Figure 3b. The centroid location climatalogy in Figures 3 and 4 also shows that the

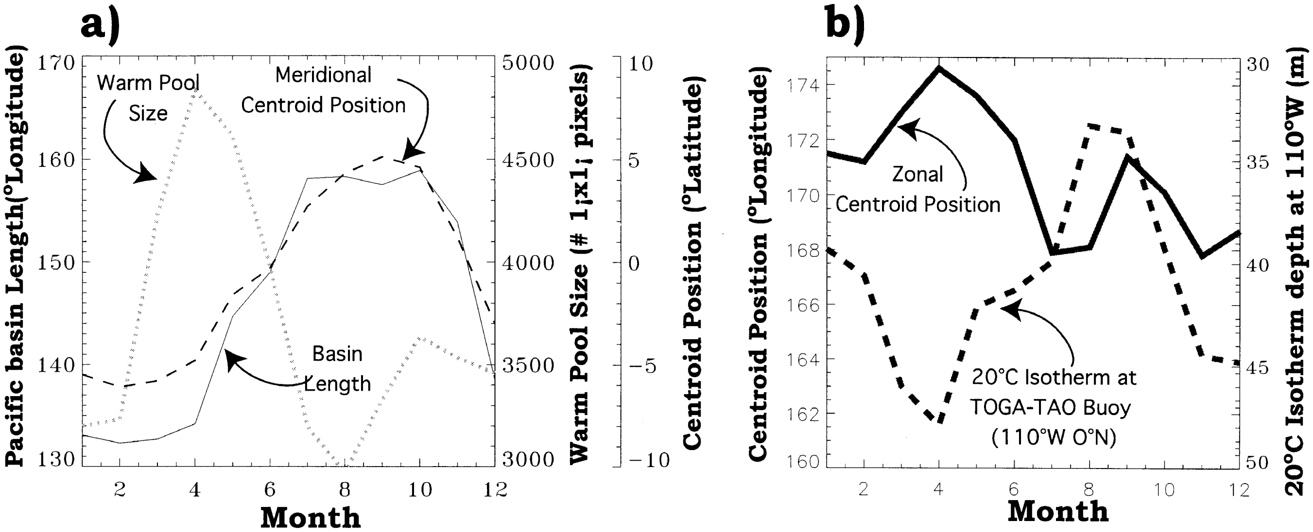

Figure 3. a) "Normal year" climatology of warm pool (> 28° C) size east of 120°E (dotted line), meridional centroid position (dashed line) and basin width (thin line) of Pacific at latitude of centroid. b) "Normal year" of zonal position (thin line) of warm pool centroid and 20° C depth at 110°W, 0°N TOGA-TOA buoy (dashed line).

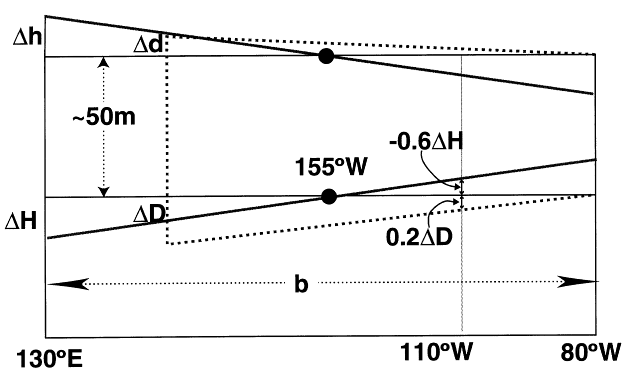

Figure 4. Schematic of the equatorial Pacific system reflecting changes in thermocline depth and sea surface elevation. Dashed line is the sea surface and thermocline response to changes in effective basin length to the Warm Pool location. The solid line represents the sea surface and thermocline change to SST gradients across the Pacific.

eastern most position of the centroid occurs during early boreal spring months. The northern position shown by meridional centroid position in Figure 4 of the warm pool also brings a reduction in the size of the warm pool as depicted by 28° C SST contour in Figure 3 and size in Figure 4. The reduction in size during the boreal summer and autumn is linked to an increase in basin length and is seen in Figure 2 and Figure 3a. The effective basin length for the warm pool at centroid latitude during each month, was calculated by using 50-m isobath in the eastern and western Pacific from the ETOP5 data set that had been smoothed twice with a 5° boxcar filter. The warm pool zonal centroid location climatalogy for a "normal year" and the 20° C depth at 110°W, 0°N are strongly negatively correlated, with the zonal centroid location leading by about a month in the boreal summer and autumn (Figure 3b).

Conceptually, the warm pool can be thought of as a two-layer system where the basin length changes with each month, as does the zonal SST gradient (Figure 4). Based on meridional centroid movement, the mean basin length is 150° of Longitude, with the western position at 130°E and eastern position at 80°W (Figure 5). The change in the length of the basin for the warm pool and the zonal SST gradient bring about changes in the thermocline depth and sea surface elevation. The total dynamic height difference is from three mechanisms: 1) that due to the zonal SST difference; 2) that due to changes in basin length; and 3) that caused by wind stress over and above what is needed to maintain the dynamic height due to the first two causes. If the warm pool system in the TWP is to be just at equilibrium, the wind stress just balances the dynamic height difference due to basin length changes Δd and zonal SST differences 2Δh. Then the total dynamic height is

$$\Delta h_{tot}=2\Delta h+\Delta d$$

The dynamic height difference due to east-west temperature difference can be given by

$$2\Delta h_{130E} = (T_e-T_w)*\alpha*H$$

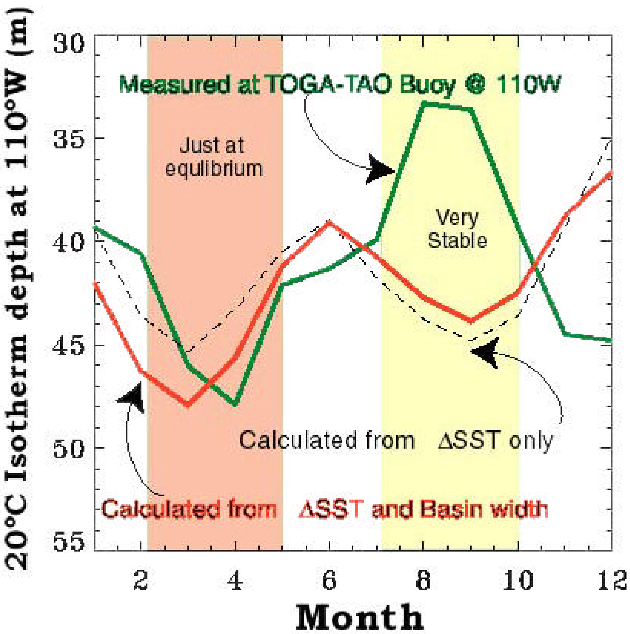

Figure 5. Calculated thermocline depth due to "normal year" zonal SST difference (dashed line), due to basin length changes and zonal SST difference (solid line) and measured at 110°W, 0°N TOGA-TAO buoy (thin line). (For a color version of this figure, please see *http://www.arm.gov/docs/documents/technicalconf_9803/barr(3)-98.pdf*.)

where $2\Delta h$ is the dynamic height difference due to zonal SST differences, and Δh_{130E} is the sea surface elevation in the TWP. H is the mixed layer height of 50 m, and α the expansion coefficient of water, of 3×10^{-4} K^{-1}.

Now, considering the Pacific basin to be a two-layer system, the change in the 20° C isotherm depth ΔH_{130E} response in the TWP, due to the dynamic height difference caused by the zonal SST difference, can be given by Meyers (1979) and Wyrtki (1979).

$$\rho_w \, \Delta h_{130E} = (\rho_{20°C} - \rho_{(w+e)/2}) \, \Delta H_{130E}$$

where $\rho_{20°C}$ is the density below the 20° C isotherm and $\rho_{(w+e)/2}$ is the mean density of the upper layer taken to be the mean of the eastern Pacific and TWP waters. The differences are considered to be primarily due to temperature, with uniform salinity of ~35 ppt (Picaut et al. 1996). Salinity in the TWP is fresher, with values of about 34 ppt, and higher salinities in the central Pacific of about 35.25 ppt (Picaut et al. 1996). The existence of the salinity front has been shown to be the barrier that prevents mixing of cold deeper water as well as defining the nature of zonal SST gradients (Picaut et al. 1996). East of the salinity front, there is a progressive decline in SST, whereas west of the salinity there is almost zero SST gradient (Picaut et al. 1996).

Because it is the warm pool that moves north and south in a sun synchronous cycle, the greatest changes in the 20° C isotherm depth due to basin length changes occur in the TWP. Allowing the 20° C depth change by basin width change to decrease linearly toward the east, the thermocline depth change in the east is given by

$$\Delta D_{130W} = H\,(300/b - 2)$$

The dynamic height response to this change in thermocline depth will be

$$\rho_w \, \Delta d_{130E} = (\rho_{20°\,C} - \rho_{(w+e)/2}) \, \Delta D_{130E}$$

The isotherm response to SST difference-induced changes pivots around the mid point of the system, 155°W. In contrast, changes in the basin length cause the isotherm to get depressed or elevated assuming no change in the meridional extent of the warm pool. From basin geometry considerations shown in Figure 4, the total change in the 20° C depth at 110°W can be given relative to the 20° C depth at 130°E.

$$\Delta H_{tot\,,110W} = -0.6\Delta H_{130E} + 0.2\Delta D_{130E}$$

It is clear that a 20° C depth increase at 130°E due to an increase in the SST gradient will cause a reduction in the 20° C depth at 110°W. On the other hand, an increase in 20° C depth at 130°E due to basin length changes will cause a 20° C depth increase at 110°W as well. Using the "normal year" climatology of SST differences, the 20° C depth change ranges from 7.7 m in March to 25 m in December. The basin length changes gives a 20° C depth change ranging from -5.6 m in October to 13.4 m in February.

During "normal years," the equatorial Pacific system is almost exactly in equilibrium during the late boreal winter and early spring months with the wind stresses appearing to just balance the thermocline depth, whereas in the boreal summer and autumn the system is very stable with wind-stress appearing to cause thermocline depth changes over and above that caused by SST differences and basin width changes (Figure 5). In Figure 5 is a plot of the calculated 20° C depth at 110°W, due to SST differences and basin length changes. The close correspondence of the calculated 20° C depth and that measured by the TOGA-TAO buoy during late winter to early spring probably indicates that the system is just balanced by the wind stresses (Figure 5). The lag of the measured depth by about a month probably indicates the time scale needed for the basin to adjust to changes in basin length and SST differences (Figure 5). Late winter and early spring are also the times when basin length significantly influences the 20° C depth of the system (Figure 5). During the boreal summer and autumn, the measured 20° C depth is shallower than that calculated due to zonal SST differences and basin length changes, indicating a very stable condition and probably indicates that zonal easterly wind stresses are over and above that needed to maintain the sea surface elevation due to zonal SST gradients and basin length changes (Figure 5). The excess wind stresses creates a deep 20° C depth in the western part of the Pacific, in response to an increased sea surface elevation. Analogous to the response of the 20° C depth at 110°W to increasing zonal SST differences, the increase in 20° C depth at 130°E due to wind stresses brings an attendant decrease in thermocline depth at 110°W. The excess wind stresses during the boreal summer and fall results in a very stable equatorial Pacific.

The equatorial Pacific is just at equilibrium during the late boreal winter and early spring during normal years, with the climatological wind stresses appearing to just balance the sea surface elevation due to zonal SST gradients and basin length changes. The stability can be compromised, by a decrease in wind stresses, an increase in volume of the warm pool, an increase in zonal SST differences or any combination of the aforementioned causes. The average size and temperature of the warm pool has been increasing since 1984 to 1987 (Yan et al. 1992). This may explain the observation of increased El-Nino warm events and reduced La-Nina events since 1976 (Trenbeth and Hoar 1997). The extended El Nino of 1991 to 1994 may also be explained in similar fashion. The extent of these warm episodes did not extend to the NINO 1+2 regions in the eastern Pacific to sufficiently raise the temperatures significantly above normal. This can be interpreted as reduced dissipation of the warm pool heat content and size, leading into another warm event in the following year, and in the case of the 1991 to 1994 a prolonged set of weak warm events (Wallace et al. 1998).

Instability of the equatorial Pacific system during the boreal spring can apparently be explained in the context of the "normal year" climatalogy of zonal SST differences and basin length. A reduction in the easterly winds is not a necessary precursor for an warm event to occur, and can be induced by increases in warm pool size. Including these concepts, special basin-width changes of the Pacific in numerical models may improve the success of predictions made by models during the boreal spring (Cane et al. 1986).

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