The Role of the Dynamic Ocean–Atmosphere Interactions in the Tropical Seasonal Cycle

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ABSTRACT

The role of dynamic ocean–atmosphere interactions on the evolution of the tropical seasonal cycle is explored using a simple coupled model. It is shown that the seasonal cycle in the coupled system can be divided into two parts—a forced part that is a direct response to seasonal changes in the solar radiation and a coupled part that involves dynamic feedbacks between the oceans and atmosphere. The latter part contributes significantly to the pronounced annual cycle in the eastern equatorial Pacific, but is less influential in the western Pacific, owing to the different climate mean conditions. The study further suggests that the ocean–atmosphere interactions in the meridional and zonal direction play different roles in the evolution of the tropical annual cycle. The former is crucial to the development of the strong annual cycle in the near-coastal zone of the eastern Pacific (eastward of 100°W), whereas the latter is instrumental in the westward expansion of the annual cycle along the equator. The results of this study suggest that many important features of the tropical seasonal cycle can be modeled with a relatively simple coupled model, provided that the climate mean conditions are correctly established.

1. Introduction

In the Tropics, the seasonal cycles over the eastern Pacific and Atlantic Oceans are strikingly different from those over the western Pacific and Indian Oceans (Horel 1982; Mitchell and Wallace 1992; Nigam and Chao 1995). The maximum seasonal fluctuations of sea surface temperature in the eastern Pacific and Atlantic occur over the cold tongue regions just south of the equator. The variabilities in these regions are dominated by a strong annual cycle, which is asymmetric about the equator despite semiannual solar forcing near the equator. The amplitudes of the fluctuations are comparable to those in the subtropical oceans (larger than 2°C; Fig. 1a). Interestingly, the corresponding variabilities in the surface winds exhibit local maxima on the opposite side of the equator (Fig. 1b). Studies show that the annual variations of the sea surface temperature in these regions are highly correlated with fluctuations in the local winds, both the zonal component and the meridional component (Mitchell and Wallace 1992; Wang 1994; Nigam and Chao 1995), but are less correlated with the fluctuations in the thermocline (Chang 1993; Chang and Philander 1994; and Nigam and Chao 1995). In contrast, the seasonal cycles in the tropical western Pacific and Indian Oceans are more symmetric about the equator and much weaker, but the semiannual component appears to be more pronounced near the equator than in the eastern Pacific and Atlantic Oceans (Figs. 1c,d).

The different character of the seasonal cycle in the Tropics can be, to a large extent, attributed to different climate-mean conditions that result from complex interactions between the atmosphere, ocean, and land surfaces. In the eastern Pacific and Atlantic Oceans, because of prevailing easterly trade winds in the Tropics, the thermocline is usually much shallower than in the western Pacific and Indian Oceans. The ocean–atmosphere feedbacks are most effective in these regions because the changes in the winds can readily affect SST and vice versa. As a result, the climate-mean conditions are asymmetric relative to the equator. That is, the regions of maximum SST, cloudiness, and rainfall are north of the equator. Chang and Philander (1994, hereafter referred as CP) demonstrated that the ocean–atmosphere feedbacks involving the meridional SST gradient and meridional wind can convert an equatorially symmetric mean state into an asymmetric state. Xie and Philander (1994) show that the feedbacks between the oceanic mixed layer and evaporation in the atmosphere also contribute to the maintenance of the asymmetry of the climate-mean state in these regions. Philander et al. (1996) further pointed out, in a recent coupled general circulation model study, that low-level stratus clouds over cold waters play a crucial role in maintaining the asymmetries.

This study focuses on the issues concerning the effects of various climate-mean conditions on the tropical seasonal cycle. In previous studies of the tropical seasonal cycle, CP suggested that the dynamic ocean–atmosphere interactions not only contribute significantly to the asymmetric climate-mean states the eastern Pa-
Fig. 1. Observed seasonal variations of the SST and the surface meridional wind as a function of latitude and time. Both the temperature and wind are zonally averaged between 110°W and 90°W in (a) and (b) and between 160°E and 180°E in (c) and (d). Contour interval is 0.5°C for the SST and 0.5 m s⁻¹ for the wind.

cific and Atlantic, but also play a vital role in the development of the pronounced annual cycles in these regions. In particular, the interaction between the meridional wind and SST may be instrumental in rapidly reestablishing the cold tongues in the eastern Pacific and Atlantic Oceans during the Northern Hemisphere summer, whereas the interaction between the zonal wind and SST may contribute to the westward propa-
gating feature in the annual cycle. Giese and Carton (1994) found, by experimenting with a coupled general circulation model using different solar forcings, that when the period of solar forcing was lengthened, the near-equatorial SST response can change from an annual cycle to a semiannual cycle in the eastern Pacific Ocean. Their results imply that the intrinsic oceanic timescale, which depends on the oceanic mean condition, is important in determining the nature of the seasonal cycle in the Tropics. In this study, we attempt to show, using a simple coupled model, that the seasonal cycle in the Tropics can be divided into a forced part and a coupled part. In the western Pacific, where the thermocline is deep and thermal gradients in both vertical and horizontal directions are weak, the seasonal cycle is essentially governed by the forced solution. However, in the eastern Pacific, where thermocline is shallow and the thermal gradients are intense, the coupled model considerably modifies the direct forced response near the equator, producing a pronounced annual cycle. The study further reveals the different roles of ocean–atmosphere interactions in the meridional and zonal directions in the evolution of the tropical annual cycle.

The arrangement of the paper is the following: Section 2 gives a description of a simple coupled model and analytical solutions derived from the model, section 3 presents the results of numerical experiments with the simple model, and section 4 discusses the major findings and other neglected factors that could influence the seasonal cycles.

2. A simple coupled ocean–atmosphere model

a. Model description

Following CP, we choose a simple paired ocean–atmosphere model, which has the Lindzen–Nigam model as the atmospheric component (Lindzen and Nigam 1987) and the Cane–Zebiak model as the oceanic component (Zebiak and Cane 1987). The Lindzen–Nigam model (hereafter the LN model) can be written as follows:

\[ E\tau_x - f \tau_y = -a \left( \frac{\partial \phi}{\partial x} - A \frac{\partial T}{\partial x} \right) \] (1)

\[ E\tau_x + f \tau_y = -a \left( \frac{\partial \phi}{\partial y} - A \frac{\partial T}{\partial y} \right) \] (2)

\[ \phi + B \left( \frac{\partial \tau_x}{\partial x} + \frac{\partial \tau_y}{\partial y} \right) = 0, \] (3)

where \( \tau_x \) and \( \tau_y \) are the zonal and meridional components of the wind stress, \( \phi \) is the geopotential at the top of the boundary layer, \( T \) is the sea surface temperature; \( E \) is a mechanical damping due to vertical diffusion of momentum and surface drag; \( A = gH_0/T_0 \) and \( B = gH_0/\mu a \) measure the strength of the pressure force induced by the SST gradients and the strength of the ”back pressure” effect; \( H_0 \) is the depth of the boundary layer; \( T_0 \) is a reference temperature; \( \mu \) is an inverse relaxation time for the adjustment of the boundary layer height; and \( \alpha \) converts the wind speeds into surface wind stresses.

The oceanic component is an extension of the conventional 11/2-layer reduced-gravity system that includes the physics of the surface mixed layer and allows prediction of the sea surface temperature. Following Cane (1979), the surface mixed layer model has a constant depth and the entrainment velocity \( w \) is determined as the divergence of the surface-layer flow. It has been shown by CP that for the coupled modes of relevance to the seasonal cycle, one can neglect the horizontal advection of heat due to the geostrophic currents \( u \) and \( v \). Therefore, the linearized SST equation in the Cane–Zebiak model can be simplified as follows:

\[ T_x + \bar{u}_x T_x + \bar{v}_y T_y + \frac{\bar{w}}{H_1} (T - yh) + u_1 \tilde{T}_x \]

\[ + v_1 \tilde{T}_y + w \tilde{T}_z + \epsilon T - \kappa \nabla^2 T = \frac{Q}{\rho_0 C_p H_1} \] (4)

\[ \begin{bmatrix} u_1 \\ v_1 \end{bmatrix} = \frac{H_2}{H} \begin{bmatrix} u_e \\ v_e \end{bmatrix}, \quad w = \frac{H_2}{H} \nabla \cdot \mathbf{v}_e \] (5)

\[ \begin{bmatrix} u_e \\ v_e \end{bmatrix} = \frac{1}{\Delta_0} \begin{bmatrix} r_x \tau_x + f \tau_y \\ r_x \tau_y - f \tau_x \end{bmatrix}, \quad \Delta_0 = H_1 (r_x^2 + f^2), \] (6)

where \( H_1 \) and \( H \) are the depth of the mixed layer and the mean depth of the thermocline; \( H_2 \) is defined as the difference between \( H \) and \( H_1 \); \( \bar{u}_x, \bar{v}_y \) and \( u_1, v_1 \) represent the zonal and meridional component of the mean and anomalous surface currents; \( \bar{w} \) and \( w \) denote the mean vertical upwelling and the anomalous upwelling due to the surface Ekman divergence; \( \gamma \) is a parameter that relates the entrained subsurface water to the thermocline depth; \( r_x \) the linear damping coefficient in the mixed layer; \( \epsilon \) and \( \kappa \) are the linear damping and diffusion coefficients in the temperature equation; and \( Q \) is the surface heat flux forcing. A list of model parameters is provided in Table 1.

Chang and Philander have shown that subsurface oceanic dynamics plays a minor role in the seasonal variation of the SST in the cold tongue region. This finding was supported by the recent observational study of Nigam and Chao (1995). Therefore, the influence of \( h \) on SST is ignored by setting \( \gamma \) to zero in (4). In the following analysis, we further simplify the problem by neglecting zonal variations so that analytical solutions to the governing equations (1)–(4) can be sought.

b. A general procedure to solve the simple coupled model

The seasonal variations of SST are caused by changes in both surface heat fluxes and surface winds.
Since the problem at hand is linear, we can divide the total SST response $T$ into two parts: a forced part $T_f$, which is a direct response of SST to changes in solar radiation, and a coupled part $T_c$, which results from dynamic interactions between the ocean and atmosphere. The forced part $T_f$ is governed by the following equation:

\[
\left( \frac{\partial}{\partial t} + F \right) T_f = \frac{Q}{\rho_0 c_p H_1}, \tag{7}
\]

where

\[
F = \bar{v}_1 \frac{\partial}{\partial y} \left( \frac{\bar{w}}{H_1} + \epsilon \right) - \kappa \nabla^2
\]

describes mean advection and diffusion processes and is a function of latitude $y$. The corresponding meridional wind stress $\tau_f^y$ is determined by the LN model from $T_f$; that is,

\[
D(\tau^y) = \frac{\partial T_f}{\partial y}, \tag{8}
\]

where

\[
D(\tau) = -\frac{B}{A} \frac{\partial^2 \tau}{\partial y^2} + \frac{\Delta \tau}{H_1 E \alpha A}, \quad \Delta = H_1 (r_s E - f^2),
\]

\[
\Delta_u = H_s (E^2 + f^2).
\]

Once the forced part is determined, a single equation for the meridional wind stress $\tau_c^y$ of the coupled mode can be derived:

\[
\left( \frac{\partial}{\partial t} + F \right) D(\tau_c^y) + C(\tau_c^y) = -C(\tau_f^y), \tag{9}
\]

where

\[
C(\cdot) = \frac{H_2}{HH_1 E} \frac{\partial}{\partial y} \left( \frac{\Delta}{\Delta_0} \bar{T}_f \right) + \frac{H_2}{HE} \frac{\partial}{\partial y} \left[ \bar{T}_c \frac{\partial}{\partial y} \left( \frac{\Delta_u}{\Delta_0} \right) \right]
\]

describes the meridional advection by anomalous surface currents and the vertical entrainment induced by changes in surface winds. Generally, $C(\cdot)$ is a function of latitude $y$. The proper boundary conditions for (9) are

\[
\frac{d\tau_c^y}{dy} = 0 \quad \text{as} \quad y \to \pm \infty. \tag{10}
\]

The SST response $T_c$ of the coupled mode can be found by solving Eq. (8) provided that $\tau_c^y$ is determined. The complete SST response $T$ is then given by the superposition of $T_f$ and $T_c$.

In the absence of forcing $Q$, $\tau_c^y$ and $T_c$ are both zero, and (9) reduces to an eigenvalue problem by assuming that the solutions are of the form $\tau_c^y = \tau(y) \exp(\sigma t)$. Chang and Philander discussed the properties of the coupled modes in great detail. They showed that some of the coupled modes can become unstable in the eastern equatorial Pacific and Atlantic Oceans. These unstable modes are argued to play a vital role in setting up the asymmetric climate-mean states in the regions. In this study, we assume that climate-mean states are given and we are interested in the questions of how the mean states can affect the coupled modes, how these modes are excited by seasonal changes in the solar radiation, and how they influence the evolution of the seasonal cycle. Therefore, we shall assume, in the following analyses, that all coupled modes are either neutrally stable or damped. Before proceeding to discuss the effects of coupled modes, it is useful to first understand the SST response in the absence of these modes.

c. Forced SST response to changes in the solar radiation

The forced solution $T_f$, which represents the direct SST response to changes in the solar radiation in the absence of the coupled modes, can be obtained by the method of separation of variables. For simplicity, considering a uniform symmetric mean state with $\bar{v} = \bar{T}_f = 0$, using boundary conditions $\partial T_f/\partial y = 0$ at $y = \pm L$ and initial condition $T_f = T_{f0}$ at $t = 0$, $T_f$ can be written as

\[
T_f = T_{f0} + \sum_{n=0}^{\infty} A_n \cos \frac{n\pi}{2L} (y - L), \tag{11}
\]

where

\[
A_n = e^{-\sigma_n^j} \int_0^L e^{\sigma_n^j s} F_f(s) ds, \quad \sigma_n^j = \frac{\bar{w}}{H_1} + \epsilon + \kappa \left( \frac{n\pi}{2L} \right)^2
\]
and
\[ F_j = \frac{1}{L} \int_{-L}^{L} \cos \frac{n\pi}{2L} (y - L) \frac{Q}{\rho_0 c_p H_1} \, dy. \]

In the absence of damping and diffusion, it is readily shown that the SST response is 90° out of phase with the forcing. As damping $\bar{w}/H_1 + \epsilon$ and diffusion $\kappa$ increase, the phase lag increases and the amplitude of $T_j$ decreases.

The seasonal variation of the actual solar radiation consists of an annual component and a semiannual component. In Fig. 2 we plot the seasonal variation of the solar radiation $Q$ (Fig. 2a) along with its annual and semiannual components (Figs. 2b, c). Here $Q$ is computed according to Berland and Berland’s (1952) formula with a mean cloudiness value of 0.8. It is clear from Fig. 2 that within an equatorial band the symmetric semiannual component dominates over the antisymmetric annual component. The maximum semiannual variation of the solar radiation is about 10 W m⁻² just north of the equator (Fig. 2c) where the annual component nearly vanishes (Fig. 2b). Therefore, if the seasonal variation of SST were solely determined by the changes in the solar radiation, one anticipates that the near-equatorial SST response would be dominated by a semiannual cycle. Indeed, the solution of the forced mode obtained by using the realistic solar forcing shows a relatively strong semiannual cycle in the narrow equatorial band, as shown in Fig. 3.

The maximum SST variation of the semiannual component is about 0.2°C near the equator (Fig. 3c), where the annual component is nearly zero (Fig. 3b). The amplitude of total SST variability, therefore, varies from a semiannual cycle with an amplitude of 0.2°C near the equator to an annual cycle with an amplitude of about 2°C in the subtropics. There is, in general, a 90° phase delay in the SST response to the solar forcing. The structure of the forced mode bears, to a certain degree, similarity to the observed SST variation in the western Pacific, implying that the seasonal SST variation in that region may be a direct response to the solar forcing and the dynamic feedbacks play a secondary role. In the above calculation, we have assumed a constant value of 0.25 K m⁻¹ for the mean vertical temperature gradient $\bar{T}$, and a constant value of 5 x 10⁻⁴ cm s⁻¹ for the mean upwelling $W$. It will be demonstrated in the following analysis that the dynamic feedback between the SST and winds considerably modifies the near-equatorial SST response.

d. Effects of dynamic feedbacks on the near-equatorial SST response

The effect of dynamic feedback between the SST and winds on the seasonal cycle is governed by Eq. (9), whose homogeneous part gives a set of an infinite number of coupled modes $\tau^n$; that is, $\tau^n$ is an eigenvector given by the following eigenvalue problem:
\[ F(D(\tau^n)) + C(\tau^n) = -\sigma^n D(\tau^n) \]  

Fig. 2. (a) Seasonal variation of solar radiation as a function of latitude and time with (b) its annual component and (c) its semiannual component. Contour interval is 20 W m⁻² for (a) and (b), and 5 W m⁻² for (c).

with boundary condition (10), where $\tau^n$ is the eigenvalue and linear operators $F$, $C$, and $D$ were previously defined in (7) – (9). The properties of the governing coupled modes are discussed in great details in CP and can be briefly summarized as follows: 1) the eigenmodes have both symmetric and antisymmetric meridi-
which shows that the growth rate is proportional to the vertical gradient of the mean temperature but inversely proportional to the thermocline depth. In the eastern Pacific, because of the shallow thermocline and strong vertical temperature gradient, the gravest coupled modes are probably close to neutrally stable, or even unstable in some seasons such as late spring and early summer. However, in the western Pacific, where the thermocline is much deeper and the temperature gradient is much weaker, all the coupled modes are likely to be severely damped by the frictional processes. It is because of the strong damping that the coupled modes are much less important in the western Pacific than in the eastern Pacific, as will be demonstrated later.

The solution to (9) can be obtained by the normal mode expansion using the governing coupled modes $\tau^n$ as the trial functions. However, because the eigenvalue problem at hand is, in general, non-selfadjoint unless the mean state is strictly symmetric about equator, the solution must be expressed in terms of the governing eigenmodes $\tau^n$ and their adjoint components $\Gamma^n$:

$$\tau^n = \sum_{n=0}^{\infty} B_n(t) \tau^n,$$

where $B_n(t)$ is given by

$$B_n(t) = \frac{\int \exp(-\sigma_n^s t) \int_{-\infty}^{\infty} \Gamma^n C(\tau^n) dy ds}{\int_{-\infty}^{\infty} \Gamma^n D(\tau^n) dy}.$$

Once $\tau^n$ is determined, the SST response $T_s$ of the coupled mode can be obtained by solving Eq. (8) and the total response $T$ is then given by the superposition of $T_e$ and $T_s$. The above solutions to the coupled model can be readily evaluated numerically provided that the forcing function is specified. The following results are based on a finite difference discretization with a grid resolution of 0.4° in a computational domain between 20°S and 20°N.

Figure 4a shows the total SST response to the actual changes in the solar radiation shown in Fig. 2, assuming a uniform symmetric mean state (i.e., $\overline{\varepsilon} = \overline{F} = 0$) of $\overline{F} = 0.25$ K m$^{-1}$ and $W = 5 \times 10^{-4}$ cm s$^{-1}$. For comparison, we also plot the SST responses to the semiannual and annual changes in the solar radiation in Figs. 4b and 4c, respectively. In contrast to Fig. 3, the annual cycle of SST is greatly amplified near the equator with an antisymmetric meridional structure. The width of the near-equatorial annual cycle is about 250 km and the maximum amplitude is 2°C. Apparently, this narrow SST response is excited mainly by the annual component of the solar forcing that is nearly antisymmetric about the equator, but not by the symmetric semianual component (Figs. 4b,c). This is because the antisymmetric annual solar forcing has a much
stronger amplitude than the symmetric semiannual component and preferably excites the antisymmetric SST modes in the coupled system.

However, in reality, the SST response in the eastern equatorial Pacific is asymmetric relative to the equator with maximum variability located just south of the equator (Fig. 1a). This asymmetric structure is likely to be caused by the asymmetric mean state of the climate system in that region. To test this hypothesis, we repeat the calculations with two different mean conditions taken from the eastern and western Pacific Ocean, respectively. The first set of mean conditions is derived as follows: \( \overline{T}_e \) and \( \overline{T}_w \) are obtained by zonally averaging the vertical and meridional gradient of the annual-mean temperature between 110°W and 90°W. The temperature data are taken from the Levitus climatological dataset. An estimate of the annual-mean meridional velocity \( \overline{v} \) is obtained by taking a similar zonal average of the surface meridional velocity from ship drifter observations. The annual-mean upwelling \( \overline{w} \) is estimated from the horizontal divergence of the surface velocities, assuming a constant depth mixed layer of 50 m. The resultant mean conditions in the eastern Pacific Ocean are displayed in Fig. 5a. As can be seen, both the meridional SST gradient and the meridional surface velocity have maximum values just north of the equator. The mean vertical velocity below the mixed layer shows a strong upwelling near the equator and a downwelling near 5°N where convergence of surface currents occurs. The vertical temperature gradient at 50 m shows maximum value of 2°C m\(^{-1}\) near 5°N. The SST response in the presence of such a mean condition is shown in Fig. 5b. In comparison with the response in the symmetric mean state case (Fig. 4a), the SST variability north of the equator is greatly reduced, giving an asymmetric SST annual cycle. This asymmetric structure of the equatorial annual cycle, to a certain degree, resembles the observed seasonal variability of SST in the eastern equatorial Pacific, as shown in Fig. 1a. The reduction of annual cycle in the northern equatorial region is primarily due to the strong downwelling, which prevents the changes in the subsurface temperature from influencing the surface temperature variation. This effect is clearly demonstrated in Fig. 5c, where meridional advections of heat are removed from the calculation. A comparison of Figs. 5b and 5c suggests that, while the downwelling north of equator causes asymmetry of the SST annual cycle, the meridional advections of heat cause the strong annual cycle south of equator to move northward and enhance the annual variation of SST near the equator.

The mean conditions in the western Pacific are derived from similar datasets by taking zonal averages between 160°E and 180°. The results are shown in Fig. 6a. In comparison with the mean conditions in the eastern Pacific, both vertical and meridional temperature gradients are much weaker and the vertical velocity becomes more symmetric about the equator. In particular, no strong downwelling is observed in the off-equatorial regions. The SST response in this case, as shown in Fig. 6b, shows little effect due to the coupled modes. The semiannual signal, even though weakened, is still noticeable near the equator. The resultant seasonal cycle of SST bears a certain resemblance to observations in the western Pacific (Fig. 1c). The main reason for minimal impact of coupled modes on the

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**Fig. 4.** SST response to changes in (a) the total solar forcing, (b) annual component, and (c) semiannual component of the solar forcing in the presence of the symmetric mean climate condition. Contour interval as in Fig. 3.
seasonal cycle in the western Pacific is that the weak vertical and meridional mean temperature gradients prevent effective ocean–atmosphere interactions occurring in that region so that all coupled modes are severely damped.

3. Numerical experiments of the simple coupled model

In the calculations described in the previous section, we neglected zonal variations in the coupled model so that ocean–atmosphere interaction between the zonal wind and SST and the effects of meridional boundaries are excluded. In this section, we further examine the seasonal cycle by experimenting with a simple coupled numerical model that allows variations in both the zonal and meridional directions. The simple coupled model uses the LN model as its atmospheric component [see Eqs. (1)–(3)] and a Cane–Zebiak model as its oceanic component. Both component models, despite their simplicity, include dynamical processes of the low-level atmosphere and upper-ocean circulations essential for capturing the basic seasonal surface wind and SST variabilities in the eastern tropical Pacific. The version of the Cane–Zebiak ocean model is identical to the one used by Chang (1994) and Chang and Philander (1994). The model uses linearized momentum equations and continuity equation to calculate upper-ocean dynamic response to changes in surface winds and a fully nonlinear thermodynamic equation to compute changes in sea surface temperature. A simple parameterization scheme for the subsurface temperature change is given by

$$T_d = T_{sub} + \partial \overline{T}_{sub}(h - \overline{h}), \quad (15)$$

where $T_{sub}$ and $\partial \overline{T}_{sub}$ are respectively the annual-mean subsurface temperature and its vertical derivative at 50 m, taken from the Levitus's data, and $h - \overline{h}$ is the thermocline perturbation. The model is integrated using a finite difference scheme with a grid resolution of 1 degree in longitude and 0.5 degree in latitude. The model domain extends from 30°S to 30°N and 160 degrees in longitude. No-flux boundary conditions in temperature and free-slip boundary conditions in velocity are applied at the boundaries of the model ocean. A more detailed description of the coupled model is provided by CP.

In all the experiments described below, the coupled model was first integrated for 5 years with the observed annual-mean surface wind stresses and surface heat fluxes plus a Newtonian damping that relaxes the model SST to the observed annual-mean SST with a 60-day $e$-folding time scale. The model was then integrated for another 10 years forced by the seasonal varying solar radiation. The atmospheric and oceanic components are coupled using a simple linear relation between the anomalous surface wind stresses and wind speeds; that is,

![Diagrams showing mean climate condition in the eastern Pacific](image-url)
zonal direction by setting $|V| = 0$ in (16a). Finally, in the fourth experiment (expt 4) coupling is switched off in the meridional direction by setting $|V| = 0$ in Eq. (16b). The results from these four experiments are displayed in Figs. 7, 8, and 9.

Figure 7 shows SST responses at 100°W in the eastern Pacific Ocean. The effect of the mode coupling is evident by comparing the near-equatorial SST response in experiments 1 (Fig. 7a) and 2 (Fig. 7b). In experiment 2, where ocean–atmosphere interaction is fully active, the annual cycle of SST near the equator is dramatically amplified. This result is consistent with the analytical solution discussed in the previous section. An examination of the results from experiment 3 (Fig. 7c) and experiment 4 (Fig. 7d) reveals that the ocean–atmosphere interaction in the meridional direction, involving dynamic feedbacks between the meridional wind surface and SST, plays a more important role in producing a strong annual cycle in the eastern Pacific Ocean than does the interaction in the zonal direction. Similar SST plots taken at 180° in the western Pacific Ocean are shown in Fig. 8. It clearly demonstrates that the ocean–atmospheric interactions, both in the zonal and meridional directions, have little impact on the seasonal cycle. This result is again consistent with the analysis in the previous section.

Finally, Fig. 9 depicts SST responses along the equator. For comparison, the observed SST annual cycle along the equator is presented in Fig. 9e. In the absence of ocean–atmosphere interaction (expt 1), the seasonal SST variation along the equator is clearly dominated by a semiannual cycle. In this case there is little zonal variation in the SST response along the equator, as indicated in Fig. 9a. As ocean–atmosphere interaction is introduced in experiment 2, the SST annual cycle exhibits a well-defined westward phase propagation along the equator with decreasing amplitude (Fig. 9b). A similar feature is well observed in the realistic annual cycle of SST, as shown in Fig. 9e. In comparison with the observation, the simulated SST annual cycle suffers from defects in that the westward propagation is too slow and the amplitude is too weak. The westward propagation appears to be more closely associated with the interaction between the zonal wind stress (symmetric modes) and SST than the interaction between the meridional wind stress and SST (antisymmetric mode). However, the latter contributes more dominantly to the strong SST annual cycle in the eastern equatorial Pacific. These findings result from a direct comparison between Fig. 9c (expt 3) and Fig. 9d (expt 4). It shows that when symmetric modes are suppressed, the annual cycle is confined closely to the eastern Pacific (eastward of 100°W) (Fig. 9c), whereas when the antisymmetric modes are removed, the seasonal variation of SST along the equator is essentially dominated by a semiannual cycle (Fig. 9d). These results suggest that the ocean–atmosphere interactions in the zonal and meridional direction, although playing different roles, both contribute to the evolution of the
tropical ocean–atmosphere annual cycle in the eastern Pacific. The interactions between the meridional wind and SST are crucial to the development of the strong annual cycle in the near-coastal zone of the eastern Pacific (eastward of 100ºW). Without these interactions the near-equatorial SST variability would be dominated by the semiannual cycle. On the other hand, the interactions between the zonal wind and SST are instrumental in the westward expansion of the annual cycle along the equator. Without them the annual cycle would be confined to the eastern boundary region. These results appear to be in agreement with the earlier studies by Chang (1994) and Chang and Philander (1994). Recent analysis of the tropical ocean–atmosphere annual cycle by Nigam and Chao (1995) also gives support to the results.

4. Discussion

The tropical ocean–atmosphere seasonal cycle is likely to be influenced by both dynamic and thermodynamic interactions between the oceans and atmosphere. The former involves feedbacks between the surface winds and SST, whereas the latter may include
feedbacks between surface heat fluxes and SST through changes in surface evaporation and cloudiness. The study presented here focuses exclusively on the dynamic feedbacks, and thus it only reveals some of the important physical processes that influence the evolution of the tropical ocean–atmosphere seasonal cycle. Throughout this study, it is also assumed that the climatic mean state of the coupled system is given, and thus the influence of the seasonal cycle on the mean state is not taken into consideration. In reality, the mean state of the system is likely to be a result of the asymmetry of the seasonal cycle. Therefore, strictly speaking, the seasonal cycle should not be separated from the mean state. Nevertheless, the results from this study shed light on the role of dynamic ocean–atmosphere interactions in the evolution of the tropical seasonal cycle. They suggest that many important features of the seasonal cycle can be modeled with a relatively simple coupled model, provided that the climate mean conditions are correctly established. A similar conclusion is drawn by D. Battisti (1995, personal communication) in his study of the seasonal cycle using his version of the Cane and Zebiak coupled model. Xie (1994) found, using a different simple coupled model that mainly involves the feedbacks between surface evaporation and vertical mixing in the upper ocean, that the nature of
Fig. 9. Seasonal variations of the SST along the equator from (a) expt 1, (b) expt 2, (c) expt 3, (d) expt 4, and (e) observation. Contour interval is 0.1°C for (a) and 0.25°C for (b), (c), and (e).
the near-equatorial annual cycle such as its period, amplitude, and westward propagation is determined to a large extent by the mean state of the coupled system and that the meridional wind is the most important forcing for the SST annual cycle. He further noted that the evaporative feedback can contribute to the westward propagating feature in the annual cycle of near-equatorial zonal winds and SST. Recently, P. Schopf (1995, personal communication) also found that the unrealistically weak meridional wind in the eastern equatorial Pacific appears to be one of the major factors that contribute to the unrealistic seasonal cycle in their coupled general circulation model. More recently, in a coupled general circulation model study, Philander et al. (1996) pointed out a potentially important feedback between low-level stratus clouds over cold water and SST in the evolution of the tropical ocean–atmosphere annual cycle. These clouds, which form in regions of subsidence where the sea surface temperature is relatively cold, are capable of reflecting more than 30% of the incident solar radiation (Klein and Hartmann 1993). Philander et al. (1996) showed that the stratus clouds are crucially important to the formation of the cold tongue in the eastern Pacific and Atlantic Oceans, owing to a positive feedback between the clouds and SST; that is, lower SST strengthens the atmospheric inversion and hence favors more stratus clouds, which lower the temperature even further (Philander et al. 1996). It is anticipated that these clouds would have a significant impact on the seasonal cycle in the cold tongue regions. However, the extent to which these clouds could affect the tropical seasonal cycle is still undetermined. This is mainly due to a lack of understanding of stratus cloud dynamics so that a reliable parameterization scheme of the positive feedbacks between the clouds and SST is currently not available. Furthermore, Mitchell and Wallace (1992) argued for a leading role of the Central American monsoons in promoting the development of the SST cold tongue during May–September. Modeling studies also suggest the potentially important effects of orography, such as the Andes along the coasts of Peru and Chili, on the equatorial cold tongue. Currently, the interactions among oceans, atmosphere, and land are poorly understood. Future efforts to advance our understanding of the seasonal cycle must take these factors into account.

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