Detection of the Eastern Edge of the Equatorial Pacific Warm Pool Using Satellite-Based Ocean Color Observations

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Abstract

The analysis of satellite-based ocean color data shows that low concentrations of surface chlorophyll-*a* (chl-*a*) found in the equatorial region of the Pacific Ocean varies in phase with the eastern edge of the warm pool. As is true for high sea surface temperatures, the existence and maintenance of these low concentrations are linked to the upper ocean stratification due to salinity. The present study also establishes the quasi permanence of a frontal zone in chlorophyll-*a* separating the regimes of the western region and the eastern-central cold tongue and, through the identification of this front in satellite-based ocean color data, it provides, for the first time, a reliable method for locating the eastern edge of the warm pool from surface observations only. Finally, the recognition of this front offers the opportunity to define a simple and robust index of the horizontal extension of the equatorial Pacific warm pool within the context of the ENSO variability.

1. Introduction

The processes controlling the warm sea surface temperatures (SSTs) permanently found in the western equatorial Pacific and known as the warm pool (Wyrtki 1989), are central to defining climate and determining the character of large scale and deep atmospheric convection. Detection of very high SST (around 30°C) within the warm pool is associated with the strong intraseasonal variability of hot events (Qin et al. 2007). Zonal displacement of the warm pool is also important for the onset of the El Niño Southern Oscillation (ENSO) phenomenon (Picaut et al. 1996; Ando and Hasegawa 2009). The separation between the cold tongue in the eastern-central equatorial Pacific and the warm pool relies at the heart of the revised delayed oscillator theory of ENSO proposed by Picaut et al. (1997). The variability of the eastern edge of the warm pool is thus crucial to understand and to monitor within the context of seasonal-to-interannual climate variations. However, the eastern edge of the warm pool along the equator is not properly defined by a front in SST but has distinct hydrological features and ecosystem dynamics (Le Borgne et al. 2002).

Several studies have attempted to find a simple way to identify the eastern edge of the warm pool (e.g., Picaut et al. 2001). Although the warm pool is characterized by warm, rainfall-induced low salinity, oligotrophic waters and the cold tongue by cold, high-salinity, mesotrophic waters, this task has proven more difficult than expected. The problem has been that although the eastern edge of the warm pool may be well defined by a convergence of zonal current (Picaut et al. 1996; Maes et al. 2004) or by biogeochemical properties (Eldin et al. 2004), it cannot be reliably identified by more readily observable phenomena such as SST or even sea surface salinity (SSS) front. The fact that Eldin et al. (2004) were unable to identify a signature in SSS is particularly significant in this respect. Another feature of the warm pool regards the salinity stratification within the upper ocean mixed layer that is known as the barrier layer (Lukas and Lindstrom 1991). Considering the temperature and salinity profiles collected by the Argo floats, Maes et al. (2006a) and Bosc et al. (2009) were able to document the quasi permanence of the barrier layer near the eastern edge of the warm pool. More importantly, both studies show that the presence of thick barrier layer is associated with the occurrence of very high SSTs. However, before the advent of the Argo data in early 2000, subsurface observations are too sparse to document the long term variability of subsurface salinity stratification associated with the eastern edge of the warm pool along the equatorial Pacific Ocean.

In this study, we intend to document the possibility of using satellite-based observations of chlorophyll-a (the major photosynthetic pigment found in living algae, hereafter "chl-*a*") to detect, accurately and consistently, the eastern edge of the warm pool in the equatorial Pacific Ocean. This possibility arises because of the relationship between low chlorophyll values (typically less than 0.1 mg m^{-3}) and the subsurface salinity stratification characterizing the warm pool region. This latter point will be verified first with in situ observations collected during oceanographic cruises. Section 2 describes the different data sets we used, Section 3 presents the results of our analysis and Section 4 concludes with a discussion.

2. Data and study method

Observations collected on oceanographic cruises provide the means to identify accurately the eastern edge of the equatorial Pacific warm pool through a combination of several parameters. To this end, we use the observations collected during the Frontalis-3 cruise staged from the IRD center in New Caledonia on the R/V Alis in April-May 2005 (Maes et al. 2006b). Chlorophyll samples were pressure filtered through Whatman GF/F filters and stored in liquid nitrogen and, upon return to Nouméa, analyzed by fluorometry after methanol extraction (Le Bouteiller et al. 1992). The vertical profiles of salinity and temperature are provided by the autonomous Argo floats that have been analyzed previously by Maes et al. (2006a).

The satellite data used in this study are the chl-*a* concentrations obtained from the NASA/Goddard Earth science (GES-DAAC). The datasets are issued from the August 2005 reprocessing version 1.1 of MODIS/Aqua (http://oceancolor.gsfc.nasa.gov/REPROCESSING/Aqua/R1.1) and from the July 2005 reprocessing version 5.1 for SeaWiFS (http://oceancolor.gsfc.nasa.gov/REPROCESSING/SeaWiFS/R5.1). We used the L3 weekly products on a regular grid of 9 by 9 km and all the data were used regardless of flag settings.

3. Results and discussion

We begin with a comparison of satellite-based and in situ observations. The middle panel of Fig. 1 shows the surface chl-*a* concentration along the equator observed between April 30 and May 3, 2005, during the Frontalis-3 cruise and the weekly composites of SeaWiFS and MODIS/Aqua data measured during the same week. Along the entire equatorial section the agreement is quite good, although the strongest observed values, eastward of

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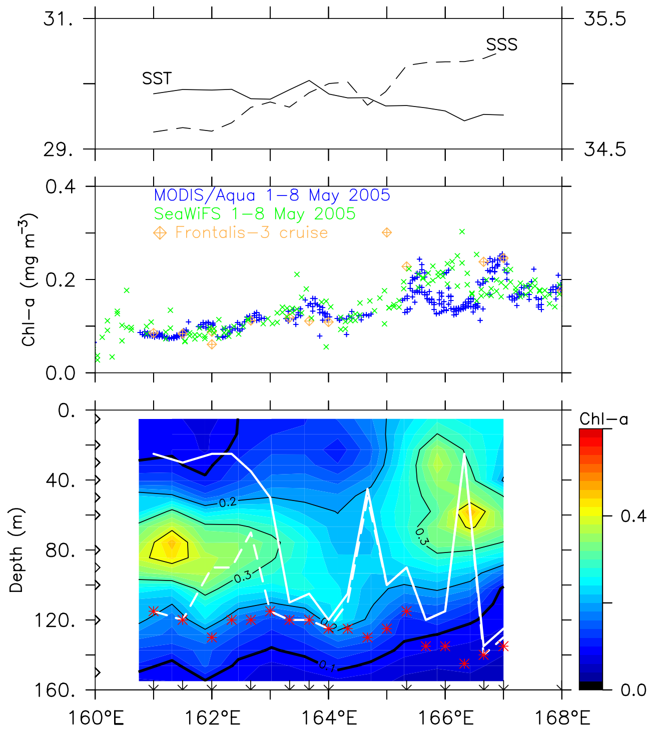

Fig. 1. The surface distributions (top) of SST, SSS and chl-*a* along the equator measured during the Frontalis-3 cruise in May 2005. In the middle panel, the values derived from SeaWiFS and MODIS/Aqua are represented by the green and blue symbols, respectively. The bottom panel shows the vertical distribution of chl-*a*, the position of the main thermocline (stars) and the bounds of the strongest salinity stratification in the upper water column (defined through the Brunt-Väisälä frequency, see Maes, 2008 for details). The symbols on the x and y axes show the positions of the stations and sampling depths. The contour interval for the chl-*a* concentration is equal to 0.1 mg m^{-3}.

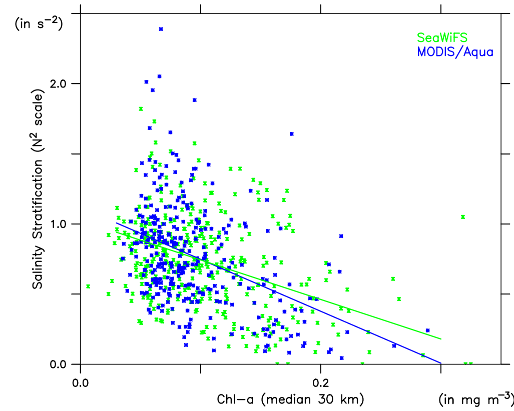

Fig. 2. Scatter diagram of the chl-a concentration (mg m^{-3}) vs. ocean salinity stratification (scaled by 10^{-4} s^{-2}) for the period 2002–2004, where the green and blue symbols represent SeaWiFS and MODIS/Aqua values, respectively. The salinity stratification is estimated from the vertical profiles of temperature and salinity collected by the Argo floats. Each corresponding value for the chl-*a* concentration is obtained through a median filter of all pixel values within a 30 km radius around the position of the profile.

164°E, are slightly underestimated by the satellite-based products. In agreement with other physical and biogeochemical parameters, the chl-*a* observations show that the eastern edge of the warm pool is located near 165°E (Maes et al. 2006b; Maes 2008), a position that is not detectable from SST, as expected, but also not quite easily from SSS field only (Fig. 1, top).

The vertical distribution of the chl-*a* concentration (Fig. 1, bottom) reveals another very interesting feature. In the western part of the section, the chl-*a* profiles exhibit a deep maximum between 60 and 100 m depth, a region that is clearly inside the warm pool. Based on the static stratification through the N^2 computation, the region where the maximum values in chl-*a* are observable is also characterized by a strong stability of the water column that is controlled entirely by the salinity (white lines in Fig. 1). This stratification produces the same net effect as the barrier layer (Maes 2008) and prevents cooling surface layer from the main thermocline, which here is deeper than 100 m depth (identified by the red stars). The nutrients from below are also not able to supply the euphotic zone and thus to enhance productivity in the surface layers. If the subsurface maximum is probably controlled by the nutrient and light distribution, the low values in chl-*a* at the surface of the western Pacific warm pool may persist only as a consequence of the salinity stratification of the upper layers. Using zeaxanthin pigments of the phytoplankton, Matsumoto and Ando (2009) also found a close relationship between the physical and surface biological processes. In the following we will see how we can exploit this relationship to determine the longitude corresponding to the eastern edge of the warm pool.

The consistency between the salinity stratification at depth and chlorophyll values at the surface suggests the mechanism that produces low values of chl-*a* in the warm pool surface. To extend such a result within the entire warm pool region and beyond, we used the computation of the salinity stratification from the temperature and salinity profiles observed by the Argo floats during the period 2002–2004. These values are then associated with the satellite-based chl-*a* concentration values through application of a daily median filter to data within a 30 km radius of each Argo profile. The scatter diagram for the pairs (salinity stratification, chl-*a* concentration) is shown in Fig. 2. It is based on data for the western Pacific Ocean only and includes 365 SeaWiFS and 324 MODIS/Aqua values. The larger values of chl-*a* are generally associated with low salinity stratification and as stratification increases the chlorophyll concentration decreases. For chl-*a* concentrations below about 0.1 mg m^{-3} the salinity stratification is always non-zero but the scatter is large, indicating that the salinity stratification is only one process acting on the complex evolution of chl-*a* within the warm pool region. The global scatter also reflects that on each side of eastern edge of the warm pool, the biogeochemical dynamics and ecosystems are quite different. This result however suggests that an accurate determination of the eastern edge of the warm pool based on a pivotal value of chl-*a* may be useful. This latter point will be investigated in details.

We found that we could locate a significant transition point in the chl-*a* concentration by searching east to west for the longitude where the satellite-derived chl-*a* concentration first falls below 0.1 mg m^{-3}. The procedure consists to apply a median filter at 30 km scale to remove the effects of the small scale structures. Using this procedure to determine the transition point, we computed the time-averaged mean and standard deviations of the chlorophyll data around the transition point for the period of July 2002 through 2006. Figure 3 shows the results for longitudes extending from 15° west to 15° east of the transition point. Westward of the transition point, the chl-*a* concentration is nearly constant and remains below 0.1 mg m^{-3}. Eastward of the transition point, the concentration increases abruptly over 5–6 degrees to approach a constant value just below 0.2 mg m^{-3}. These characteristics of the SeaWiFS and MODIS/Aqua curves on either sides of the central position suggest that this transition point consistently represents a frontal region that separates the oligotrophic and mesotrophic regimes of the western Pacific warm pool. It appears also to be associated with the eastern edge of the warm pool. The warm pool region is characterized by a low temporal standard deviation and this weakness is apparent over 10° of longitude westward of the chl-*a* transition point. The two satellite-based data sets agree extremely well in their mean structure and only the standard deviation is slightly higher for SeaWiFS than for MODIS/Aqua in the warm pool. Indeed, following the recommendations of the

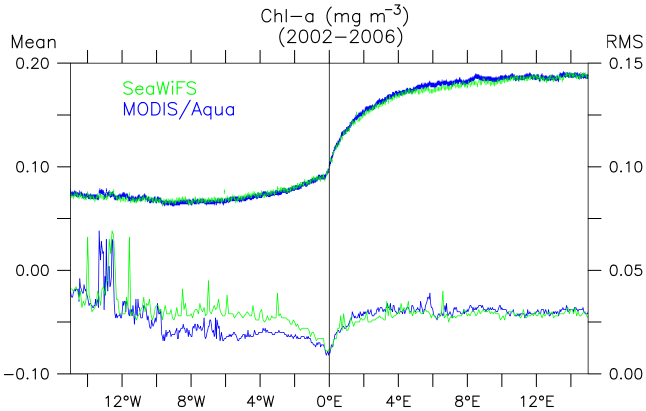

Fig. 3. Time-averages (left axis, upper lines) and standard deviations (right axis, lower lines) of chl-*a* concentrations (mg m^{-3}) for locations within 15 degrees east and west of the 0.1 mg m^{-3} value. The vertical black line represents the central position. The time period for the calculations begins in July 2002 when both ocean color archives are available, and continues through 2006.

Ocean Biology Processing Group (Franz 2005), Pottier (2006) determined over the year 2003 a standard deviation for chl-*a* concentration derived from SeaWiFS of 0.01 to 0.02 mg m^{-3} higher than that from MODIS/Aqua within the tropical Pacific (10°N−10°S).

The ability to locate accurately the eastern edge of the warm pool provides a tool to monitor the progress of ENSO events. The time-longitude variation of the 0.1 mg m^{-3} transition point is shown in Fig. 4 superimposed on contours of the equatorial SSTs (Reynolds et al. 2002), higher than 28°C. With the exception of the end of 1997 through the mid-1998 period when the exceptionally strong ENSO event completely disrupted the biogeochemical properties in the equatorial Pacific Ocean (Chavez et al. 1999), the detection of the chl-*a* transition point is always possible and it is always found within waters warmer than 28°C. During this period, the chl-*a* transition point may be as much as 20° of longitude farther west than the 28°C isotherm. With respect to temporal variability, Table 1 shows that the position of the chl-*a* transition point varies in phase with the SST anomalies in the NINO3.4 box and with the Southern Oscillation Index (SOI). The correlations using the chl-*a* data are similar to those using the 28°C isotherm, giving us further confidence in our use of the satellite-based chl-*a* data for monitoring the eastern edge of the warm pool. Moreover, Fig. 4 shows that the important hot spots, defined by Waliser (1996) as SSTs warmer than 29.75°C, are systematically found westward of the chl-*a* transition point. A similar relationship was also reported by Maes et al. (2006a) and by Bosc et al. (2009) in their analyses of the variations of the eastern edge of the warm pool as defined by the SSS variability. Finally, we also note the consistency with the few estimates established from the three Frontalis cruises that differ in their position of the frontal zone by less than one degree in longitude (see the squares on Fig. 4).

4. Conclusion and perspectives

The co-variability of SST and chlorophyll along the equatorial Pacific Ocean has been known for some time and either could be used to mark the eastward progression of the western Pacific warm pool (e.g., Ryan et al. 2002). However, the choice of a specific isotherm value to represent the leading edge of the warm pool remains problematical. Following a suggestion of Picaut et al. (2001), we undertook the analysis of satellite-based chlorophyll data and found that the eastern edge of the oligotrophic waters characterized by a sharp transition of the chl-*a* concentration to values lower than 0.1 mg m^{-3}, could be identified with the eastern edge of the equatorial Pacific warm pool. During the

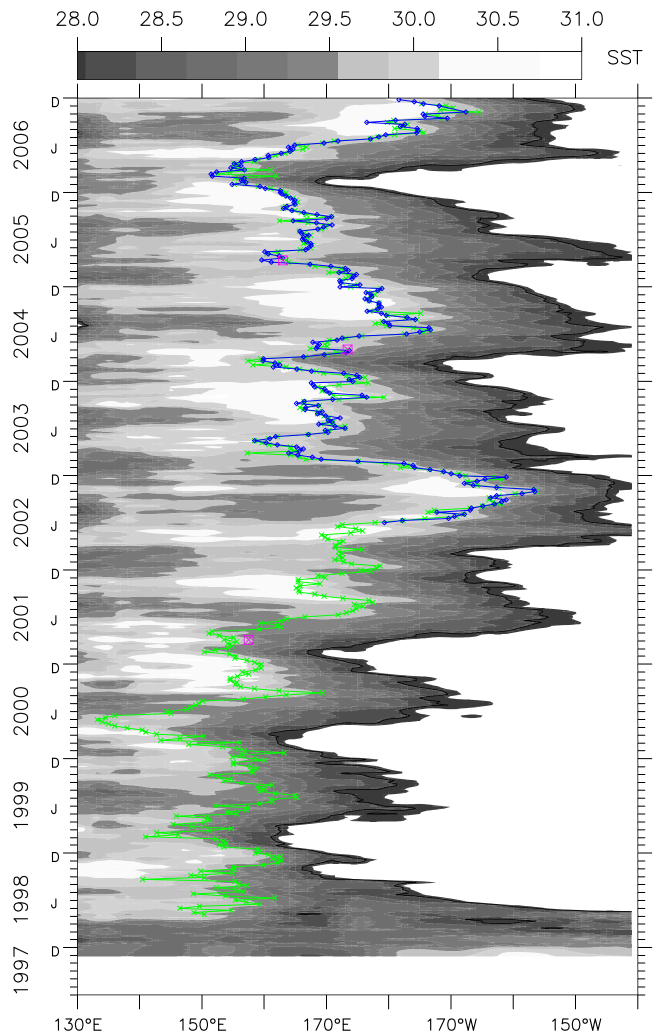

Fig. 4. Longitude vs. time plots along the equatorial Pacific Ocean between 130°E and 140°W. SST values warmer than 28°C are shown as shaded contours. The positions of the eastern edge of the warm pool as defined by the 0.1 mg m^{-3} value in chl-*a* are represented in green and blue for the SeaWiFS and MODIS/Aqua values, respectively. The pink squares show the position of the eastern edge of the warm pool estimated through a multi-parameter analysis of the Frontalis1-3 cruises.

Table 1. Correlation coefficients between the monthly position of the eastern edge of the warm pool as detected by the chl-a front or the 28°C isotherm, and the NINO3.4 SST anomalies (upper) and SOI (lower) for the two different satellite periods. Note that, for the same periods, the correlation coefficients between the NINO3.4 SST and SOI are equal to −0.73 and −0.92, respectively. The SOI and NINO3.4 SST data are provided by the NOAA Climate Prediction Center.

	28°C	MODIS/Aqua	SeaWiFS
08/2002−10/2006	0.83 −0.77	0.90 −0.64	0.90 −0.62
10/1997−10/2006	0.90 −0.92	no data available	0.88 −0.85

period 1998−2006 the zonal displacements of this transition point exhibit significant variations between 140°E and 160°W, that are consistent with previous analyses of the convergence zone of equatorial currents (Maes et al. 2004). Also, the correspondence between temperatures greater than 29.75°C and chl-*a* concentrations less than 0.1 mg m^{-3} emphasizes the importance of low chl-*a* concentrations in the heat budget of the region (Lewis et al. 1990). The warm pool region is a complex system whose the variations influence both physical properties and biogeochemical activities. The latter one could be used to detect conjointly with the physical parameters the eastern edge of the warm pool where air-sea interactions are so crucial. It is expected that a closer attention to the variability of the warmest SSTs in the warm pool could lead to a better representation of the large scale gradient that drives the lower circulation of the ocean-atmosphere coupled system of the equatorial Pacific Ocean.

The western equatorial Pacific Ocean is of primary importance in the climate system through its role in the deep organized atmospheric convection, in the circulation and stratification of the upper ocean layers and, ultimately, in ENSO variability. Although the western warm pool and the cold tongue of the central-eastern Pacific Ocean represent very different ocean regimes, it has been a challenge to identify a single parameter that marks the transition between them. The simple approach proposed here, based on the satellite-derived chlorophyll-*a* concentrations, succeeds in detecting this transition zone, with a high spatio-temporal accuracy not previously reported. Finer analyses of the eastern edge of the warm pool will be even available with the merging of different satellite-based products (Pottier et al. 2006). Finally, the fact that the chlorophyll-a transition point can be reliably determined throughout the 1997−2006 period suggests the nearly permanent presence of a frontal zone between the warm-pool and cold-tongue regimes and demonstrates the usefulness of the chlorophyll-*a* transition as an index of the zonal extension of the equatorial Pacific warm pool within the context of ENSO.

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References

Ando, K., and T. Hasegawa, 2009: Annual zonal displacement of Pacific warm pool in association with El Niño onset. *SOLA*, **5**, 149−152, doi:10.2151/sola.2009-038.

Bosc, C., T. Delcroix, and C. Maes, 2009: Barrier layer variability in the western Pacific warm pool from 2000 to 2007. *J. Geophys. Res.*, **114**, C06023, doi:10.1029/2008JC005187.

Chavez, F. P., P. G. Strutton, G. E. Friedrich, R. A. Feely, G. C. Feldman, D. G. Foley, and M. J. McPhaden, 1999: Biological and chemical response of the equatorial Pacific ocean to the 1997−98 El Niño. *Science*, **286**, 2126−2131.

Eldin, G., T. Delcroix, and M. Rodier, 2004: The frontal area at the eastern edge of the western equatorial Pacific warm pool in April 2001. *J. Geophys. Res.*, **109**, C07006, doi:10.1029/2003JC002088.

Franz, B. A., 2005: Methods for assessing the quality and consistency of ocean color products, Available online at http://oceancolor.gsfc.nasa.gov/DOCS/methods/sensor_analysis_methods.html, (accessed 2010.9.21).

Le Borgne, R., R. T. Barber, T. Delcroix, H. Y. Inoue, D. J. Mackey, and M. Rodier, 2002: Pacific warm pool and divergence: temporal and zonal variations on the equator and their effects on the biological pump. *Deep Sea Res. II*, **49**, 2471−2512.

Le Bouteiller, A., J. Blanchot, and M. Rodier, 1992: Size distribution patterns of phytoplankton in the western Pacific: towards a generalization for the tropical ocean. *Deep Sea Res.*, **39**, 803−823.

Lewis, M. R., M.-E. Carr, G. C. Feldman, W. Esaias, and C. McClain, 1990: Influence of penetrating solar radiation on the heat budget of the equatorial Pacific Ocean. *Nature*, **347**, 543−545.

Lukas, R., and E. Lindstrom, 1991: The mixed layer of the western equatorial Pacific Ocean. *J. Geophys. Res.*, **96**, 3343−3357.

Maes, C., 2008: On the ocean salinity stratification observed at the eastern edge of the equatorial Pacific warm pool. *J. Geophys. Res.*, **113**, C03027, doi:10.1029/2007JC004297.

Maes, C., J. Picaut, Y. Kuroda, and K. Ando, 2004: Characteristics of the convergence zone at the eastern edge of the Pacific warm pool. *Geophys. Res. Lett.*, **31**, L11304, doi:10.1029/2004GL019867.

Maes, C., K. Ando, T. Delcroix, W. S. Kessler, M. J. McPhaden, and D. Roemmich, 2006a: Observed correlation of surface salinity, temperature and barrier layer at the eastern edge of the western Pacific warm pool. *Geophys. Res. Lett.*, **33**, L06601, doi:10.1029/2005GL024772.

Maes, C., E. Kestenare, A. Ganachaud, F. Gallois, M. Rodier, D. Varillon, G. Eldin, R. Chuchla, and A. Lapetite, 2006b: Rapport de la mission FRONTALIS 3 à bord du N.O. Alis du 22 avril au 19 mai 2005, 22°S−2°N/161°E−172°E, IRD Nouméa. *Rapports de Missions*, Sciences de la Mer, Océanographie Physique, No 20, 167 pp.

Matsumoto, K., and K. Ando, 2009: Use of cyanobacterial pigments to characterize the ocean surface mixed layer in the western Pacific warm pool. *J. Mar. Syst.*, **75**, 245−252.

Picaut, J., M. Ioualalen, C. Menkes, T. Delcroix, and M. J. McPhaden, 1996: Mechanism of the zonal displacements of the Pacific warm pool: implications for ENSO. *Science*, **274**, 1486−1489.

Picaut, J., F. Masia, and Y. du Penhoat, 1997: An advective-reflective conceptual model for the oscillatory nature of the ENSO. *Science*, **277**, 663−666.

Picaut, J., M. Ioualalen, T. Delcroix, F. Masia, R. Murtugudde, and J. Vialard, 2001: The oceanic zone of convergence on the eastern edge of the Pacific warm pool: a synthesis of results and implications for ENSO and biogeochemical phenomena. *J. Geophys. Res.*, **106**, 2363−2386.

Pottier, C., 2006: Merging of ocean color data: Application in Operational Oceanography, PhD Thesis, University Paul Sabatier, Toulouse III, 320 pp.

Pottier, C., V. Garçon, G. Lamicol, J. Sudre, P. Schaeffer, and P.-Y. Le Traon, 2006: Merging SeaWiFS and MODIS/Aqua ocean color data in North and equatorial Atlantic using weighted averaging and objective analysis. *IEEE transactions on geoscience and remote sensing*, **44**, 3436−3451.

Qin, H., H. Kawamura, and Y. Kawai, 2007: Detection of hot event in the equatorial Indo-Pacific warm pool using advanced satellite sea surface temperature, solar radiation, and wind spped. *J. Geophys. Res.*, **112**, C07015, doi:10.1029/2006JC003969.

Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Wang, 2002: An improved in situ and satellite SST analysis for climate. *J. Climate*, **15**, 1609−1625.

Ryan, J. P., P. S. Polito, P. G. Strutton, and F. P. Chavez, 2002: Unusual large-scale phytoplankton blooms in the equatorial Pacific. *Prog. Oceanogr.*, **55**, 263−285.

Waliser, D. E., 1996: Formation and limiting mechanisms for very high sea surface temperature: Linking the dynamics and the Thermodynamics. *J. Climate*, **9**, 161−188.

Wyrtki, K., 1989: *Some thoughts about the West Pacific warm pool. Western Pacific International meeting and workshop on TOGA COARE*. J. Picaut, R. Lukas, and T. Delcroix, Eds, Nouméa, 833 pp.

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