NOTES AND CORRESPONDENCE

On the Genesis of the Equatorial Annual Cycle

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14 July 1993 and 17 March 1994

ABSTRACT

A linear theory is proposed that can explain both the period and the westward propagation of the equatorial annual cycles in the SST and zonal wind. A coupled model linearized about a mean state of the air–sea system is used. This model allows the surface winds to directly change the SST through surface evaporation and vertical mixing, in contrast to the formulation of the conventional air-sea coupling models that emphasize the effects of winds on ocean dynamics. It is demonstrated that the characteristics of the equatorial seasonal cycle, including its period, phase, and amplitude, are determined to a large extent by the mean state of the air-sea coupling system. The meridional wind, which is specified here, is the most important forcing for the annual cycle in SST, suggesting that the equatorial annual cycle is the response to an annual solar forcing in the off-equatorial regions.

1. Introduction

In most of the world oceans the annual cycle in sea surface temperature (SST) is directly forced and readily explained by that in the solar radiation. An exception occurs on the equator, where the solar radiation is dominated by a semiannual cycle. Nevertheless, a pronounced SST annual cycle is observed at the equator in the Atlantic, eastern, and central Pacific, whose phase propagates westward from the eastern coasts (Horel 1982). The lack of an annual cycle in solar radiation and its phase variation in the zonal direction suggests that this equatorial annual cycle in SST is a product of ocean–atmosphere interactions.

Although both the annual and El Niño/Southern Oscillation (ENSO) cycles involve the development and decay of the equatorial cold tongue, there is an apparent difference in their thermodynamics. Figure 1 shows the time–depth section of ocean temperature obtained from a TOGA-TAO mooring at the equator and 110°W. The 1986–87 El Niño is associated with a significant deepening of thermocline. The change in the thermocline depth has been demonstrated to be the primary cause of the SST variation on the ENSO timescale (Philander and Seigel 1985). Although the SST annual cycle has a comparable amplitude with the ENSO cycle, the thermocline depth changes little in regular seasonal cycles of 1984 and 1985, compared to the large change during a period from 1986 to 1987. The annual signal in temperature is confined to the top 40 m of water (see also Fig. 8a of McPhaden and Taft 1988). In the boreal spring, the surface water starts warming up and stratifying. In the boreal fall, on the other hand, it cools down and a mixed layer develops. This stratification and destratification of the upper 40-m water column suggests that there is a large change in the strength of vertical mixing in the upper ocean on the seasonal timescale.

This study explores this vertical mixing mechanism for the SST annual cycle using a simple coupled ocean–atmosphere model. We consider a simple setting in which the solar radiation is the external forcing with a finite semiannual and a zero annual cycle at the equator. The model is linearized about a mean state of the air–sea coupling system. We will argue that the northerly position of the intertropical convergence zone (ITCZ) is a key to setting the period of the seasonal cycle. The maintenance of the northerly ITCZ is studied and reported in separate papers (Xie 1994; Xie and Philander 1994). The dynamic response of the ocean to the seasonally varying surface winds associated with the ITCZ and its thermodynamic consequences have been discussed by Philander and Pacanowski (1981) in an ocean-only model and by Mitchell and Wallace (1992) in an observational study. The same wind variation could have a direct effect on the mixed-layer temperature since it changes the turbulence kinetic energy flux and evaporation at the ocean surface, as revealed in limited turbulence measurements in the eastern equatorial Pacific (Herbert et al. 1991; Gregg et al. 1985). These diabatic effects will be emphasized in our model.

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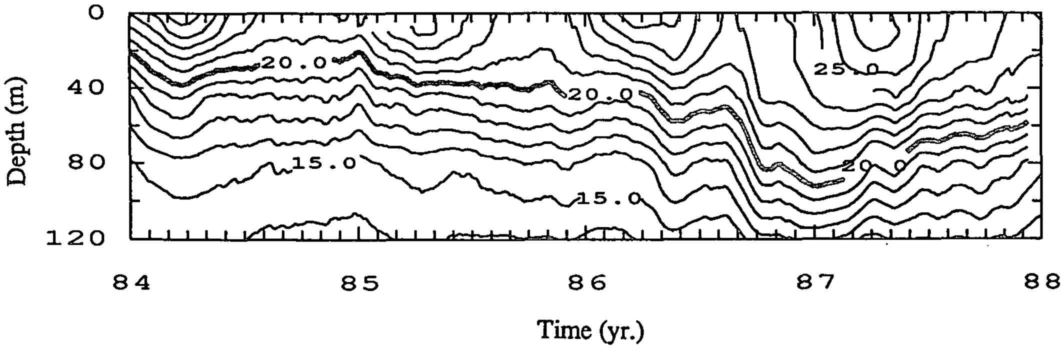

FIG. 1. Time–depth section of ocean temperature at 110°W from January 1984 to December 1987. Daily data are used and a 3-month running mean is applied to damp the disturbances with periods of 30 to 60 days. The significant deepening of the thermocline in 1986 and 1987 is associated with the El Niño.

We will show that this model is able to produce both an SST annual cycle and its westward phase propagation.

2. Model

The governing equation for the mixed-layer temperature T or SST may be written as

$$\frac{\partial T}{\partial t} + \mathbf{u}_0 \cdot \nabla T = \frac{Q/(\rho c_p) - w_e \Delta T}{h} + A\nabla^2 T, \quad (1)$$

where

$$Q = Q_{\text{shortwave}} + Q_{\text{longwave}} - L\rho_a C_E \Delta q |\mathbf{u}|, \quad (2)$$

$$w_e \Delta T = \frac{2mu_*^3}{\alpha g h} - \frac{Q}{\rho c_p}, \quad (3)$$

$\mathbf{u}_0$ denotes oceanic current velocity, $\mathbf{u}$ is the surface wind velocity, h is the depth of the mixed layer, and Δq is the moisture content difference between the sea surface and 10-m height. Other notations are standard. Because of its small size, sensible heat has not been included in the surface heat flux Q. The entrainment velocity (w_e) formula (3) is derived based on a turbulence energy conservation consideration (Kraus and Turner 1967), where ΔT is the temperature jump across the bottom of the mixed layer and $u_* = |\mathbf{u}| [(\rho_a/\rho)C_D]^{1/2}$ is the friction velocity. Note that both the latent heat flux and the entrainment rate are dependent on the surface wind speed. For the derivation of (1) and its application to the ENSO study, see Anderson and McCreary (1985) and Xie et al. (1989).

We assume that the thermocline depth does not change in time, which is true to the first order for a regular seasonal cycle (Fig. 1). This assumption filters out the ENSO cycle and allows us to focus on the annual cycle. We will neglect most of the advection terms since they are small on the seasonal timescale at the equator in the observations (Hayes et al. 1991). Results from GCM simulations are mixed; the advection plays a secondary role in some simulations (Koeberle and Philander 1994) and is important in others (E. Harrison 1993, personal communication). Under these assumptions and by linearizing (1)–(3) about an annual-mean state, an equation for SST perturbations at the equator can be obtained

$$\frac{\partial T'}{\partial t} + \bar{u}_0 \frac{\partial T'}{\partial x} = \frac{2}{H} \frac{1}{\rho c_p h_0} \times \left[Q_s - \left(1 + \frac{\eta}{H}\right) \bar{Q}_E \frac{Uu' + Vv'}{|\mathbb{U}|^2} \right] - \epsilon T', \quad (4)$$

where

$$\eta = \left(\frac{3m}{\alpha g h_0} U_*^3 \right) \Big/ \left(\frac{\bar{Q}_E}{\rho c_p} \right),$$

$H = h/h_0$ with h_0 denoting the depth at the eastern boundary, the overbar and prime denote the annual mean and perturbation, $\mathbb{U} = (U, V)$ the zonal and meridional components of the mean surface wind velocity, and Q_s and $\bar{Q}_E$ are perturbation solar radiation and mean latent heat flux, respectively. Hereafter we will drop the prime for perturbation for clarity. In the linearized equation (4), the relative importance between the latent heat flux and vertical mixing is measured by η/H. Note that diffusion terms and part of the latent heat flux term have been combined and represented in a Newtonian cooling form. Equation (4) is valid only at the equator or at the latitude of the center of the equatorial cold tongue where the meridional SST gradient vanishes. Off the equator the meridional advection becomes important. In fact, a substantial part of the annual signal in SST in the off-equatorial region within 5° latitude could simply be advected from the equator by the poleward Ekman flow.

The annual signal in the zonal wind in the Pacific exhibits the same westward propagation as in the SST (Horel 1982; Philander and Chao 1991), indicating a local coupling between the two. Lindzen and Nigam (1987) proposed that the pressure gradient in the surface boundary layer of the atmosphere be directly related to the SST gradient. Following Lindzen and Nigam, we assume

$$u = \mu \frac{\partial T}{\partial x}, \quad (5)$$

at the equator. We choose a value of $\mu = 2 \times 10^6$ $m^2 s^{-1} K^{-1}$, with which a change of 0.5 $m s^{-1}$ in surface wind speed results from SST change of 2.5 K in 10 000 km.

Substitution of (5) in (4) yields

$$\frac{\partial T}{\partial t} + (\bar{u}_0 - c)\frac{\partial T}{\partial x} = \frac{2}{H}\frac{Q_s}{\rho c_p h_0} - \frac{2}{H}\left(1 + \frac{\eta}{H}\right)\frac{\bar{Q}_E}{\rho c_p h_0}\frac{V}{|\mathbf{U}|^2}v - \epsilon T, \quad (6)$$

where

$$c = -\mu\frac{2}{H}\left(1 + \frac{\eta}{H}\right)\frac{\bar{Q}_E}{\rho c_p h_0}\frac{U}{|\mathbf{U}|^2}.$$

Note that the variation in the zonal wind enters (6) as a term associated with wave propagation at a phase speed c. This happens because of a 90° phase difference between the zonal wind and SST. The mean easterly trades are necessary for this westward propagation; $c = 0$, if $U = 0$. Since $U < 0$ in the central and eastern Pacific, $c > 0$ leads to a westward propagation. By using the parameters in Table 1 that are typical of the eastern equatorial Pacific, the phase speed of this westward propagation is estimated to be 1.5 $m s^{-1}$ for a 20-m deep mixed layer, much faster than the mean zonal current that is about 0.2 $m s^{-1}$. Neelin (1991) discussed a westward propagating coupled instability in a different model context where the Ekman pumping changes the SST and the direction of the mean winds is not important.

At the equator the solar radiation is dominated by a semiannual cycle,

$$Q_s = -\mathrm{Re}(\tilde{Q}_s e^{2i\omega t}), \quad (7)$$

where $\omega = 2\pi/(1\ \mathrm{yr})$. The clock is so set that $t = 0$ corresponds to 1 January. We need to determine another forcing function in (6), v. It seems that the meridional wind variation at the equator has less to do with the local SST change than the changes off the equator. As evidence, neither the amplitude nor the phase of the annual cycle in the meridional wind changes very much across the Pacific along the equator, whereas those of the SST cycle do (see Fig. 2 of Philander and Chao 1991). Outside a narrow equatorial band, the annual cycle dominates the solar radiation, which gives rise to an SST contrast between the hemispheres. This hemispheric SST difference, causing changes in atmospheric heating both through sensible heat and moist static stability, generates an annual cycle in the meridional wind that is nearly zonally uniform. Although it is possible to build a model that includes the off-equatorial region,[1] for simplicity, we prescribe v as a function of time,

TABLE 1. Model parameters.

c_p	4×10^3 $KJ^{-1} kg^{-1}$	ρ	10^3 $kg\ m^{-3}$	h_0	20 m
U	-2.5 $m s^{-1}$	V	2.5 $m s^{-1}$	$\tilde{v}$	1.25 $m s^{-1}$
$\bar{u}_0$	-0.2 $m s^{-1}$	η	1.5	μ	2×10^6 $m^2 s^{-1} K^{-1}$
$\tilde{Q}_s$	15 $W m^{-2}$	$\bar{Q}_E$	50 $W m^{-2}$	ϵ	$= 10^{-7}$ s^{-1}

$$v = -\mathrm{Re}(\tilde{v}e^{i(\omega t - \pi/4)}). \quad (8)$$

The annual cycle in the meridional wind at the equator can be thought of as the result of air–sea interactions in the off-equatorial regions. In (8), v reaches a maximum on 15 August, consistent with observations (Horel 1982). The lag behind the summer solstice can be due to such factors as the thermal inertia of the upper ocean in the off-equatorial region.

3. Results

By assuming

$$T = \tilde{T}_1 e^{i(\omega t - \pi/4)} + \tilde{T}_2 e^{2i\omega t}, \quad (9)$$

(6) gives

$$(\epsilon + in\omega)\tilde{T}_n + (\bar{u}_0 - c)\frac{\partial \tilde{T}_n}{\partial x} = \tilde{Q}_n, \quad n = 1, 2, \quad (10)$$

where

$$\tilde{Q}_1 = \frac{2}{H}\left(1 + \frac{\eta}{H}\right)\frac{\bar{Q}_E}{\rho c_p h_0}\frac{V}{|\mathbf{U}|^2}\tilde{v} \quad (11a)$$

$$\tilde{Q}_2 = -\frac{2}{H}\frac{\tilde{Q}_s}{\rho c_p h_0}. \quad (11b)$$

Here $\tilde{Q}_1$ and $\tilde{Q}_2$ are the annual and semiannual harmonics of the forcing and are due to the meridional wind and solar radiation, respectively; $\tilde{T}_1$ and $\tilde{T}_2$ are the annual and semi-annual harmonics of the SST response. The model parameters are listed in Table 1. The thermal damping timescale is about 100 days, a value typical for the Tropics (Hirst 1986; Zebiak and Cane 1987; Xie et al. 1989). The value of η is estimated to be 1.5 with $C_D = 1.5 \times 10^{-3}$ and $\alpha = 2 \times 10^{-4}$ K^{-1}, indicating that the wind-stirring effect is larger than the latent heat flux in the eastern ocean. The value of m is less certain and a range of values have been used in previous mixed-layer model studies. A value of 2.5, which is twice as large as the value adopted by Kraus and Turner (1967) for their midlatitude study, is used to reflect the observations that the vertical mixing is maximum on the equator presum-

[1] For example, consider three boxes of equal latitudial extent that are centered along 10°S, 0°, and 10°N, respectively. The SST in the equatorial box is governed by (6). In the off-equatorial boxes, the SST variation is largely determined locally by the solar radiation forcing that has a dominant one-year period. If one assumes that the meridional winds at the equator are driven by the SST gradient forcing in the atmospheric surface boundary layer, a central difference scheme for T_y involves only the SSTs in the two off-equatorial boxes. As a result, one could obtain such an annual cycle in v as that in (8). Liu and Xie (1994) have extended this model to include an explicit treatment of the off-equatorial region.

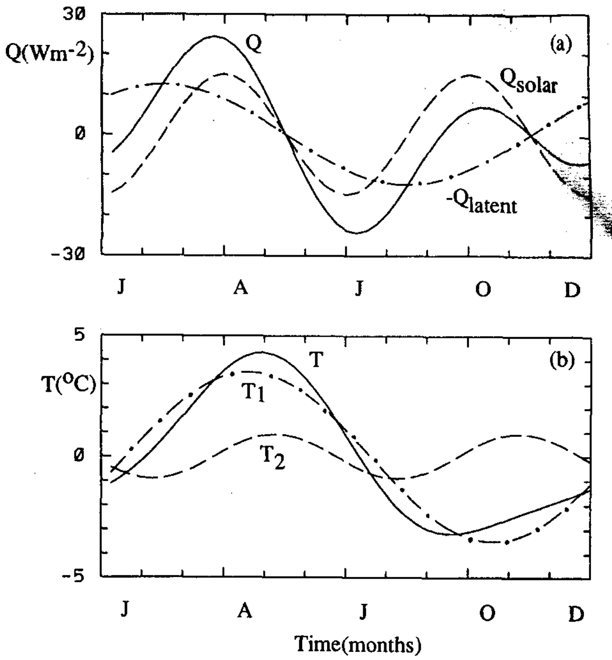

FIG. 2. (a) Surface heat flux ($Q = Q_{\mathrm{solar}} - Q_{\mathrm{latent}}$) and (b) SST ($T = T_1 + T_2$) anomalies (solid line) at the eastern boundary as functions of time. The first (dash-dotted line) and second (dashed line) harmonics of surface heat flux and SST are also plotted.

ably due to the large vertical shear of currents (Moum and Caldwell 1985). Equations (6) or (10) state that the local perturbation in SST is controlled by 1) weak westward advection of SST perturbations by the mean current, 2) "effective" westward advection of SST perturbations due to the effects of the zonal wind through entrainment and evaporation, 3) heating/cooling due to perturbations in solar forcing, 4) heating/cooling that results from entrainment and evaporation associated with perturbations in the meridional wind, and 5) dissipation.

To solve Eq. (10), a condition

$$u|_{x=0} = \mu T_x|_{x=0} = 0, \tag{12}$$

is imposed at the eastern boundary. Consistent with this boundary condition, the perturbation zonal wind and the zonal variation in perturbation SST are small around 100°W on the equator in observations (Horel 1982). Therefore, we may think of the eastern boundary of the model to be located at 100°W. East of 100°W, the seasonal cycle may be affected by coastal effects from the American continent. Under this boundary condition, the SST variation at the eastern boundary is controlled locally by solar and meridional wind forcings:

$$\tilde{T}_n = \frac{\tilde{Q}_n}{\epsilon + in\omega}, \quad \text{at} \quad x = 0. \tag{13}$$

An equatorially asymmetric mean state with the ITCZ in the Northern Hemisphere is necessary for the annual cycle in SST, which has a nonzero amplitude only if $V \neq 0$. Figure 2 shows the surface heat flux and SST variation at $x = 0$. Although the net surface heat flux is dominated by a semiannual cycle (Fig. 2a), the predominant annual cycle appears in SST (Fig. 2b). The larger amplitude of the first SST harmonic is attributed partly to the contribution from the vertical mixing, which does not show up in the surface heat flux, and is attributed partly to a frequency-selecting mechanism in Eq. (13). The frequency appears in the denominator on the rhs of (13) and the SST response is thus larger to lower-frequency forcing. In the boreal winter (months DJF) when the sun is in the Southern Hemisphere, the southerly winds converging onto the northerly ITCZ weaken. This weakens the vertical mixing and reduces surface evaporation, causing the equatorial warming in the boreal spring (MAM). The cooling in early fall (ASO), on the other hand, is caused by the strengthening of the southerlies. Compared to observations in the eastern Pacific, the model SST reaches its warmest value in late April or early May, one month too late. This is partly due to the late equinox on 30 March in the model compared with the real one. The phase of the model annual cycle depends also on the value of ϵ. It is interesting to note that the first harmonic of SST actually has a reasonably right timing.

The tilt of the thermocline along the equator, maintained by prevalent trades, can cause the zonal change in the amplitude of the seasonal cycle [see Eq. (11)]. This zonal variation in the SST perturbation will then propagate westward at a local speed of $(c - \bar{u}_0)$. To elucidate this effect, we let the depth of the mixed layer H be a simple function of x,

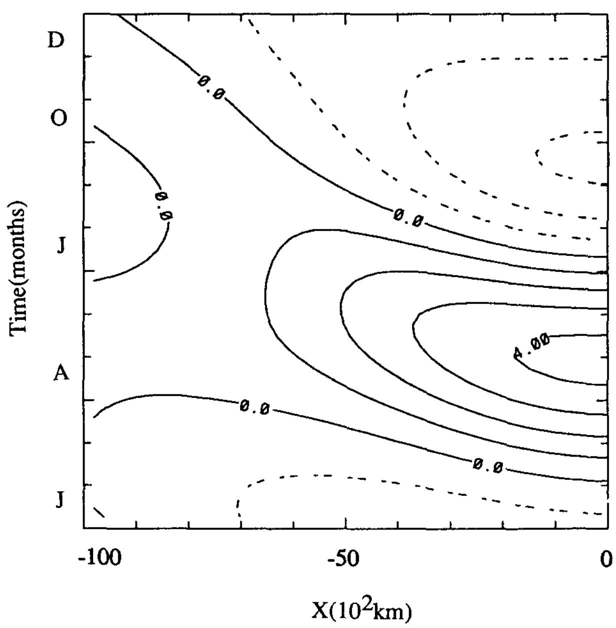

FIG. 3. Time–longitude section of SST anomaly (°C; contour interval = 1°C) at the equator. Contours with negative values are plotted in dashed line.

$$H(x) = 1 + ax, \quad a = -4 \times 10^{-7}\ \mathrm{m}^{-1}, \qquad (14)$$

corresponding to a distribution that the dimensional depth increases from 20 m at the eastern boundary to 100 m 10 000 km west. Figure 3 shows the time–longitude section of the SST anomaly. In the central and eastern part of the domain, the SST variation is dominated by an annual cycle and shows a westward propagation. Because of the westward increase of the mean heat content of the upper ocean, the amplitude of the SST annual cycle is largest on the eastern boundary and decreases westward in response to a zonally homogeneous relaxing of southerly winds in spring. This zonal SST gradient induces westerly wind anomalies along the equator, which weaken the prevalent easterly trades. The weakening of the trades reduces the vertical mixing and surface evaporation and therefore increases the SST in the interior. Since the warming by the trades relaxing is 90° out of phase with SST, the SST anomalies as well as the zonal wind anomalies propagate westward. The same is true for the cooling in fall. The westward propagation is primarily a result of the interaction between the SST and zonal wind, although the zonal advection by the oceanic current also makes a small contribution. When the advective term is removed, the major features of the SST variation in Fig. 3 do not change except for a small reduction in the distance that the annual signal can penetrate (not shown). In the western part of the ocean where the thermocline is deep and the vertical mixing is weak (or η is small), the change in SST is basically a direct response to the solar radiation forcing and is dominated by a semiannual cycle.

4. Discussion

A simple theory of tropical seasonal variation has been proposed, in which the permanent location of the ITCZ in the Northern Hemisphere is crucial for setting the one-year period. This note emphasizes a new air–sea coupling mechanism in which surface winds force the ocean in the thermodynamic equation rather than through momentum flux as in the conventional coupled models (Philander et al. 1984; Hirst 1986). In the present model, annual signals propagate westward in the SST and zonal wind, although the meridional winds change simultaneously along the equator. These features of the model annual cycle are consistent with the observations (Philander and Chao 1991). Although the SST response to wind variation is locally determined, the response of the zonal wind to SST is nonlocal, giving rise to the westward propagation.

This study suggests that the characteristics of the equatorial seasonal cycle be determined to a large extent by the mean state of the air–sea system. 1) A hemispherically asymmetric circulation of the atmosphere, particularly the northerly ITCZ and the mean southerly cross-equatorial winds, are necessary for the annual cycle in SST. 2) The presence of the easterly trades is crucial for the westward propagation of the annual cycle. 3) The tilt of the mean thermocline along the equator causes the zonal changes in the period, phase, and amplitude of the seasonal cycle. In the eastern part of the ocean where the thermocline is shallow, the annual cycle prevails, while in the west where the thermocline is deep, a semiannual cycle is excited by direct solar forcing.

Locally at the equator, an apparent mismatch between the principal periods of the net heat flux and SST is observed at 110°W (Hayes et al. 1991). This mismatch is not surprising since the direct solar forcing has only a minor effect on the SST variation. It is the vertical mixing effects, though not appearing in the surface heat flux, that account for most of the model annual cycle in SST. The annual cycle in the strength of the vertical mixing as inferred from Fig. 1, on the other hand, can to a large extent be attributed to that in the meridional winds. The equatorial annual cycle is not locally determined but is remotely forced by the annual cycle in solar radiation in the off-equatorial region, in the sense that the meridional wind variation at the equator is remotely forced by the solar radiation. For simplicity, the annual cycle in the meridional wind is specified as a given function in this note. Taking one step further, Xie and Philander (1994) are able to produce an annual cycle in SST at the equator in a coupled ocean–atmosphere model that develops its own northerly ITCZ and is forced solely by the solar radiation. The genesis mechanisms proposed here are responsible for their SST annual cycle.

The ENSO and the equatorial annual cycle share some development features as far as SST is concerned. This study indicates that they are generated by different mechanisms and have different dynamic characteristics. First, from the observations, the annual cycle has a much shallower vertical structure than the ENSO. Indeed, it is shown that local surface mixed-layer processes can produce an annual cycle with a reasonable amplitude as well as a realistic westward phase propagation. Second, the annual cycle is essentially a forced problem at the equator. Indeed, the unforced free solution to Eq. (6) represents a damped mode, in contrast to the ENSO that is caused by the internal instability of the coupled ocean atmosphere. Third, the meridional wind is an important forcing for SST on the seasonal timescale, whereas it plays at most a secondary role in the ENSO.

Between 10°N and 10°S the SST annual cycle has a maximum amplitude along the equator, whereas the variation in southerlies has a much broader meridional scale and is maximum north of the equator. If the latter cycle is the major forcing of the former, there is an apparent discrepancy between the two in the meridional scale. The equatorially trapped structures in the thermocline depth and in the vertical shear of the zonal current favor a similar trapped structure in SST and might account for the discrepancy. The shallow thermocline and strong vertical shear are presumably the cause of the strong turbulence dissipation rate observed near the equator (Moum and Caldwell 1985; Hebert et al. 1991). There are other effects, such as those of the cloudiness, of the change in the mixed-layer depth, and of the off-equatorial upwelling and downwelling, which should be taken in account and assessed in the future study.

Acknowledgments. The author would like thank G. Philander, I. Held, J. Wallace, A. Kubokawa, C. Koeberle, and R. Pacanowski for discussions; M. McPhaden, the TOGA-TAO Project Office of NOAA/PMEL, for providing the mooring data; and E. Sarachik, T. Mitchell, and an anonymous referee for comments on the manuscript. This study is funded by a grant from the National Oceanic and Atmospheric Administration (NA26G0102-01). The views expressed herein are those of the author and do not necessarily reflect the views of NOAA or any of its agencies. This work was presented at a workshop on the ENSO and annual cycle at Princeton University in June 1992.

REFERENCES

Anderson, D. L. T., and J. P. McCreary, 1985: Slowly propagating disturbances in a coupled ocean–atmosphere model. *J. Atmos. Sci.,* **42,** 615–929.

Gregg, M. C., H. Peters, J. C. Wesson, N. S. Oakey, and T. J. Shay, 1985: Intensive measurements of turbulence and shear in the equatorial undercurrent. *Nature,* **318,** 140–144.

Hayes, S. P., P. Chang, and M. J. McPhaden, 1991: Variability of the sea surface temperature in the eastern equatorial Pacific during 1986–1988. *J. Geophys. Res.,* **96,** 10 553–10 566.

Hebert, D., J. N. Moum, and D. R. Caldwell, 1991: Does ocean turbulence peak at the equator? Revisited. *J. Phys. Oceanogr.,* **21,** 1690–1698.

Hirst, A. C., 1986: Unstable and damped equatorial modes in simple coupled ocean–atmosphere models. *J. Atmos. Sci.,* **43,** 606–630.

Horel, J. D., 1982: On the annual cycle of the tropical Pacific atmosphere and Ocean. *Mon. Wea. Rev.,* **110,** 1863–1878.

Koeberle, C., and S. G. H. Philander, 1994: Seasonal variations of sea surface temperatures in the tropical Pacific Ocean. *Tellus,* **46A,** 481–496.

Kraus, E. B., and J. S. Turner, 1967: A one-dimensional model of the seasonal thermocline. II: The general theory and its consequences. *Tellus,* **19,** 98–109.

Lindzen, R. S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the Tropics. *J. Atmos. Sci.,* **44,** 2418–2436.

Liu, Z., and S.-P. Xie, 1994: Equatorward propagation of coupled air–sea disturbances with application to the annual cycle of the eastern tropical Pacific. *J. Atmos. Sci.,* **51,** 3807–3822.

McPhaden, J. M., and B. A. Taft, 1988: Dynamics of seasonal and intraseasonal variability in the eastern equatorial Pacific. *J. Phys. Oceanogr.,* **18,** 1713–1732.

Mitchell, T. P., and J. M. Wallace, 1992: The annual cycle in equatorial convection and sea surface temperature. *J. Climate,* **5,** 1140–1156.

Moum, J. N., and D. R. Caldwell, 1985: Local influences on shear-flow turbulence in the equatorial ocean. *Science,* **230,** 315–316.

Neelin, J. D., 1991: The slow sea surface temperature and the fast-wave limit: Analytic theory for tropical interannual oscillations and experiments in a hybrid coupled model. *J. Atmos. Sci.,* **48,** 584–606.

Philander, S. G. H., and R. C. Pacanowski, 1981: The oceanic response to cross-equatorial winds (with application to coastal upwelling in low latitudes). *Tellus,* **33,** 201–210.

——, and A. D. Seigel, 1985: Simulation of El Niño of 1982–1983. *Coupled Ocean–Atmosphere Models,* J. C. J. Nihoul, Ed., Elsevier, 517–541.

——, and Y. Chao, 1991: On the contrast between the seasonal cycles of the equatorial Atlantic and Pacific Oceans. *J. Phys. Oceanogr.,* **21,** 1399–1406.

——, T. Yamagata, and R. C. Pacanowski, 1984: Unstable air–sea interactions in the Tropics. *J. Atmos. Sci.,* **41,** 604–613.

Xie, S.-P., 1994: The maintenance of an equatorially asymmetric state in a hybrid coupled GCM. *J. Atmos. Sci.,* **51,** 2602–2612.

——, and S. G. H. Philander, 1994: A coupled ocean–atmosphere model of relevance to the ITCZ in the eastern Pacific. *Tellus,* **46A,** 340–350.

——, A. Kubokawa, and K. Hanawa, 1989: Oscillations with two feedback processes in a coupled ocean–atmosphere model. *J. Climate,* **2,** 946–964.

Zebiak, S. E., and M. A. Cane, 1987: A model of El Niño–Southern Oscillation. *Mon. Wea. Rev.,* **115,** 2262–2278.