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The tropical Western Hemisphere warm pool

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Abstract. The Western Hemisphere warm pool (WHWP) of water warmer than 28.5°C extends from the eastern North Pacific to the Gulf of Mexico and the Caribbean, and at its peak, overlaps with the tropical North Atlantic. It has a large seasonal cycle and its interannual fluctuations of area and intensity are significant. Surface heat fluxes warm the WHWP through the boreal spring to an annual maximum of SST and areal extent in the late summer/early fall, associated with eastern North Pacific and Atlantic hurricane activities and rainfall from northern South America to the southern tier of the United States. SST and area anomalies occur at high temperatures where small changes can have a large impact on tropical convection. Observations suggest that a positive ocean-atmosphere feedback operating through longwave radiation and associated cloudiness is responsible for the WHWP SST anomalies. Associated with an increase in SST anomalies is a decrease in atmospheric sea level pressure anomalies and an anomalous increase in atmospheric convection and cloudiness. The increase in convective activity and cloudiness results in less longwave radiation loss from the surface, which then reinforces SST anomalies.

1. Introduction

The tropical Eastern Hemisphere has the western Pacific warm pool comprised of the warmest sea surface temperature (SST) in the world oceans. The eastward expansion and westward contraction of warm waters in the western Pacific warm pool, associated with large changes in atmospheric convection, plays an important role in the evolution of the El Niño-Southern Oscillation (ENSO). The shift of this warm pool and convective activity leads to altered Walker and Hadley circulations and then affects global climate variations.

In this paper, we discuss the second-largest tropical warm pool on Earth — the Western Hemisphere warm pool (WHWP). *Weisberg* [1996] noted it at the 76th AMS Annual Meeting. As we will show, the WHWP also has a large annual cycle and significant interannual departures in area and intensity, although it does not undergo large anomalous zonal excursions such as occur in the western Pacific. The annual development and demise of the WHWP are associated with marked seasonal changes in tropospheric heat, moisture and stability over the inter-tropical Americas, resulting in contrasting wet and dry seasons and the annual development of tropical storms and hurricanes on both sides of Central America. Research on the WHWP and its relation to other climate phenomena is needed to understand its role in mediating the climate fluctuations of the Western Hemisphere.

We define the WHWP as the region covered by water warmer than 28.5°C. These are temperatures that have a significant impact on organized tropical convection [*Graham and Barnett*, 1987] and nearly always subtend a closed region. Unlike the western Pacific warm pool, which straddles the equator, the WHWP is entirely north of the equator. It is comprised of the eastern North Pacific (ENP) west of Central America, the Gulf of Mexico, the Caribbean, and the western tropical North Atlantic (TNA). Despite separation by the narrow Meso-American landmass, with possibly distinct ocean dynamics and thermodynamics on either side, the atmosphere may respond to the WHWP as a monolithic whole.

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2. Western Hemisphere Warm Pool Variability

Monthly SST, sea level pressure (SLP), surface wind, surface heat fluxes, and cloud cover of the *da Silva et al.* [1994] data from 1945-1993 are used in this paper. The seasonal variations of SST and surface wind are shown in Fig. 1. During the boreal winter, there is no water warmer than 28.5°C in the ENP or in the Intra-Americas Sea (IAS), i.e., the Gulf of Mexico and the Caribbean. The warm pool starts to develop in the ENP during the boreal spring. The warming expands into

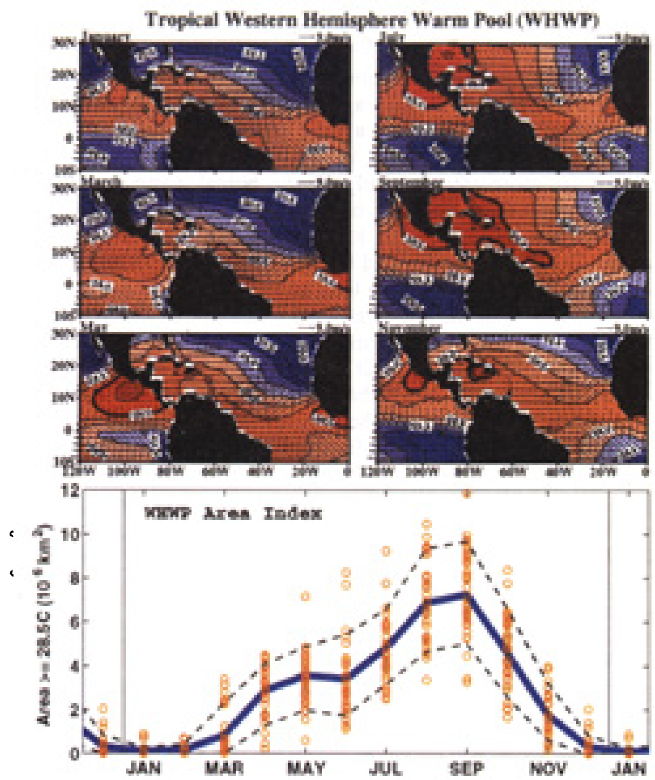

Figure 1. Upper: Seasonal distributions of SST and surface wind for the WHWP. The darker contour encloses water warmer than 28.5° (dark red). Lower: Seasonal variation (heavy curve) of the area index for water warmer than 28.5°C. Orange circles indicate the non-seasonal departures of the area index about the annual mean curve. The dashed curves are the averaged upper and lower quartiles of the departures.

the Gulf of Mexico in the boreal summer. Finally, the warm pool moves south into the Caribbean and expands eastward into the TNA in the boreal late summer and early fall. Both the Pacific intertropical convergence zone (ITCZ) and the Atlantic ITCZ are southernmost (northernmost) and weaker (stronger) in the boreal spring (fall). Off the west coast of Mexico and Central America, northeasterly winds prevail (weaken) before (during) the ENP warming whereas winds are easterly in the IAS. Just as the western Pacific warm pool is associated with easterly winds in the equatorial Pacific, the WHWP is correspondingly associated with easterly winds in the TNA.

The WHWP area of water warmer than 28.5°C is shown in the bottom panel of Fig. 1. It undergoes a strong annual cycle from nil in winter to a maximum in September. An early, smaller plateau of WHWP development in the late spring is associated with the May warming in the ENP, and precedes the more extensive warming on the Atlantic side. The larger September peak occurs when the warm pool has expanded southward to northern South America and eastward into the western TNA. The interannual variability of the WHWP appears as significant as its annual cycle. The range (dashed curves) of the non-seasonal index values (orange circles) is as large as the climatological mean area, while the upper quartile warm pool developments are about twice as extensive as for the lower quartile.

As the warm pool develops and the tropical rainy season begins, warmer SSTs are associated with a warmer and moister troposphere, reduced SLP, weaker easterly trade winds, less vertical wind shear and weakened subsidence aloft [*Gray*, 1968; *Knaff*, 1997]. The summer maximum of the WHWP area index is therefore consistent with the factors that favor a contemporaneous, strong rainy season and a higher frequency and greater intensity of tropical storms and hurricanes [*Landsea*, 1993; *Enfield*, 1996].

While the seasonal evolution of tropical storms in the Atlantic and eastern Pacific agrees in general with WHWP development, they do not always agree in their details. Typically, storms that form in the Caribbean and Gulf of Mexico are more frequent early in the boreal summer (July-August), while those that form off NW Africa and develop over the TNA are more frequent in August-September [*Landsea*, 1993]. The former are favored by, and consistent with an early warm pool development in the Gulf of Mexico, while the latter are consistent with the later, eastward extension of the WHWP into the western TNA in August-October. However, a second peak in Caribbean and Gulf of Mexico storms in late September until early November is not favored by the cooler SSTs there, at that time. The genesis locations of eastern Pacific storms are tightly confined to the ENP portion of the WHWP [*Gray*, 1968]. As with the Gulf of Mexico storms, they occur more frequently in early and late summer and they specifically do not coincide with the April-May maximum in the ENP portion of the WHWP. It may be that the greater frequency of the Pacific storms, in general, and the tighter confinement of their development to the region of warmest water, may actually be a factor in cooling the ENP SSTs in June-July. The heat given up by the upper ocean to a hurricane may reach values of up to six times normal, but the mixed layer cooling due to entrainment from below is even greater. The reduced ENP warm area in June-July, prior to the subsequent increase in area in the Gulf of Mexico, coincides with a mid-season trade wind increase and relaxation of summer rainy conditions over Central America [*Magaña et al.*, 1999]. Clearly, however, the relationships between warm pool development and the evolution of the tropical storm season are complex and we cannot argue simply that the genesis of storms is consistently more frequent at the times and places of warmest SST.

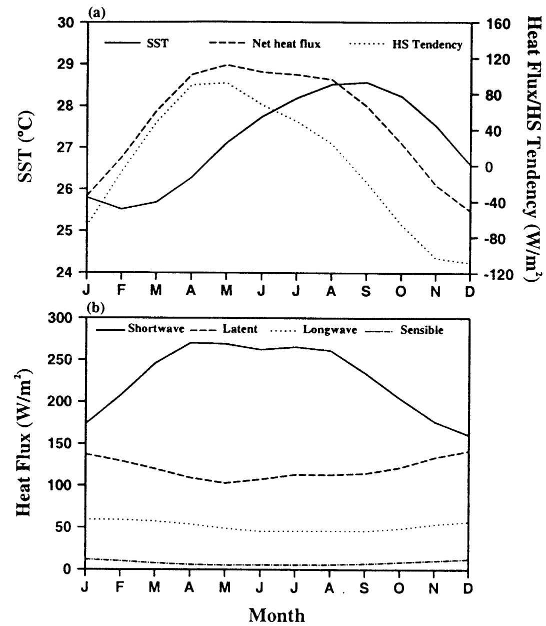

Figure 2. (a) Seasonal variations of SST, surface net heat flux (shortwave radiation minus latent heat flux, longwave radiation, and sensible heat flux), and heat storage (HS) tendency in the Western Hemisphere warm pool (WHWP) region (7°N-27°N, 110°W-50°W). (b) Seasonal variations of shortwave radiation, latent heat flux, longwave radiation, and sensible heat flux in the WHWP region.

3. Seasonal Heat Balance

To study what causes the WHWP variations, we need to define a WHWP intensity index. As suggested by Fig. 1, we define the intensity index as the average SST over the region of 7°N-27°N, 110°W-50°W. The seasonal variations of the WHWP SST and surface heat fluxes are shown in Fig. 2a. The WHWP SST is highest in September and lowest in February. What are the physical processes that control the SST seasonal variations? If the surface net heat flux is responsible for the SST variations, there will be a temporal phase lag between SST and net heat flux. Mathematically, if $\partial T / \partial t = Q_{Net} / \rho C_P h$ where T is SST, Q_{Net} is surface net heat flux, t is time, ρ is density of seawater, C_P is heat capacity of seawater, and h is ocean mixed layer depth, then there will be a 90° phase difference between T and Q_{Net}. For the annual cycle, the 90° phase difference means a 3-4 month phase lag between SST and net heat flux. Fig. 2a shows that maximum of net heat flux occurs around May which leads maximum SST in August-September by 3-4 months. In fact, the heat storage tendency ($\rho C_P h \partial T / \partial t$) is in phase with the surface net heat flux. These phase relationships suggest that the SST variations in the WHWP region are induced primarily by surface net heat flux. Notice that, during the decay phase of the heat storage tendency, the net heat flux is larger than the heat storage tendency, suggesting that a process other than surface fluxes is needed to cool SST, such as horizontal advection or entrainment from below. We have separately calculated SST and net heat flux for the component regions separated by the Meso-American landmass, i.e., the ENP and IAS portions of the WHWP. The calculations show that the heat budget of the IAS is similar to that of the WHWP as a whole, but that in the ENP the development phase as well as the decay phase requires extra cooling. Moreover, XBT data show that the mixed layer depth is shallower during late summer and fall (compared to spring) in the Caribbean region. Thus, the seasonal

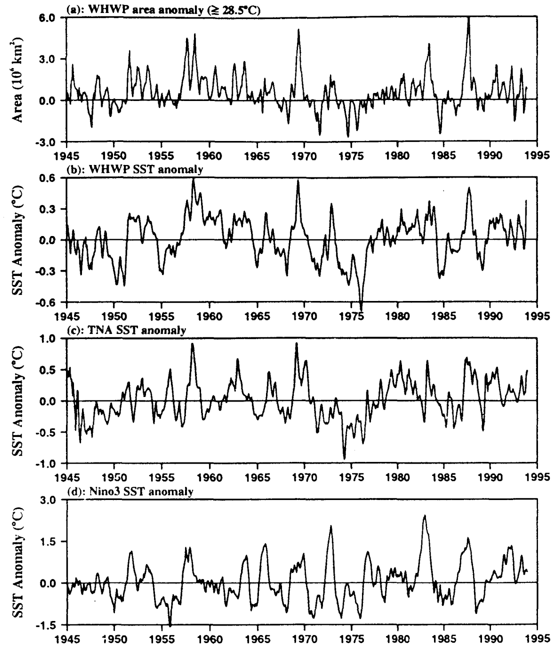

Figure 3. Three-month running means of (a) area anomalies for water warmer than 28.5°C, (b) SST anomalies in the Western Hemisphere warm pool (WHWP) region (7°N-27°N, 110°W-50°W), (c) SST anomalies in the tropical North Atlantic (TNA) region (6°N-22°N, 60°W-15°W); and (d) SST anomalies in the Nino3 region (150°W-90°W, 5°S-5°N).

variations of the mixed layer will make the heat storage tendency curve in Fig. 2a even lower during the decay phase.

The surface net heat flux is composed of shortwave radiation, latent heat flux, longwave radiation, and sensible heat flux. The longwave radiation in the *da Silva et al.* data is defined as net longwave radiation across the sea surface, which is a function of air temperature, SST, cloudiness, and atmospheric vapor pressure. Fig. 2b shows that the shortwave radiation and the latent heat flux are the largest terms whereas the longwave radiation and sensible heat flux have relatively small amplitudes. The shortwave radiation is maximum from April to August, and the latent and sensible heat fluxes have their minimum values around May owing to lower wind speeds associated with the seasonal south-north movement of the ITCZs. This results in the maximum of net heat flux occurring in May. Minimum values of net longwave radiation occur in the boreal summer, associated with greater cloud cover and a corresponding increase in the downward longwave radiation from cloud ceilings. In spite of relatively small amplitudes, both longwave radiation and sensible heat flux play a secondary role by keeping the net heat flux from being larger in the late spring and early summer.

4. A Positive Ocean-Atmosphere Feedback

The WHWP also shows variability on interannual and longer time scales, as evidenced in Fig. 1 (the bottom panel) and Figs. 3a and b for the area and regional intensity indices of SST. The correlation between the area anomalies and the regional SST anomalies is 0.75. Thus, the size of the WHWP is highly related to the intensity of the WHWP warming. The relationship is not perfect, as we would expect of a fixed area in comparison with the complex warm pool evolution. Phenomena important for the climate variability of the intertropical Americas include the Pacific ENSO, its Atlantic extension [*Enfield and Mayer*, 1997] and the tropical Atlantic meridional SST gradient [*Chang et al.*, 1997; *Enfield et al.*, 1999]. Therefore, the WHWP SST anomalies may be related to the Pacific ENSO and the Atlantic meridional gradient mode, the latter being impacted more by the TNA [*Enfield*, 1996]. The TNA and Nino3 SST anomalies are shown in Figs. 3c and d, respectively. The maximum correlation coefficients of the WHWP and TNA SST anomalies and of the WHWP and Nino3 SST anomalies are 0.60 and 0.59 with zero and three month lags, respectively. Maximum correlation between the TNA and Nino3 SST anomalies of 0.45 occurs when the TNA SST anomalies lag the Nino3 SST anomalies by 5 months. The amplitude of the WHWP SST anomalies is smaller than that for either the Nino3 or the TNA. However, as in the western Pacific warm pool [*Wang et al.*, 1999; *Wang*, 2000], the WHWP anomalies occur at high SSTs where even small intensity changes, coupled with the large changes in area, can have a significant impact on the convection development.

A positive ocean-atmosphere feedback operating through latent heat flux has been hypothesized for the variations of TNA SST anomalies [*Chang et al.*, 1997]. This feedback is physically different from the positive feedback of ENSO operating dynamically through momentum flux in that the atmosphere interacts thermodynamically with the ocean through latent heat flux. An increase in the TNA SST anomalies is associated with a decrease in wind speed, which in turn acts to decrease latent heat flux and further increase SST anomalies. We test for a westward extension of this hypothesis by calculating the correlation between the WHWP SST and latent heat anomalies. The WHWP SST anomalies do not significantly correlate with the WHWP latent heat anomalies at any time lags (Fig. 4a). This suggests that the positive feedback operating through latent heat flux is not a mechanism for the WHWP SST anomaly variations.

We also calculate the lagged cross-correlations between the different WHWP variables. Cause-effect relationships and positive ocean-atmosphere feedbacks may be inferred from distribution of lagged cross-correlation. The relation between zonal winds and SST in the equatorial Pacific (associated with ENSO) is a classic example of this. Normally, a positive feedback implies that the ocean and atmosphere mutually interact with each other, yielding a cross-correlation that is more or less symmetrical about zero lag. Fig. 4b shows that the maximum negative correlation between the WHWP

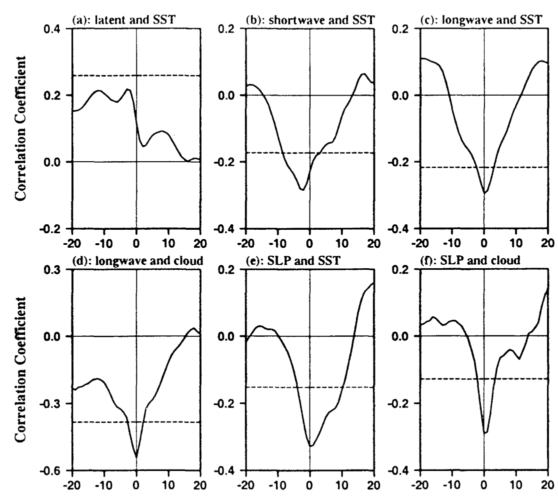

Figure 4. The lagged cross-correlation coefficients between (a) the WHWP latent heat flux and SST anomalies, (b) the WHWP shortwave radiation and SST anomalies, (c) the WHWP longwave radiation and SST anomalies, (d) the WHWP longwave radiation and cloud cover anomalies, (e) the WHWP SLP and SST anomalies, and (f) the WHWP SLP and cloud cover anomalies. The horizontal dashed lines represent 90% significance level. Negative lags imply that the second variable leads the first.

shortwave radiation and SST anomalies occurs when the shortwave radiation anomalies lag the WHWP SST anomalies by 2-3 months. The phase lag implies that shortwave radiation anomalies do not force SST anomalies in the WHWP, rather they are a consequence of the SST anomalies. Warm SST anomalies increase convective cloudiness which reduces the solar radiation into the ocean. The decrease of shortwave radiation anomalies tends to damp the original warm SST anomalies, and thus reduces the persistence of SST anomalies.

In contrast, the cross-correlation between the surface longwave radiation and SST anomalies show a nearly symmetrical distribution about zero lag (Fig. 4c). A negative correlation between the longwave radiation and cloud anomalies also occurs at zero lag (Fig. 4d). Together, Figs. 4c and d suggest a positive feedback between the longwave radiation and SST anomalies associated with cloudiness variations. Thus, an initial warming causes increased convection and cloudiness, the latter reduces longwave radiative heat loss from the sea surface, and the surface temperature increases, reinforcing the original SST anomalies.

Based on these results, we propose the hypothesis that a positive ocean-atmosphere feedback operating through surface longwave radiation and associated cloudiness is responsible for the WHWP anomaly warmings. Given an increase in SST anomalies, the atmosphere shows a decrease in SLP anomalies (Fig. 4e). The decrease of SLP anomalies results in an anomalous increase in deep convective activity and atmospheric cloudiness (Fig. 4f). The increase in convection and cloudiness results in less longwave radiation loss from the surface, which then reinforces the original SST anomalies (Figs. 4c and d). Therefore, it seems that the positive feedback operating through longwave radiation increases the WHWP SST that is damped by shortwave radiation 2-3 months later.

Previous research is consistent with these results. It has been reported that clouds reduce surface longwave radiation by about 16-19 W/m^2 [*Ramanathan et al.*, 1995], suggesting that longwave radiation and associated cloudiness are indeed capable of changing SST. Another study showed that the IAS radiates more (less) energy to space during periods of high (low) SLP [*Knaff*, 1997], a fact that is consistent with the inverse relationship between SST and SLP (Fig. 4e) and with the relationships among longwave radiation, SST, and cloud (Figs. 4c and d). This feedback is also consistent with the results of *Ramanathan and Collins* [1991] who observed a positive feedback between SST and cloud longwave forcing during the Pacific 1986-87 El Niño. They also argued that the positive feedbacks due to the clouds and greenhouse gases in the atmosphere are diminished by the significant decrease in cloud shortwave forcing. *Philander et al.* [1996] proposed a positive ocean-atmosphere feedback operating through shortwave radiation due to low-level stratus clouds over the cold water region of the southeast tropical Pacific. In their feedback, cold SSTs strengthen the atmospheric inversion and hence produce more low-level stratus clouds, which decrease the temperature further by reducing shortwave radiation. However, the WHWP is associated with warm water which produces high-level and convective clouds. These environmental differences seem to favor the clouds participating through longwave radiation in the WHWP region rather than through shortwave radiation.

5. Summary

The WHWP shows a large seasonal cycle and significant interannual and longer time scale variations. Seasonally, the surface net heat flux seems to be primarily responsible for the WHWP warming, although the secondary roles of reduced entrainment cooling, in general, and advection in the ENP region, deserve further attention. Although the SST anomalies are not large compared to the annual cycle or to the Pacific El Niño, the WHWP variability occurs at high temperatures where small changes induce large effects on tropical convection. Moreover, the changes in SST anomalies are highly correlated with large expansions and contractions in the boreal summer area covered by SSTs over 28.5°C. The data suggest that a positive ocean-atmosphere feedback operating through surface longwave radiation and associated cloudiness may be responsible for the WHWP anomaly warming. A better understanding of WHWP SST variations may lead to improved forecasts of hurricanes in the eastern Pacific and Atlantic, and rainfall over its surrounding countries. Finally, the results of this paper are inferred from the *da Silva et al.* data, thus a further study of the WHWP is needed given potential errors of this data set.

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References

Chang, P., L. Ji, and H. Li, A decadal climate variation in the tropical Atlantic Ocean from thermodynamic air-sea interactions, *Nature*, *385*, 516-518, 1997.

da Silva, A. M., C. C.Young, and S. Levitus, *Atlas of surface marine data 1994. Vol. 1. Algorithms and procedures*, NOAA Atlas NESDIS 6, US Dept. of Commence, Washington, DC, 83 pp, 1994.

Enfield, D.B., Relationships of inter-American rainfall to tropical Atlantic and Pacific SST variability, *Geophys. Res. Lett.*, *23*, 3505-3508, 1996.

Enfield, D. B., and D. A. Mayer, Tropical Atlantic sea surface temperature variability and its relation to El Niño-Southern Oscillation, *J. Geophys. Res.*, *102*, 929-945, 1997.

Enfield, D. B., A. M. Mestas-Nunez, D. A. Mayer, and L. Cid-Serrano, How ubiquitous is the dipole relationship in tropical Atlantic sea surface temperature?, *J. Geophys. Res.*, *104*, 7841-7848, 1999.

Graham, N. E., and T. P. Barnett, Sea surface temperature, surface wind divergence, and convection over tropical oceans, *Science*, *238*, 657-659, 1987.

Gray, W. M., Global view of the origions of tropical disturbances and storms, *Mon. Wea. Rev.*, *96*, 669-700, 1968.

Knaff, J. A., Implications of summertime sea level pressure anomalies in the tropical Atlantic region, *J. Climate*, *10*, 789-804, 1997.

Landsea, C. W., A climatology of intense (or major) Atlantic hurricanes, *Mon. Wea. Rev.*, *121*, 1703-1713, 1993.

Magaña, V., J. A. Amador, and S. Medina, The midsummer drought over Mexico and central America, *J. Climate*, *12*, 1577-1588, 1999.

Philander, S. G. H., D. Gu, D. Halpern, G. Lambert, N. C. Lau, and R. C. Pacanowski, Why the ITCZ is mostly north of the equator, *J. Climate*, *9*, 2958-2972, 1996.

Ramanathan, V., and W. Collins, Thermodynamic regulation of ocean warming by cirrus clouds deduced from observations of the 1987 El Niño, *Nature*, *351*, 27-32, 1991.

Ramanathan, V., B. Subasilar, G. J. Zhang, W. Conant, R. D. Cess, J. T. Kiejl, H. Grassl, L. Shi, Warm pool heat budget and shorwave cloud forcing: A missing physics, *Science*, *267*, 499-503, 1995.

Wang, C., R. H. Weisberg, and J. I. Virmani, Western Pacific interannual variability associated with the El Niño-Southern Oscillation, *J. Geophys. Res.*, *104*, 5131-5149, 1999.

Wang, C., On the atmospheric responses to tropical Pacific heating during the mature phase of El Niño, *J. Atmos. Sci.*, *57*, 3767-3781, 2000.

Weisberg, R. H., On the evolution of SST over the PACS region, *Abstracts of 76th AMS Annual Meeting*, Atlanta, Georgia, Amer. Meteor. Soc., p378, 1996.

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