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On the Seasonal Sea Surface Temperature Variations in the Tropical Pacific

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With 9 Figures

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Summary

In this paper, we investigated physical processes that control the seasonal variations of sea surface temperature in the tropical Pacific, using an intermediate ocean model. It is found that the westward propagation of sea surface temperature along the equator is attributed to dynamic response of the ocean to the wind (that consists of 3-dimensional temperature advection), whereas the northward propagation of sea surface temperature in the eastern Pacific results from the thermodynamic response of the ocean to the surface heat flux, primarily shortwave radiation that includes the effect of low-level stratus clouds. The remote response of the eastern Pacific sea surface temperature to seasonally varying wind in the western Pacific is of secondary importance, compared to the local wind forcing. The results suggest that the mechanism that controls the seasonal cycle of sea surface temperature is different from that associated with El Nino-Southern Oscillation.

1. Introduction

Annual cycle in sea surface temperature (SST) shows clear westward propagation along the equator (Horel, 1982). Figure 1a shows this feature. Note that a maximum SST anomaly occurs in March around 100° W and moves westward and reaches to the date line in June. While this positive SST anomaly continues to propagate westward, a cold SST anomaly appears in the eastern equatorial Pacific. Seasonal change in the western Pacific has a clear semi-annual variation, closely following the solar radiation, whereas the eastern equatorial Pacific is dominated by an annual harmonic.

Two processes may be responsible for the westward propagation of SST. One is latent heat (or evaporation) at the ocean surface. Another is seasonal upwelling that carries colder water below the surface. In response to a positive SST anomaly in the eastern equatorial Pacific, a westerly wind establishes just west of the SST anomaly which, on one hand, suppresses annual mean upwelling (due to mean easterly wind along the equator) and, on the other hand, reduces the latent heat from ocean to atmosphere due to decrease in wind speed. As a result, the SST anomaly propagates westward. So far it is not clear which process is dominant.

In addition to the westward propagation along the equator, there seems to be northward propagation in the eastern Pacific. Figure 1b shows a latitude-time section. It is found that a positive SST anomaly occurs at 30° S in February and moves northwest to the equator in April. Similarly, a negative SST anomaly appears at 30° S in August and move to the equator in October. As we know, the sun moves back and forth and crosses the equator twice a year. Why does the sea surface temperature evolve with time so differently? What is the effect of low-level stratus clouds on the SST evolution?

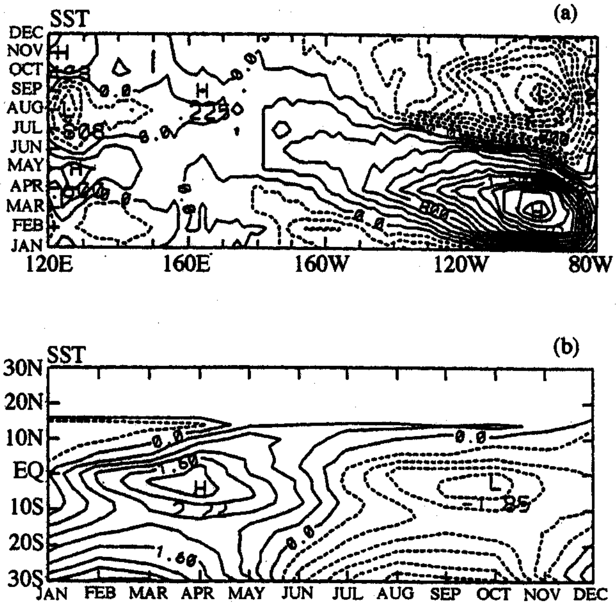

Fig. 1. (a) The time-longitude seaction of observed SST anomalies along the equator and (b) the time-latitude section of the SST anomalies along 100–110° W. The data is from Sadler et al. (1987). Contour intervals are 0.2 °C and 0.4 °C, respectively

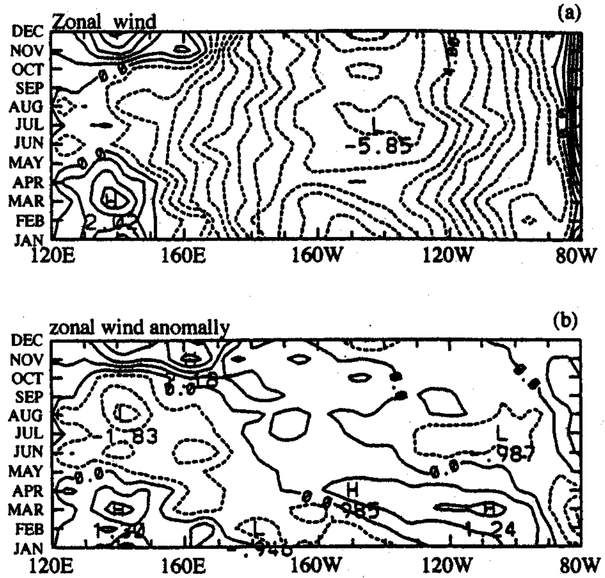

Fig. 2. Climatological monthly mean surface zonal wind (a) and the corresponding monthly mean anomaly (b) along the equator. The data is from Sadler et al. (1987). Contour interval is 0.5 m/s

The wind in the western Pacific may have a remote effect on the SST change in the eastern Pacific. Measurements indicated that there is eastward propagation in sea level signals with a particular phase speed of 2–3 m/s during a strong westerly burst in April 1980 in the western Pacific (Knox and Halpern, 1982). Lukas et al. (1984) analyzed the sea level records from a number of stations during the onset of the 1982–1983 ENSO and found that there are eastward-propagating signals that resemble the first and second-mode Kelvin waves. Seasonally, climatological monthly mean wind also shows a large variation over the western Pacific. For example, both total and anomalous winds (as shown in Fig. 2) are westerlies in the northern winter between 120° E and 170° E. A maximum westerly anomaly (2.2 m/s) occurs in November at 160° E whereas a total westerly wind (2.5 m/s) appears in December around 140° E. Lukas (1981) speculated that the seasonal change of sign of pressure gradient associated with equatorial undercurrent at 86° W–96° W during early spring is caused by the remote wind forcing in the west side of the ocean through the propagation of an oceanic Kelvin pulse. Wang (1994) suggested that the change in SST off the coast of Ecuador (80–86° W, 2° N–2° S) may be a remote response to the annual wind variation in the western Pacific, since it leads by about two months. Interannually, the change in SST results from east-west movement of warm surface water and remote forcing of the western Pacific wind. Does seasonal variation of sea surface temperature also depend on remote forcing in the western Pacific? Or does the annual cycle in the eastern Pacific result from local air-sea interaction?

This paper is an investigation of the physical mechanisms of seasonal cycle of sea surface temperature in the tropical Pacific by means of an intermediate ocean model. The specific questions we address here are: What is relative importance of dynamic and thermodynamic processes to cause the seasonal SST change? Does the annual variation in the eastern Pacific result from local air-sea interaction or result from remote response of the western Pacific wind? In Section 2, the ocean model is briefly described. To identify the relative importance of the dynamic and thermodynamic processes, four experiments are designed in terms of annual mean or annual cycle of surface wind and heat flux forcing. The results are presented in Section 3. In Section 4, we investigate the effects of remote and local wind forcing on the annual cycle of SST in the equatorial eastern Pacific. Finally, a summary is given in Section 5.

2. The Model

The ocean model used in this study is a Cane-Zebiak type model (Cane, 1979; Zebiak and Cane, 1987; Seager et al., 1988; Chang, 1994). It describes the linear dynamics of an homogeneous upper layer with a varying thermocline that overlies a motionless abyssal layer with a constant reference temperature. A constant bouyancy is assumed in the upper ocean ($b = 5.6\,\mathrm{cm/s^2}$) that corresponds to a Kelvin wave speed of 290 cm/s for a mean thermocline depth of 150 m. The mean upper-ocean currents ($\mathbf{u}$) and thermocline depth (h) equations can be written as:

$$\partial \mathbf{u}/\partial t + fk \times \mathbf{u} + \nabla bh = \tau/H + \mu\nabla^2\mathbf{u} - r\mathbf{u} \qquad (2.1)$$

$$\partial h/\partial t + H\nabla(\mathbf{u}) = -rh \qquad (2.2)$$

where τ, μ, r, and H represent the surface wind stress, diffusion coefficient, the Rayleigh friction coefficient, and the mean depth of thermocline, respectively. A constant surface (Ekman) layer is added into the upper ocean to capture the intensity of wind-driven surface currents and to predict the sea surface temperature. The equation for shear currents between the surface layer and the layer below may be written as

$$r_s u_s - f v_s = \tau^{(x)}/\rho_0 H_1 \qquad (2.3a)$$

$$r_s v_s + f u_s = \tau^{(y)}/\rho_0 H_1 \qquad (2.3b)$$

where $\mathbf{u}_s = \mathbf{u}_1 - \mathbf{u}_2$, $\tau^{(x)}$ and $\tau^{(y)}$ are wind stress in zonal and meridional directions, H_1 represents the depth of the surface layer, and subscripts 1 and 2 refer to the surface layer and underlying layer, respectively.

The vertical entrainment velocity at the base of surface layer is determined by the Ekman divergence of surface-layer currents, namely,

$$w_e = H_1(\partial u_1/\partial x + \partial v_1/\partial y) \qquad (2.4)$$

The sea surface temperature is determined by 3-dimensional temperature advection and surface heat flux, namely

$$\partial T/\partial t + \mathbf{u}_1 \cdot \nabla T = Q/\rho_0 C_w H_1 + \kappa\nabla^2 T - w_e H(w_e)(T - T_e)/H_1 \qquad (2.5)$$

where T_e represents subsurface temperature that is parameterized from thermocline depth variations, following Chang (1994), ρ_0 and C_w are the density and specific heat of water, Q is diabatic heating that consists of solar radiation, longwave radiation, latent heat, and sensible heat, and $H(x)$ is a heavyside step function which equals to 1 when $x > 0$ and equals to 0 when $x < 0$.

The model covers the tropical Pacific basin from 120° E to 80° W and from 30° S to 30° N. A realistic coast geometries along the North America and Australia are specified. The model has a spatial resolution of 2° longitude by 1° latitude with a time step of 3 hours. An absolute potential enstrophy conserving scheme is applied and the model is formulated on a staggered C grid in a spherical coordinate. No-flux boundary conditions for temperature are applied at the ocean boundary. At northern and southern boundaries, a large Rayleigh friction and Newtonian damping, with a time scale of a half day, that rapidly decays toward the ocean interior is introduced to eliminate coastal Kelvin waves. A constant heat flux correction, with a Newtonian damping time scale of 5 months toward observed annual mean SST, is applied. The climatological monthly mean surface winds derived from Sadler et al. (1987) and heat fluxes from Esbensen and Kushnir (1981) are used.

The model contains a number of parameters that have the following values. The density of ocean water, ρ_0, equals to 1 g/cm³, the air density at surface, ρ_a, is $1.2 \times 10^{-3}\,\mathrm{g/cm^3}$, the drag coefficient, C_D, is set to be 1.5×10^{-3}, the heat capacity of water, C_w, is $4.2 \times 10^3\,\mathrm{J\,kg^{-1}\,K^{-1}}$, the Rayleigh friction coefficient, r, is set to be 1/150 day^{-1}, the diffusion coefficients in momentum and thermodynamic equations, μ and κ, are set to be $5 \times 10^7\,\mathrm{cm^2/s}$, the mean depth of thermocline, H, is 150 m, the depth of the constant surface layer, H_1, is 50 m, and the friction coefficient in the shear Eq. (2.3), r_s, is set to be 0.5 day^{-1}.

3. Surface Wind vs. Heat Flux Forcing

The change in SST depends primarily on dynamic response of the ocean to the wind that includes horizontal and vertical temperature advection and thermodynamic response to the heat flux. To identify the relative importance of the two processes, four experiments are designed. In the first experiment, the model is forced by observed annual mean wind and heat flux. In the second

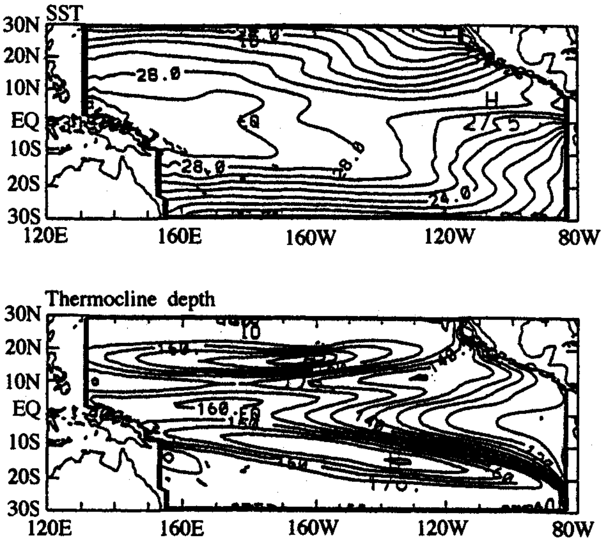

Fig. 3. The model simulated annual mean SST (upper panel) and thermocline depth (in units of m) (lower panel) in response to observed annual mean wind and heat flux forcing. The contour intervals are 1 °C and 5 m

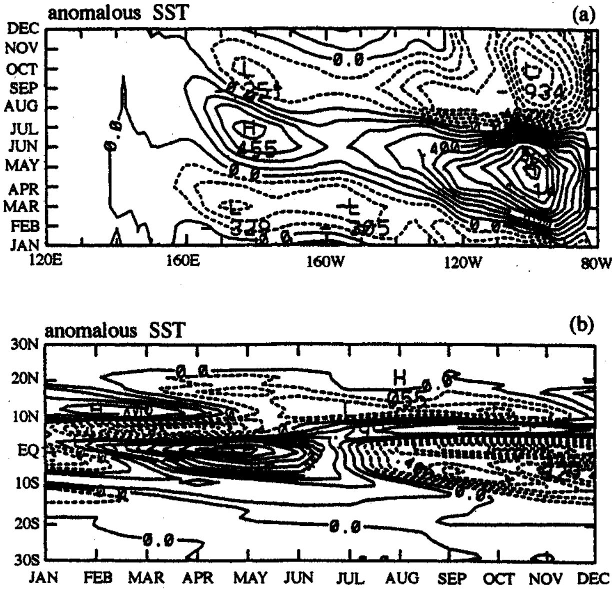

Fig. 4. The time-longitude section along the equator (upper panel) and time-latitude section along 100–110° W (lower panel) of the SST amomalies in response to seasonally varying wind forcing. In this case, surface heat fluxes are specified from climatological annual mean fields. The contour interval is 0.1 °C

experiment, the model is forced by annual cycle wind and annual mean heat flux. In the third experiment, seasonally varying heat flux is specified but the wind keeps its annual mean value. In the fourth experiment, both annual cycle wind and heat flux are specified.

Figure 3 shows the model simulation of sea surface temperature and thermocline depth in the presence of annual mean wind and heat flux forcing (experiment 1). Note that this model is capable of simulating some essential features of tropical Pacific SST. For example, both the western Pacific warm pool and the eastern Pacific cold tongue are well simulated. In most region, the error in the SST field is less than 1 °C. Maximum errors occur in the equatorial eastern Pacific (1.3 °C) and along the South America coasts (2.5 °C). Generally, the thermocline is shallow in the eastern equatorial Pacific and South America coast, and becomes deep in the western Pacific and subtropical regions. The success in reproducing annual mean conditions add confidence to further investigate the seasonal cycle.

Seasonal change in surface wind stress (experiment 2) essentially captures the westward propagation of SST along the equator. As shown in time-longitude section (Fig. 4a), there is clear westward propagation of SST. In pure response to dynamic forcing of the wind, the SST in the eastern Pacific yields a dominant annual cycle. The amplitude of the SST anomaly generated by the dynamic forcing is close to that in presence of pure thermodynamic forcing (experiment 3, shown in Fig. 5a), suggesting that both dynamic response of the ocean to the wind and the thermodynamic response to heat flux are equally important in promoting annual cycle in the equatorial eastern Pacific. The seasonally varying heat flux (that includes latent heat flux), on the other hand, does not have significant contribution on the westward propagation along the equator (Fig. 5a).

It is intersting to note that seasonal variations in heat flux, not in wind stress, contribute to the northward propagation of SST in the eastern Pacific (see Figs. 5b and 4b). Further study reveals that the northward propagation depends primarily on shortwave radiation that includes the effect of low-level stratus clouds. In the eastern equatorial Pacific are dominated low-level stratus clouds. These clouds are generally negatively correlated with sea surface temperature. This is because colder the SST is, larger atmospheric static stability, stronger the atmospheric inversion, more the low

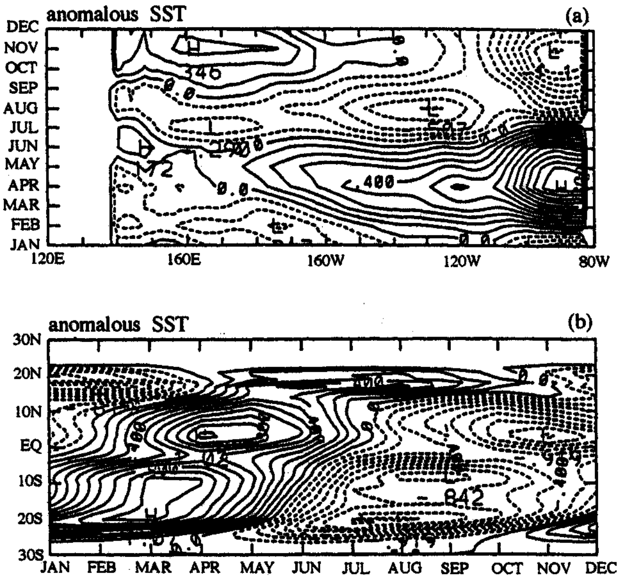

Fig. 5. The time-longitude section along the equator (upper panel) and time-latitude section along 100–110° W (lower panel) of the SST anomalies in response to seasonally varying heat flux forcing. In this case, surface wind stresses are specified from climatological annual mean fields. The contour interval is 0.1 °C

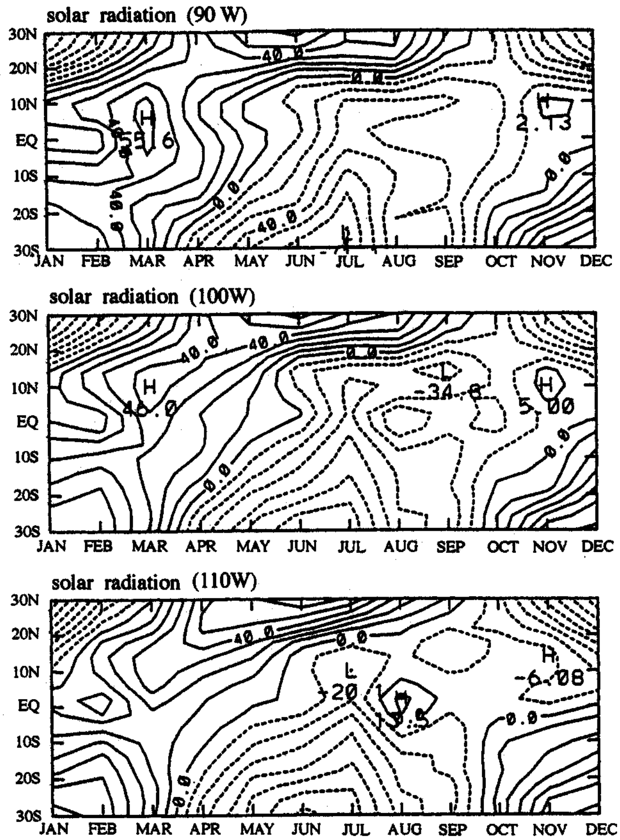

Fig. 6. The time-latitude section of downward solar radiation at the ocean surface (Esbensen and Kushnir, 1981) along 90° W, 100° W, and 110° W, respectively. The effect of clouds has been considered. The contour interal is 10 W/m^2

level stratus clouds. After the northern summer, the sun starts to move to the south. Due to the intensification of the equatorial cold tongue in September, the clouds increase dramatically, which effectively reduces solar radiation and prevents the increase of SST. As a result, no significant southward movement is observed in the SST field (Fig. 1b). Figure 6 exhibits the latitude-time section of available solar radiation (that includes the cloud effect) in the eastern Pacific. As discussed above, due to low-level stratus cloud feedback, only the northward movement of shortwave radiation is observed.

In response to both seasonally varying wind and heat flux forcing (experiment 4), the model captures both the westward propagation and the northward propagation in the SST field. The resultant SST is approximately a linear combination of the previous two experiments (experiments 2 and 3).

To sum up, both dynamic and thermodynamic responses are important to promote annual variation in the eastern Pacific. Seasonal upwelling plays a crucial role in causing the westward propagation along the equator whereas the shortwave radiation modified by low-level stratus clouds is responsible for the northward movement in the eastern Pacific.

4. Remote vs. Local Wind Forcing

In this section, we investigate the response of the eastern Pacific SST to remote and local wind forcing. Two experiments are designed. In the first experiment, seasonally varying wind is specified in the western Pacific (west of the date line), while annual mean wind is specified east of the date line. In the second experiment, we specify the annual cycle wind in the eastern Pacific (east of 180° E) but keep the annual mean wind west of 180° E. In both cases, climatological annual mean heat fluxes are used.

In response to remote forcing of the wind in the western Pacific, a moderate SST anomaly occurs

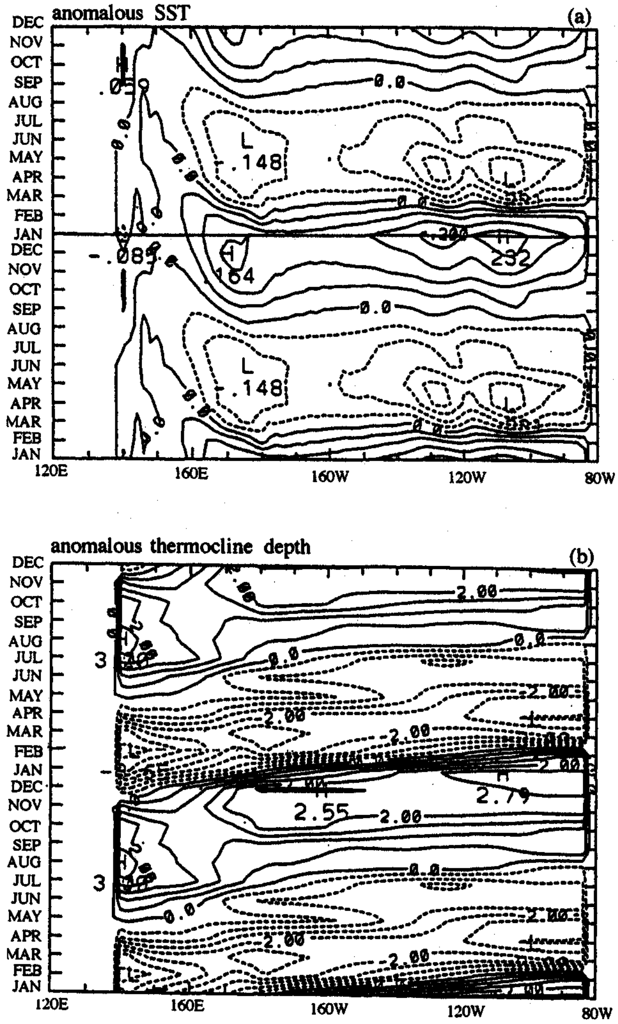

Fig. 7. The time-longitude section of the simulated SST anomaly (upper panel) and thermocline depth anomaly (in units of m) (lower panel) along the equator in response to seasonally varying wind forcing west of the date line. The contour intervals are 0.05 °C and 0.5 m

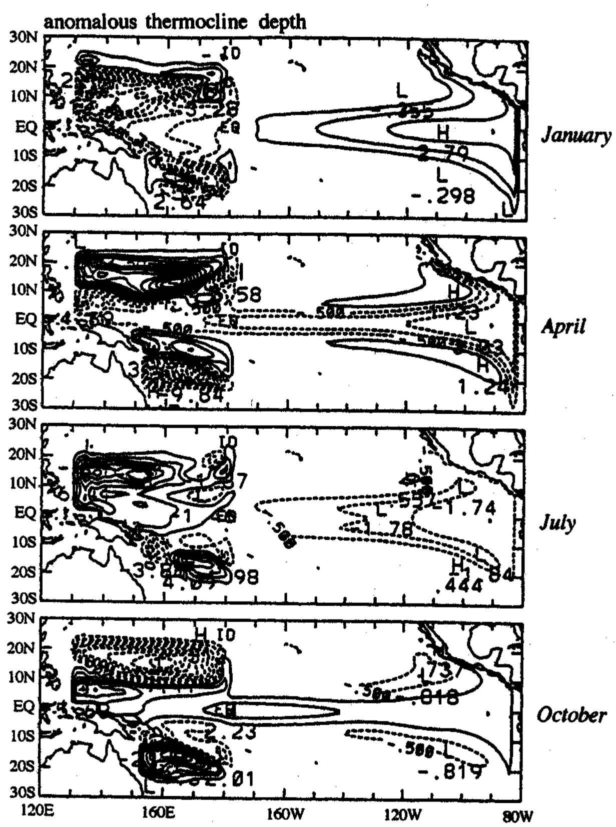

Fig. 8. Horizontal structure of the simulated thermocline depth anomaly in January, April, July, and October, respectively, in response to seasonally varying wind forcing west of the date line. The contour interval is 1 m

in the eastern Pacific. Since there is no seasonal wind forcing east of the date line, the SST anomaly in the eastern Pacific is purely due to ocean wave dynamics that carry information from the west of the ocean and cause the thermocline variation in the eastern Pacific. It is found from Fig. 7a that a positive SST anomaly (0.23 °C) occurs in January around 110° W. A smaller SST anomaly (0.16 °C) is found in December at 170° E. The positive SST anomaly seems to propagate eastward with a phase speed of 75° longitude per month (about 3 m/s) that is close to oceanic Kelvin wave speed. The eastward propagation of the SST coincides well with the thermocline depth variation (Fig. 7b). The positive (negative) thermocline depth anomalies correspond to the warm (cold) SST anomalies.

To further identify the horizontal structure of the eastward-propagating Kelvin waves, we illustrate the model simulated anomalous thermocline depth fields (Fig. 8). These maps clearly show the structure of equatorial Kelvin waves, with a maximum amplitude right at the equator. West of the date line, the thermocline depth anomaly shows complicated patterns, generally asymmetric to the equator, in response to local wind forcing. East of the date line, where there is no anomalous wind forcing, the thermocline anomaly is approximately symmetric about the equator, resembling the Kelvin wave structure. The anomaly structure extends poleward along the coasts of the North and South Americas.

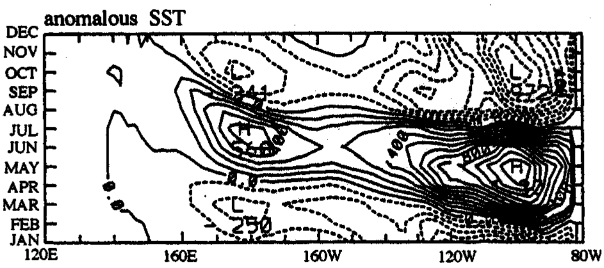

Fig. 9. The time-longitude section of the simulated SST anomaly along the equator in response to seasonally varying wind forcing east of the date line. The contour interval is 0.1 °C

An interest experiment in contrast to the above case is to examine the response of the ocean to seasonally varying winds east of the date line with a specified annual mean wind west of 180° E. Different from the eastward propagation of SST in the aforementioned case, the SST anomaly shows clear westward propagation due to local wind forcing (Fig. 9). The amplitude of SST anomaly due to local wind forcing is much larger than that in presence of remote forcing, suggesting that the seasonal variation of equatorial cold tongue is primarily controlled by local forcing, rather than the remote wind forcing. It is further confirmed by designing an additional experiment in which seasonally varying wind is only confined to east of 120° W. Compared this with the previous case, the amplitudes of the SST anomaly are very close.

5. Summary

In this study, we investigate the physical mechanisms that determine the annual cycle of SST in the tropical Pacific. A reduced-gravity ocean model is used to study the dynamic response of the tropical Pacific ocean to surface wind and the thermodynamic response to the surface heat flux. The model is a Cane-Zebiak type model (Cane, 1979; Zebiak and Cane, 1987; Seager et al., 1988; Chang, 1994) that contains a reduced-gravity upper ocean with varying thermocline depth and a constant mixed layer. Sea surface temperature is primarily determined by 3-dimensional temperature advection and diabatic heating that consists of shortwave and longwave radiations and latent and sensible heat fluxes. Climatological monthly mean surface winds (Sadler et al., 1987) and heat fluxes (Esbensen and Kushnir, 1981) are used as external forcing.

By isolating seasonally varying wind and/or heat flux forcing, we found that the westward propagation of SST along the equator results from the wind-induced dynamic processes, primarily seasonal upwelling, and the northward propagation of the SST in the eastern Pacific results from the thermodynamic response of the ocean to the heat flux, primarily shortwave radiation modified by low-level stratus clouds. It is found that both dynamic and thermodynamic processes are important in contributing the observed strength of annual cycle of SST in the equatorial eastern Pacific.

Interannually, the remote forcing of the western Pacific wind play an important role. On seasonal time scale, it is of secondary importance in changing SST. Our results indicate that the annual cycle of SST in the eastern Pacific depends primarily on local air-sea interactions.

From a meteorological point of view, the westward propagation of atmospheric quantities such as wind, sea-level pressure, and precipitation along the equator is attributed to SST forcing, and from an oceanographic point of view, the westward propagation of SST results from zonal wind forcing. The circular argument suggests that coupled ocean-atmosphere interaction is of central importance. In this study, we investigated the response of tropical Pacific ocean to described wind and heat flux forcing. Future work will focus on a coupled air-sea interaction system and examine how the coupled system responds to solar radiation forcing.

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