Chapter 2
Roles of Air–Sea Interaction in Shaping Tropical Mean Climate

Abstract In this chapter, the observed characteristics of the mean climate such as the equatorial asymmetry of the intertropical convergence zone (ITCZ) and the equatorial annual cycle in the tropical Pacific are first described. Next the physical mechanisms responsible for the ITCZ asymmetry and the annual cycle are discussed, from theoretical analysis and idealized modeling perspectives.

2.1 ITCZ Asymmetry

The tropical time–mean climate in the eastern Pacific and Atlantic exhibits a remarkable asymmetry relative to the equator, with abundant rainfall and the intertropical convergence zone (ITCZ) residing north of the equator, and cold SST tongue at and south of the equator (Fig. 2.1). From a pure meteorological point of view, the ITCZ asymmetry is attributed to the north–south SST asymmetry. From a pure oceanographic point of view, the asymmetry in SST is caused by the asymmetry in the atmospheric forcing field. The circular argument suggests that air–sea interaction may play a role.

If one regards the coupled atmosphere–ocean system as a whole, then only external forcing is solar radiation at top of the atmosphere, which, for long-term mean, is perfectly symmetric about the equator. Then a natural question is, given such a symmetric forcing, why is the response of the coupled system asymmetric? What are the fundamental physical processes that cause the asymmetry? Why does such an asymmetry appear only in the eastern Pacific and the eastern Atlantic, not in the western Pacific and Indian Ocean?

Another fascinating feature of the mean state is the annual cycle of SST and wind fields in the equatorial eastern Pacific and eastern Atlantic. Observations show that both the SST and surface wind in the regions exhibit a marked 12-month period (Fig. 2.2), but the solar radiation at the top of the atmosphere is dominated by a semiannual (6 months) period, as the sun crosses the equator twice a year. Again such an annual frequency could be easily interpreted from a pure meteorological or a pure oceanographic point of view. However, considering the Earth climate is a coupled atmosphere–ocean system, one has to explain why the annual frequency is of atmospheric and oceanic variables and is observed while the forcing is semiannual.

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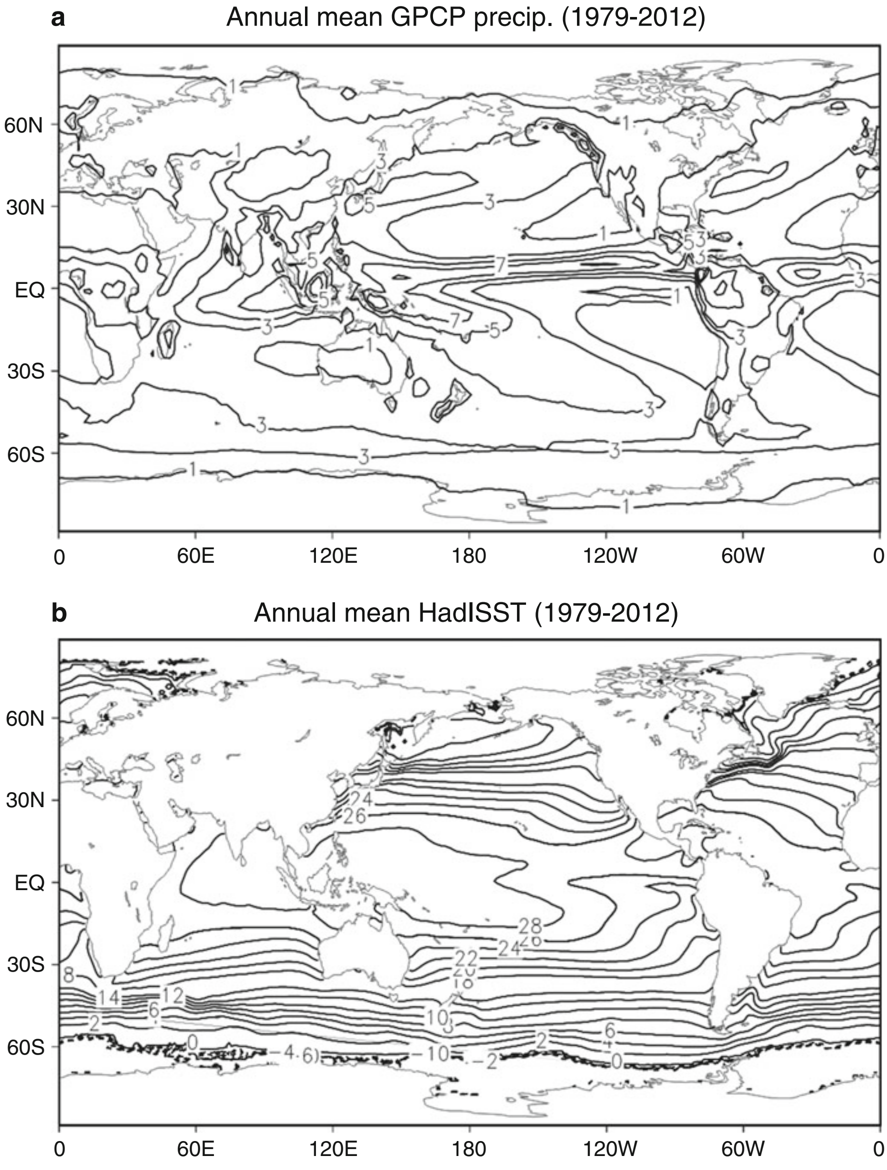

Fig. 2.1 Patterns of global annual mean precipitation (*top*, unit: mm/day) and SST (*bottom*, unit: °C) averaged during 1979–2012 (Source: Global Precipitation Climatology Project (GPCP) monthly precipitation and Hadley Centre Sea Ice and Sea Surface Temperature data (HadISST), respectively)

To address the science questions above, one needs first to examine the observed evolution features of OLR, wind, and SST in the tropical eastern Pacific domain from March (when the Pacific cold tongue is weakest) forward. Figure 2.3 shows

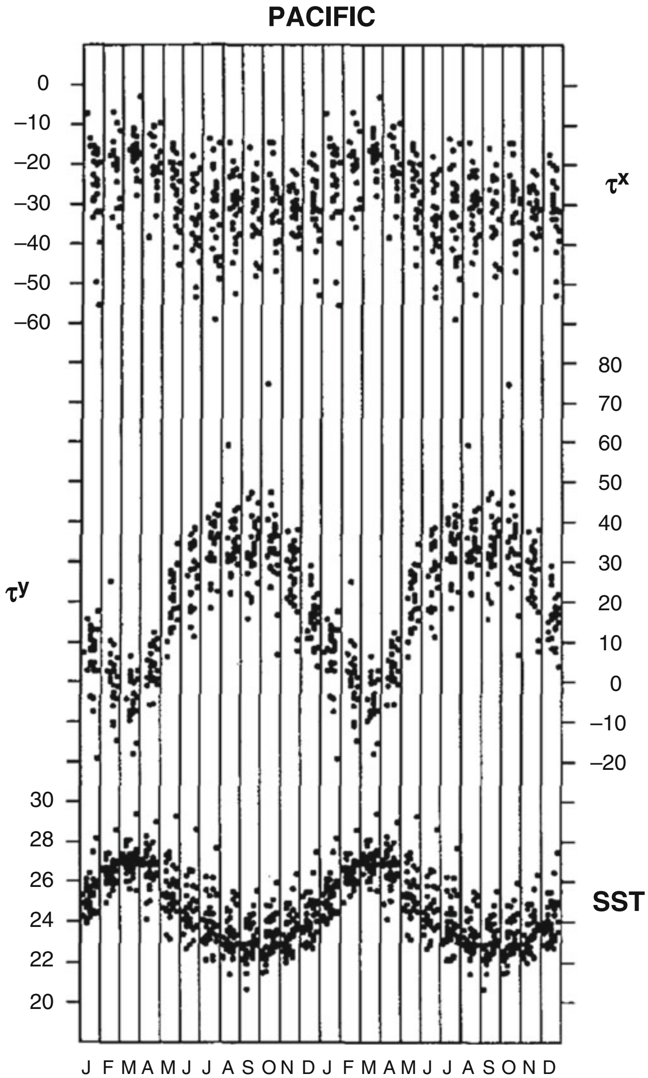

Fig. 2.2 Time evolution of zonal wind stress to the west (4°S–4°N, 130°–110°W) and meridional wind stress to the north (0°–8°N, 120°–100°W) of the Pacific cold tongue and SST at the cold tongue (4°S–4°N, 104°–86°W) (From Mitchell and Wallace (1992). © Copyright 1992 American Meteorological Society (AMS))

the difference fields from March to May (May minus March) and from May to August (August minus May). As time progresses, north–south SST gradients increase and so do the cross-equatorial flows and the antisymmetric components of SLP and OLR fields. Therefore, the observational analysis implies that the cold tongue development experiences a positive feedback loop between the atmospheric

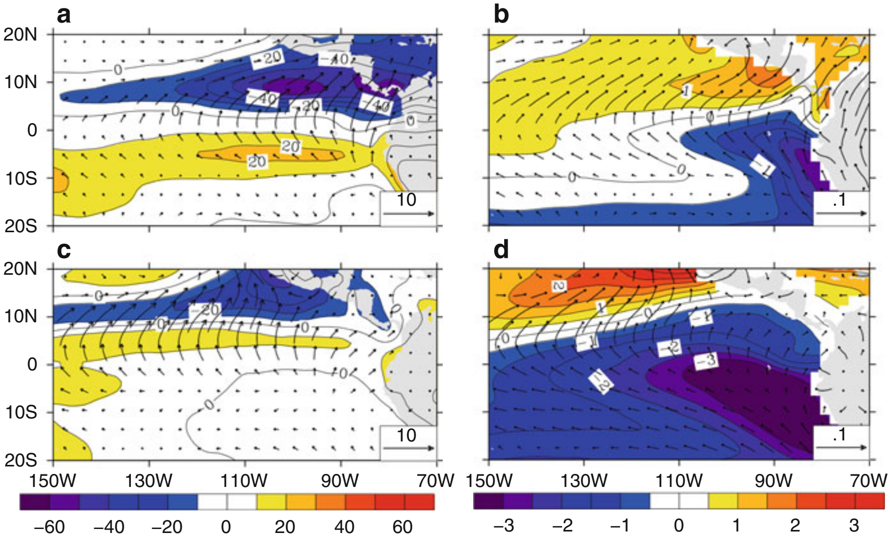

Fig. 2.3 The difference fields of observed (*left*) climatological OLR (contour interval, 10 W m^{-2}) and surface wind (m s^{-1}) and (*right*) SST (contour interval, 0.5 °C) and wind stress (Nt m). The *top panel* shows the difference between March and May (May minus March), and the *bottom panel* shows the difference between May and August (August minus May) (After Mitchell and Wallace (1992). © Copyright 1992 American Meteorological Society (AMS))

and oceanic antisymmetric modes, that is, an antisymmetric SST pattern induces cross-equatorial winds, which may further strengthen the north–south SST gradient.

2.2 Theories

Motivated by the observed evolution patterns, several theories have been proposed to understand the ITCZ asymmetry. The main idea behind the theories is that air–sea interaction may favor most unstable growth of an antisymmetric mode. The first theory, proposed by Chang and Philander (1994), focused on a positive dynamic air–sea feedback among meridional wind, ocean upwelling, and SST. They considered a simple coupled air–sea model listed below. The atmospheric component is the Lindzen–Nigam model, and the ocean component is a simplified Cane–Zebiak model:

$$E\tau^x - f\tau^y = -\alpha\left(\frac{\partial \phi}{\partial x} - A\frac{\partial T}{\partial x}\right) \tag{2.1}$$

$$E\tau^y + f\tau^x = -\alpha\left(\frac{\partial\phi}{\partial y} - A\frac{\partial T}{\partial y}\right) \tag{2.2}$$

$$\phi + B\left(\frac{\partial\tau^x}{\partial x} + \frac{\partial\tau^y}{\partial y}\right) = 0 \tag{2.3}$$

where τ^x and τ^y are the zonal and meridional component of the surface wind stress, ϕ is the geopotential at the top of the boundary layer, T is the sea surface temperature, and E is a mechanical damping due to the vertical diffusion of momentum and surface drag. $A = gH_0/2T_0$ and $B = gH_0/\mu\alpha$ 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; μ is an inverse relaxation time for the adjustment of the boundary layer height; and α converts the wind speeds into surface wind stresses.

It was shown that thermocline variation on the annual timescale is negligible. Neglecting the upper-ocean wave and thermocline variation, the Cane–Zebiak model may be simplified as

$$T_t + \bar{u}_1 T_x + \bar{v}_1 T_y + \frac{\bar{w}}{H_1}(T - \gamma h) + \bar{u}_1 T_x + v_1 \bar{T}_y + w\bar{T}_z + \varepsilon T - \kappa\nabla^2 T = 0 \tag{2.4}$$

$$\begin{bmatrix} u_1 \\ v_1 \end{bmatrix} = \frac{H_2}{H}\begin{bmatrix} u_e \\ v_e \end{bmatrix}, \quad w = \frac{H_1 H_2}{H}\nabla \cdot V_e \tag{2.5}$$

$$\begin{bmatrix} u_e \\ v_e \end{bmatrix} = \frac{1}{\Delta_0}\begin{bmatrix} r_s\tau^x + f\tau^y \\ r_s\tau^y - f\tau^x \end{bmatrix}, \quad \Delta_0 = H_1\left(r_s^2 + f^2\right), \tag{2.6}$$

where H_1 and H denote the depth of a constant mixed layer and the mean depth of thermocline; H_2 is the difference between H and H_1; ($\bar{u}_1$, $\bar{v}_1$) and (u_1, v_1) represent the mean and perturbation zonal and meridional currents; h denotes the thermocline depth anomaly, which is assumed to be zero.

An eigenvalue analysis of the simple coupled model above indicates that there are two unstable modes (Fig. 2.4). The first mode has an antisymmetric structure relative to the equator. The second mode has an equatorially symmetric structure. The former has a greater growth rate than the latter, indicating that atmosphere–ocean interaction favors mostly the equatorially antisymmetric mode. Note that the most unstable mode has a zero wave number, implying that this mode supports a zonally uniform structure.

The schematic diagram of Fig. 2.5a illustrates what air–sea interaction processes are involved in leading to unstable development of an antisymmetric mode. Consider a perfectly symmetric background mean state. Initially, we introduce a weak antisymmetric SST perturbation, with a positive (negative) SSTA north (south) of the equator. If the symmetric mean state is stable, the initial perturbation would dissipate with time, due to natural dissipation process. If the system is unstable, the

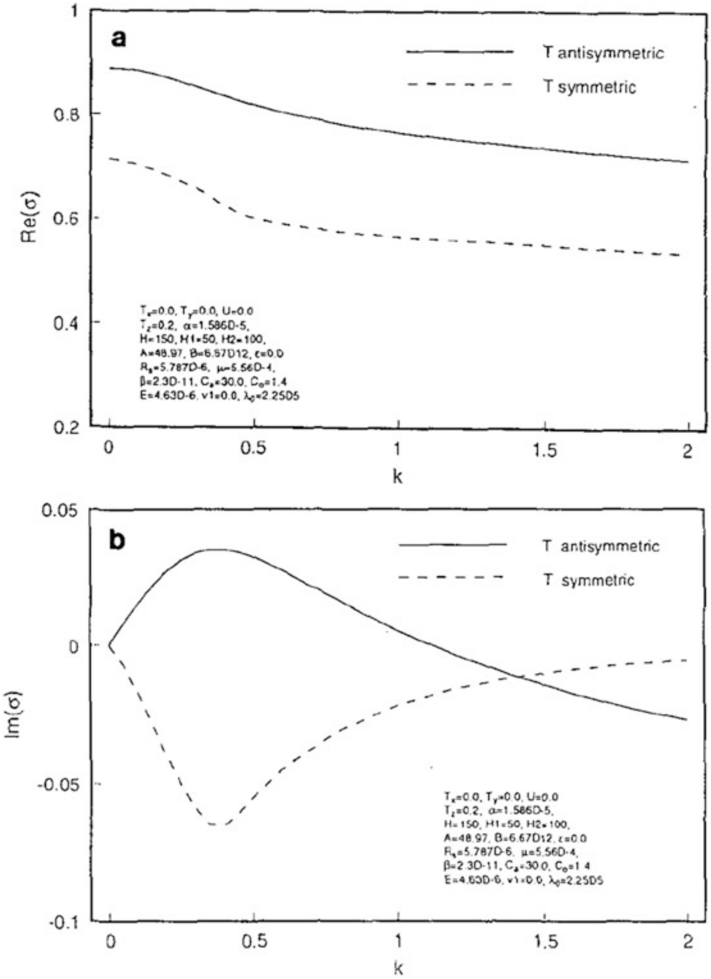

Fig. 2.4 (**a**) Growth rate and (**b**) frequency of the coupled model of Chang and Philander (1994) as a function of zonal wave number. *Solid* and *dashed lines* represent the antisymmetric and symmetric SST modes, respectively (From Chang and Philander (1994). © Copyright 1994 American Meteorological Society (AMS))

antisymmetric mode would grow exponentially with time. As a result, an antisymmetric mean state will be established.

In response to the initial weak antisymmetric SSTA forcing, a southerly wind anomaly is induced. The southerly wind converges onto the anomalous warm water, over which convection occurs, and drives the oceanic Ekman currents. The ocean Ekman current has a maximum northward component at the equator where the Coriolis force vanishes. Far away from the equator, the oceanic Ekman currents have a smaller northward component because the Coriolis force deflects the wind-driven ocean currents to the left (right) in the southern (northern) hemisphere. As a

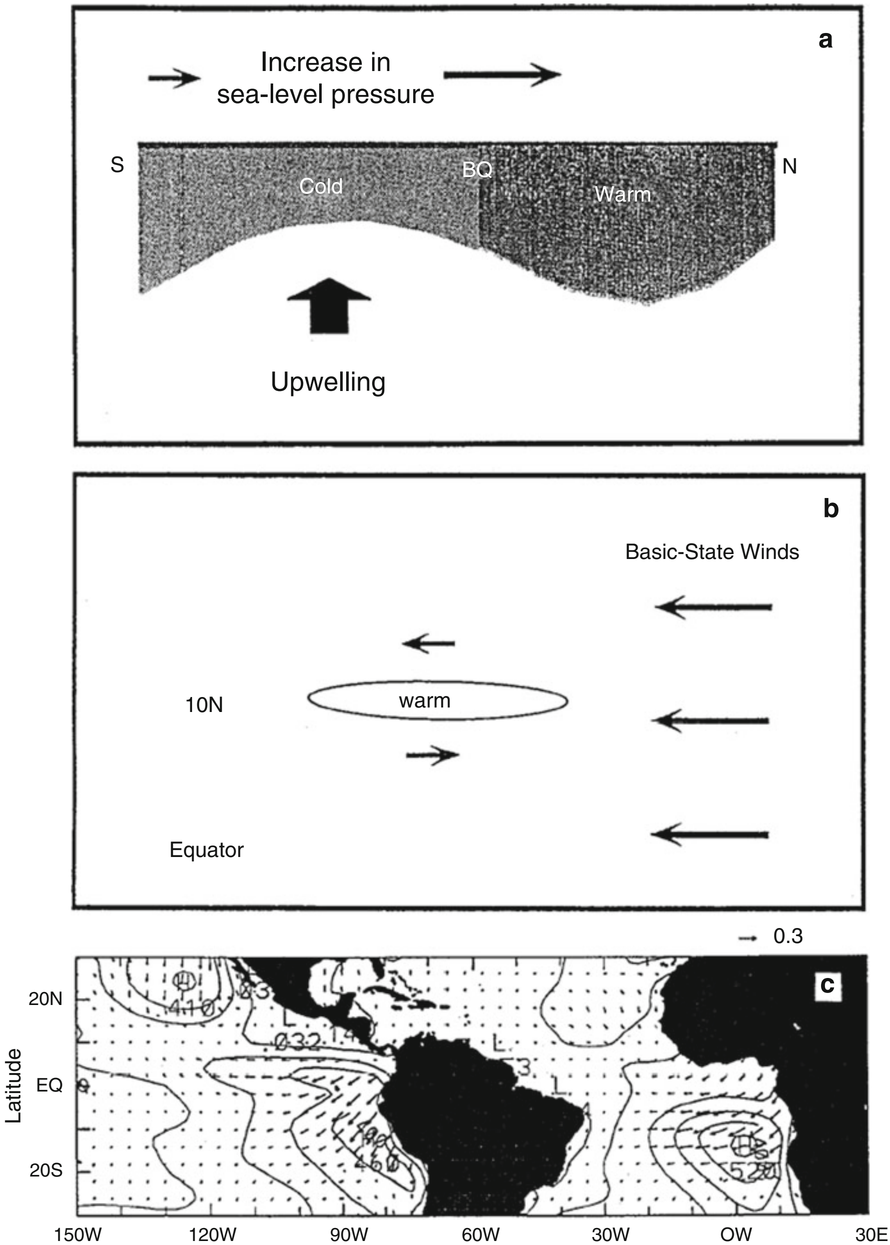

Fig. 2.5 Schematic diagrams illustrating key processes involved in the meridional wind–upwelling–SST feedback (**a**) and the wind–evaporation–SST feedback (**b**). Contours and vectors in (**c**) show the annual mean stratus cloud amount and the amplitude/phase of annual cycle of the stratus cloud (with downward/westward indicating July/October maximum) (From Li (1997). © Copyright 1997 American Meteorological Society (AMS))

result, the northward component of the oceanic Ekman current leads to surface divergence (convergence) south (north) of the equator. This promotes an asymmetry anomalous vertical velocity at bottom of the oceanic surface (or mixed) layer. An upwelling (downwelling) appears at south (north) of the equator. The asymmetric vertical velocity further strengthens the initial antisymmetric SST perturbation. As a result, there is a positive feedback loop among the anomalous SST, meridional wind, and SST. Through this positive feedback cycle, the asymmetric SST perturbation grows.

The coupled air–sea instability mentioned above was referred to as the "meridional wind–upwelling–SST" feedback (Li 1997). In addition to this dynamic air–sea feedback, there are two types of thermodynamic air–sea feedbacks, with one involving the "wind–evaporation–SST" feedback (Xie and Philander 1994) and another involving the "stratus cloud–radiation–SST" feedback (Li and Philander 1996; Philander et al. 1996).

The "wind–evaporation–SST" feedback was originally proposed by Neelin et al. (1987) and Emmanuel (1987), in studying a prominent atmospheric low-frequency oscillation, the Madden–Julian oscillation (MJO). It was further applied by Xie and Philander (1994) in studying the ITCZ asymmetry problem. The key to this mechanism lies in a background wind distribution, depicted in Fig. 2.5b. Suppose initially there is a positive SST perturbation located at 10°N. In response to this SST forcing, westerly (easterly) wind anomalies are generated to the south (north) of the SST perturbation. The westerly anomalies to the south tend to reduce the surface evaporation because the basic-state winds are easterlies. This leads to a positive time change rate for SST. As a result, the SST anomaly intensifies and propagates equatorward. Near the equator, strong oceanic upwelling induced by mean easterlies causes an extremely cold SST tongue, which suppresses the atmospheric convection and prevents further equatorward movement of the ITCZ. As a result, the positive SST perturbation and associated anomalous convection have to stay a few degrees in latitude away from the equator. It has been shown that without the equatorial cold tongue, a case in the western Pacific, the maximum SST center would move to the equator.

The third type of air–sea interaction involves the positive feedback between the low-level marine stratus clouds and SST (Li and Philander 1996; Philander et al. 1996). Whereas the convective clouds favor the warm waters in the western Pacific, the low-level stratus clouds form over colder water in the eastern Pacific (Fig. 2.5c). These stratus clouds vary seasonally and have maximum values in September when the SST is lowest. The low-level stratus clouds are particularly important in a coupled ocean–atmosphere system because they are involved in a positive feedback cycle: the lower the SST, the larger the static stability of the lower troposphere, the stronger the atmospheric inversion, and the thicker the deck of low-level stratus clouds; the increase in the clouds further shields the ocean from shortwave radiation and causes even lower SST. An observational data analysis revealed that there is a negative correlation between the low-level stratus cloud amount and SST in the eastern tropical Pacific and Atlantic. With a decrease in SST, the low-level stratus clouds increase dramatically, which further reduces the shortwave radiation into the ocean and results in a colder SST.

The discovery of the positive ocean–atmosphere interaction processes provides a theoretical basis in understanding the fundamental cause of the climate asymmetry. But what is the relative importance of the aforementioned three air–sea feedbacks in shaping the climate asymmetry? Li (1997) attempted to address the question by considering the three air–sea feedback processes in a unified framework. In a realistic parameter regime, the aforementioned three air–sea feedback processes are of equal importance in contributing to the observed asymmetry in the eastern Pacific.

Positive air–sea feedbacks favor the development of an antisymmetric mode with ITCZ being located in either hemisphere depending on the sign of the initial perturbation. Why does the nature select northern hemisphere (NH) as a preferred location of ITCZ? Furthermore, in reality the ITCZ asymmetry happens only in the eastern tropical Pacific and eastern Atlantic, not in the western Pacific and Indian Ocean. What determines the preferred longitudinal location? These questions will be addressed in the next section.

2.3 Effects of Asymmetric Land Mass and Coastal Geometry

One possible cause of the preferred NH location of ITCZ is greater land mass in NH. An atmospheric GCM (GFDL R30 model) was used to address this hypothesis (Philander et al. 1996; Li 1997). To understand the sole effect of the asymmetric land mass, a symmetric SST relative to the equator is specified as the model lower-boundary condition; the annual mean insolation is specified at the top of the atmosphere, and it is symmetric about the equator. In spite of greater land mass in NH, the mean wind simulated by the model remains symmetric about the equator in the tropical Pacific. The result is quite different in the tropical Atlantic, where there are northward cross-equatorial winds. The cause of this wind asymmetry is attributed to a thermal contrast between the heated northwestern Africa and the cooler ocean to its south.

Two important features of the surface wind field in the tropical Pacific (Fig. 2.6b) are worth noting. Firstly, there are strong easterlies at the equator even though the SST does not vary zonally. Such winds, when allowed to influence the ocean, can drive warm surface waters westward and upwell cold water from below to the surface in the east, resulting in a strong east–west asymmetry in SST (the warm-pool–cold-tongue thermal contrast). This type of air–sea interaction along the zonal direction is responsible for determining the strength of the Walker circulation. Secondly, because of the tilt of the American coast, the trade winds to the south (north) of the equator are essentially parallel (perpendicular) to the coast. As we know, the parallel-to-coast winds may induce strong upwelling along

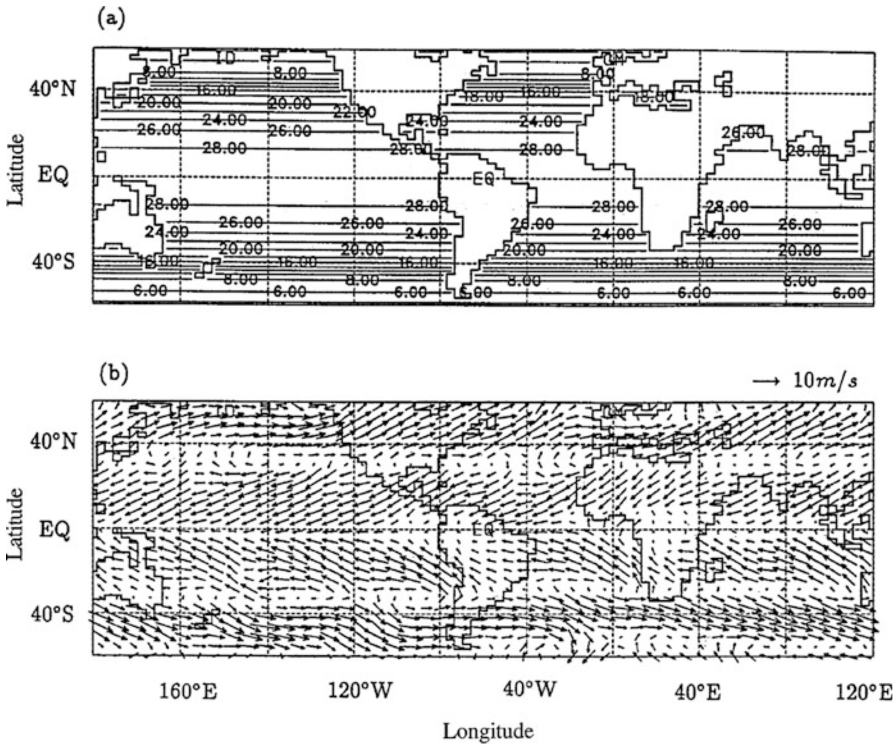

Fig. 2.6 (**a**) A prescribed zonal-mean symmetric SST (°C) field derived from the climatological monthly mean COADS dataset (averaged between 140°E and 180°). (**b**) The R30-model simulated surface wind field forced by the above prescribed symmetric SST and annual mean solar radiation (From Li (1997). © Copyright 1997 American Meteorological Society (AMS))

the coast and cool SST there. Therefore, when coupled to the ocean, the winds can help set up a north–south asymmetry in SST by inducing anomalous cooling (warming) off the coast of Peru (Panama). Once this initial weak SST asymmetry is set up, the ocean–atmosphere feedbacks mentioned in the previous section could further amplify the equatorial asymmetry, to finally reach the observed strength.

To test this hypothesis and to understand the relative role of various air–sea feedbacks, a set of coupled sensitivity experiments were conducted. In the first set, a focus was on the dynamic air–sea coupling, that is, the atmosphere influences the ocean solely through wind stresses, while the surface heat fluxes in the model were specified as a Newtonian damping term that restores the model SST toward a prescribed symmetric SST field shown in Fig. 2.6a. The ocean model starts from a symmetric SST condition (as shown in Fig. 2.6a). Because of the quick development of equatorial easterlies, an east–west asymmetry, characterized by a warm pool in the western Pacific and a cold tongue in the eastern Pacific (Fig. 2.7a), is established. Because of the tilt of the western coasts of the Americas, a north–south asymmetric SST pattern develops (Fig. 2.7a). Accompanied with the development

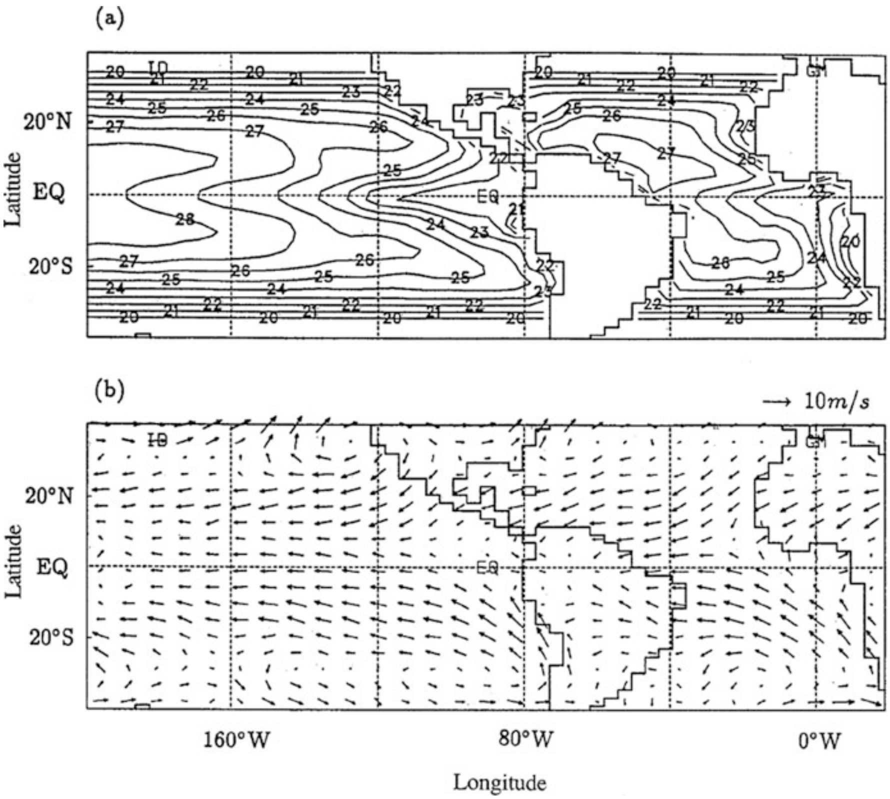

Fig. 2.7 The simulated (**a**) SST (°C) and (**b**) surface wind fields from a hybrid coupled GCM. Only the dynamic coupling is considered in this case. The thermodynamic coupling, such as the surface evaporation and cloud effects, is intentionally suppressed. The model starts from a symmetric SST condition as shown in Fig. 2.6a (From Li (1997). © Copyright 1997 American Meteorological Society (AMS))

of the asymmetric SST pattern is northward cross-equatorial flow in the eastern tropical Pacific (Fig. 2.7b).

The dynamic coupling in this model setting contains two important, distinctive processes: (1) a coastal wind-upwelling mechanism that perturbs SST in the vicinity of the coastal regions and is responsible for the initiation of an equatorial asymmetry and (2) the meridional wind–SST feedback that further amplifies the asymmetry through a positive feedback loop. The former is a coastal mode that primarily involves air–sea interactions in the vicinity of the coast, and the latter is an equatorially trapped mode whose meridional extent is determined by the Rossby radius of deformation.

The importance of the tilted coast in initiating the climate asymmetry in the Pacific was further demonstrated by Philander et al. (1996), who specified an idealized western coast of the Americas that is parallel to a longitude. The idealized

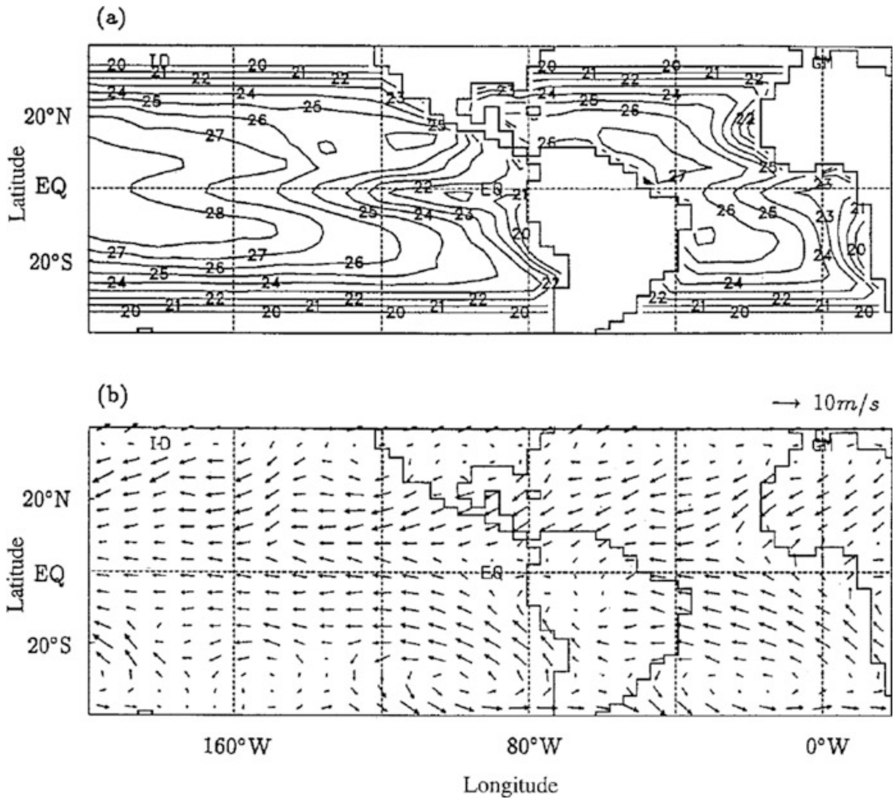

Fig. 2.8 The simulated (**a**) SST (°C) and (**b**) surface wind fields from the hybrid coupled GCM in the presence of the dynamic coupling, the evaporation–wind feedback, and the low-level stratus cloud–SST feedback (From Li (1997). © Copyright 1997 American Meteorological Society (AMS))

numerical model experiment showed that in the absence of the coast asymmetry, no equatorial asymmetric SST patterns are generated in the eastern Pacific.

The intensity of the equatorial asymmetry is further strengthened when the "wind–evaporation–SST" feedback and the "stratus cloud–radiation–SST" feedback are turned on in the coupled experiments, in addition to the dynamic air–sea feedback, as shown in Fig. 2.8.

To sum up, the coupled atmosphere–ocean GCM experiments demonstrated that the essential cause of the preferred NH ITCZ location differs in the tropical Pacific and Atlantic. In the Atlantic, it is primarily attributed to a thermal contrast between the heated northwestern Africa and the cooler ocean to its south. In the Pacific, it is due to the coastal asymmetry of America. The specific coastal and land asymmetries favor a warmer SST and enhanced convection north of the equator and northward cross-equatorial flow. The atmosphere–ocean feedbacks further magnify the coast- and land-induced asymmetry.

While the preferred NH location for ITCZ is attributed to the aforementioned coastal and land asymmetry, the preferred longitudinal location of the ITCZ asymmetry is mainly determined by the depth of climatological mean thermocline at the equator. This is because air–sea interactions are most active in the region where mean thermocline is shallow. The contrasting mean conditions between west and east of the ocean basins are ultimately caused by the dominance of trade winds in both the tropical Pacific and Atlantic. The wind in the tropical Indian Ocean, on the other hand, is unique as it is dominated by north–south monsoonal flows. Thus the ultimate cause of longitudinal ITCZ location is the global distribution of land and ocean that determines where the monsoonal flow and trades prevail and thus determines the east–west tilting of equatorial ocean thermocline.

2.4 Annual Cycle at the Equator

Various factors may contribute to the annual periodicity of SST and wind fields at the equator. Firstly, the solar radiation at the top of the atmosphere has an annual harmonic, with both a symmetric and an antisymmetric component. The former is associated with the elliptic orbit of the Earth – the Sun is closer to the Earth in January and further from the Earth in July. The latter is associated with the tilting of the Earth's axis, which causes the season. But this antisymmetric component has a zero amplitude right on the equator. Which of the above solar radiation components is critical?

Secondly, the SST and wind annual cycles occur only in the equatorial eastern Pacific and Atlantic, where the mean climate is asymmetrically relative to the equator. What is the role of the asymmetric mean state in causing the annual variation of atmospheric and oceanic variables?

To address these questions, Li and Philander (1996) constructed a simple atmosphere–ocean coupled model in which the annual mean state is prescribed and the model is forced by time-dependent solar radiation at the top of the atmosphere.

The atmospheric component of this coupled model is the Lindzen-Nigam model:

$$EV + \beta y U = \frac{\partial \phi}{\partial y} + A\frac{\partial T}{\partial y}, \tag{2.7}$$

$$EU - \beta y V = \frac{\partial \phi}{\partial x} + A\frac{\partial T}{\partial x}, \tag{2.8}$$

$$\varepsilon\phi = -C_0^2\left(\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y}\right), \tag{2.9}$$

where U and V are zonal and meridional wind components; φ is the geopotential at the top of the atmospheric boundary; E is a boundary-layer frictional coefficient; ε is an inverse timescale for cumulus convection adjustment; $A = gH_0/2T_0$ and

$C_0^2 = gH_0$, respectively, measure the strength of the pressure gradient force induced by SST gradients and a barotropic gravity wave speed in the boundary layer; H_0 is the depth of the boundary layer; and T_0 is a reference temperature. In the present study, we set $E = 1/2.5\ \text{day}^{-1}$, $\varepsilon = 1/30\ \text{min}^{-1}$, $H_0 = 3000$ m, and $T_0 = 288$ K.

The ocean component of the coupled model is similar to that of Zebiak and Cane (1987) with slight modification in the shear current equation and the SST equation:

$$\frac{\partial \mathbf{v}}{\partial t} + f\mathbf{k}\times\mathbf{v} = -g'\nabla h + \frac{\tau}{\rho H} - r\mathbf{v} + \upsilon\nabla^2\mathbf{v}, \tag{2.10}$$

$$\frac{\partial h}{\partial t} + H\nabla\cdot\mathbf{v} = -rh + \kappa\nabla^2 h, \tag{2.11}$$

$$\frac{\partial \tilde{\mathbf{v}}}{\partial t} + f\mathbf{k}\times\tilde{\mathbf{v}} = \frac{\tau}{\rho H_1} - r_s\tilde{\mathbf{v}} + \upsilon\nabla^2\tilde{\mathbf{v}}, \tag{2.12}$$

$$\begin{aligned}&\frac{\partial T}{\partial t} + \mathbf{v_1}\cdot\nabla(\bar{T}+T) + \bar{\mathbf{v}}_\mathbf{1}\cdot\nabla T\\ &= -[M(\bar{w}+w) - M(\bar{w})]\bar{T}_z - M(\bar{w}+w)T_z + \frac{Q}{\rho C_w H^w 1} - \alpha T + \kappa\nabla^2 T,\end{aligned} \tag{2.13}$$

where $\mathbf{v}$ and $\tilde{\mathbf{v}}$ denote the mean upper-ocean current and the shear current between the mixed layer and the layer below;

$$\mathbf{v_1} = \mathbf{v} + \frac{H_2}{H}\tilde{\mathbf{v}}$$

represents the surface current; $H_1 = 25$ m and $H = 150$ m are mean mixed layer, thermocline depths, and $H_2 = H - H_1$; $w = H_1$ is the vertical entrainment velocity at the base of the mixed layer; $M(x) = x$, if $x > 0$, and $M(x) = 0$, otherwise, is the Heaviside function; h denotes the varying thermocline; T is sea surface temperature; T_z is the vertical temperature gradient between surface and subsurface; Q denotes surface heat fluxes associated with solar radiation and latent heat fluxes; g' and ρ represent reduced gravity and the density of the upper ocean; r, r_s, and α represent damping coefficients for momentum and heat and have the values $r = 1/150\,\text{day}^{-1}$, $r_s = 1\ \ \text{day}^{-1}$; and $\alpha = 1/60\,\text{day}^{-1}$ and υ and κ stand for diffusion coefficients ($3\times10^3\text{m}^2\text{s}^{-1}$). All quantities with a bar denote the annual mean fields, and the others denote departures from the mean.

The model was forced by seasonally varying solar radiation forcing at the top of the atmosphere, which contains three components: (1) a semiannual harmonic component associated with the meridional movement of the sun that crosses the equator twice a year, (2) an annual harmonic symmetric component that is associated with elliptic orbit, and (3) an annual harmonic antisymmetric component that is associated with tilting of Earth's axis. In the presence of all the solar forcing and the realistic mean state, the model is able to reproduce the observed annual cycles of SST in the tropical Pacific. The simulated SST in the equatorial western Pacific is

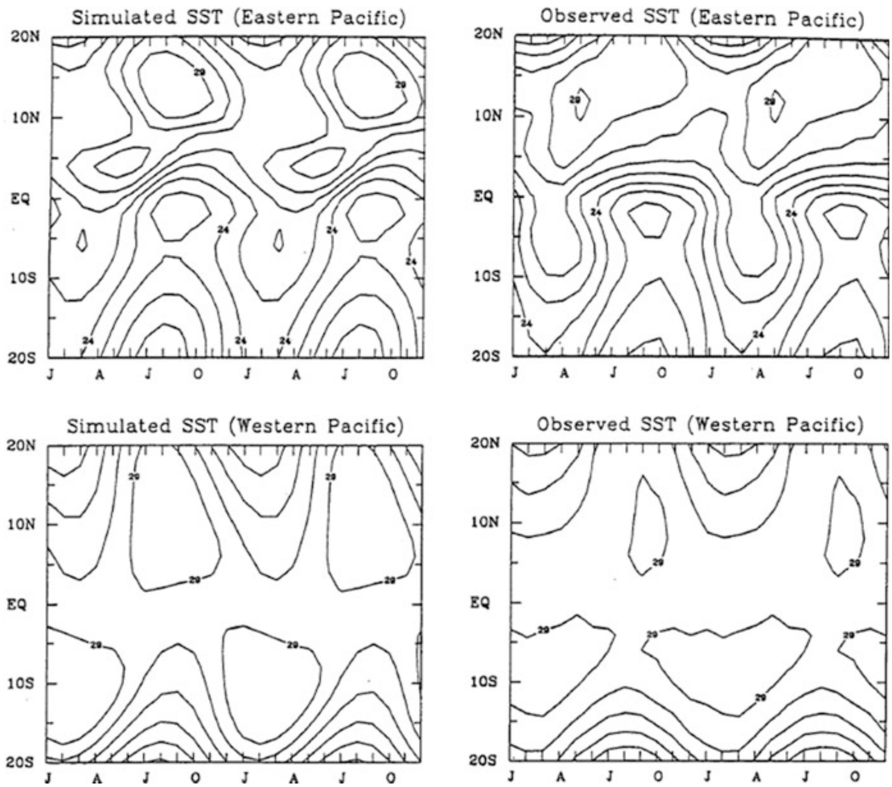

Fig. 2.9 The annual cycles of sea surface temperatures, simulated and observed, in the eastern Pacific (90°–110°W; *upper panels*) and in the western Pacific (160°E–180; *lower panels*). The model results are from the reference case in which the model is forced with the complete seasonal solar radiation; a realistic, asymmetric annual mean basic state is specified and the western coast of the Americas coincides with a meridian (From Li and Philander (1996). © Copyright 1996 American Meteorological Society (AMS))

characterized by a 6-month period, while the SST in the equatorial eastern Pacific is characterized by a 12-month period (Fig. 2.9), consistent with the observation.

Given that the model control experiment is able to reproduce the observed annual cycle in the eastern equatorial Pacific, sensitivity experiments are further carried out to understand the effect of the asymmetric mean state and the relative roles of various components of solar forcing. When a symmetric mean state is specified, no annual cycle is found in the eastern equatorial Pacific. This confirms the important role of the mean state asymmetry in causing the annual cycle. The specific roles through which the mean state impacts the annual cycle will be discussed below. As to the various components of solar forcing, it is found that the symmetric annual harmonic (related to Earth's elliptic orbit) is not important so is the semiannual harmonic. The most important forcing is the one related to the tilt of Earth's axis. Although being zero at the equator, it is critical to cause the annual

SST cycle right on the equator. This is because this antisymmetric annual-harmonic solar forcing results in an antisymmetric SST anomaly, which further drives cross-equatorial winds, leading to the SST change at the equator through anomalous temperature advection and heat fluxes.

The top panel of Fig. 2.10 shows the time–latitude section of the simulated SST field when one considers only dynamic air–sea coupling (i.e., heat flux effects including surface latent heat flux and cloud shortwave radiation variations are intentionally suppressed). There is an annual cycle in SST at the equator. The key process to cause the annual SST variation at the equator in this case is anomalous meridional temperature advection as shown in the bottom panel of Fig. 2.10. Note that the mean meridional temperature gradient $\bar{T}_y$ is positive in the eastern equatorial Pacific. Thus, the anomalous ocean current, v', which changes direction seasonally, cools the equator during the northern summer and warms it during the northern winter. In other words, the term $-v'\bar{T}_y$ leads to an annual SST cycle at the equator. The term $-\bar{v}T'_y$ also contributes to an annual SST cycle because the mean ocean current $\bar{v}$ is northward (in response to the mean northward cross-equatorial wind) and because anomalous meridional temperature gradient T'_y changes sign with the seasons in response to the seasonal change of anomalous cross-equatorial wind according to the meridional wind–upwelling–SST feedback mechanism (Chang and Philander 1994). The above anomalous meridional advection processes become important only when the asymmetric mean state is presented. In the case when the mean state is symmetric about the equator, both $\bar{T}_y$ and $\bar{v}$ vanish at the equator.

The results in Fig. 2.10 show the seasonal cycle that involves the strictly dynamical response of the ocean to the winds. There will be an annual cycle at the equator even if the winds generate no currents but merely cause evaporation, provided the time–mean state has northward winds at the equator. This is possible because the intensification of those winds during the northern summer and their weakening during the northern winter will cause an annual variation in surface evaporation and hence in SST. The dotted line in Fig. 2.11 depicts that variation on the equator at 100° W. It is modest in amplitude but is magnified if we take into account the combined effects of the surface evaporation and the dynamical response discussed in Fig. 2.10. The solid line of Fig. 2.11 shows this combined effect. The further amplification of the annual harmonic is possible because of the low-level cloud–radiation–SST feedback mentioned previously. The lower the sea surface temperatures, the thicker the cloud layer and the smaller the shortwave radiation into the ocean. The dashed line in Fig. 2.11 shows how SST varies when the dynamical response, the surface evaporation, and the stratus cloud feedbacks are all included. This case almost recovers the seasonal SST variation in the control experiment, which is close to the observed amplitude.

To sum up, the eastern equatorial Pacific has a pronounced annual cycle in fields such as SST, both components of the surface winds, and cloudiness. The challenge is to explain how the seasonal variations in solar radiation force such a response. The results described above indicate that the annual cycle is forced by the

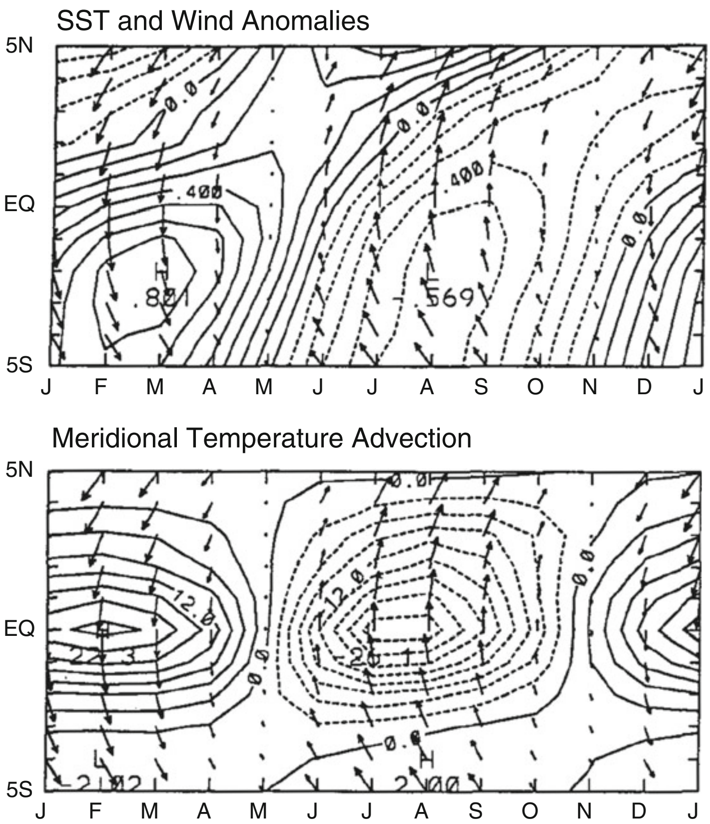

Fig. 2.10 Time–latitude section of the simulated SST field (*top*, interval, 0.1 °C) and anomalous meridional advection (bottom, interval, 3e^{-8} K/s) at 100°W in the presence of dynamic air–sea coupling only (From Li and Philander (1996). © Copyright 1996 American Meteorological Society (AMS))

antisymmetric component of the seasonally varying solar radiation. This is possible because of the mean state asymmetry, an asymmetry that is prominent in the eastern tropical Pacific but not the Indian Ocean or western tropical Pacific. That asymmetry permits an annual cycle on the equator even in a model that couples the atmosphere to a one-dimensional mixed-layer ocean. In such a coupled model, the northward winds at the equator will be intense toward the end of the northern summer and relaxed toward the end of the southern summer. Surface evaporation associated with these winds will cause SST near the equator to be low in August and September and high in March and April. In other words, the model will have an annual cycle in SST at the equator. The amplitude will be modest but can be augmented by next taking into account the dynamical response of the ocean to the winds: intense northward winds induce upwelling and low SST at and to the south of the equator. Further amplification is possible by including the effects of low-level stratus clouds. They form when cold surface waters lead to an

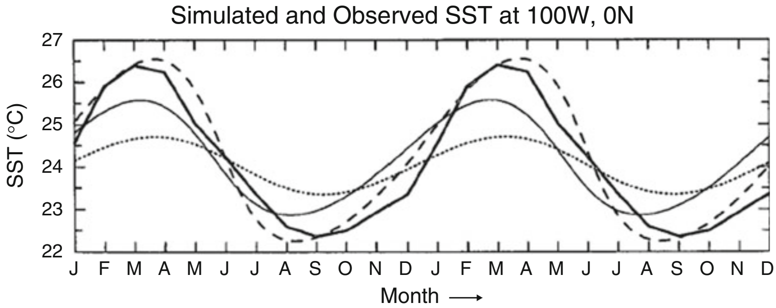

Fig. 2.11 Seasonal variations of SST at 0°, 100°W in three cases. The SST is determined strictly by evaporation in the case of the *dotted line*, by evaporation and oceanic upwelling in the case of the *solid line*, and by evaporation, upwelling, and the presence of low-level stratus clouds in the case of the dashed line. The heavy *black line* corresponds to the observed variations. In all three cases, a realistic time–mean state that is asymmetric about the equator is specified; the model is forced by part of annually varying solar radiation that is strictly antisymmetric about the equator; and the western coast of the Americas coincides with a meridian (From Li and Philander (1996). © Copyright 1996 American Meteorological Society (AMS))

atmospheric inversion and strengthen that inversion by further cooling the surface waters.

In addition to the asymmetric mean state, there are several other factors that contribute to the annual cycle at the equator. They include the annual cycle in solar radiation associated with the ellipticity of the Earth's orbit and the asymmetry, relative to the equator, of the western coast line of the Americas. Their contributions are modest and remain so even when ocean–atmosphere interactions and feedbacks associated with low-level clouds are taken into account.

Questions

1. Describe what unstable positive air–sea feedback processes are involved in the antisymmetric mode development in Chang and Philander (1994).
2. Under what conditions do the positive feedback processes mentioned in Problem 1 work?
3. Describe what unstable positive air–sea feedback processes are involved in the symmetric mode development in Chang and Philander (1994).
4. Under what conditions do the positive feedback processes mentioned in Problem 3 work?
5. In addition to dynamic air–sea feedback mentioned in Problem 1, what types of thermodynamic air–sea feedback processes are involved in the development of the antisymmetric mode (relative to the equator)?
6. Why do unstable dynamic and thermodynamic air–sea feedback processes mentioned above occur only in the tropical eastern Pacific and the tropical eastern Atlantic?
7. Air–sea feedback processes mentioned above favor the occurrence of ITCZ in either hemisphere. Why does the observed ITCZ always stay north of the

equator? What control the preferred ITCZ location north of the equator in the tropical Pacific and Atlantic?

8. Why do the annual variations of observed SST and wind move westward along the equator? What air–sea interaction processes are involved in the westward propagation?
9. Why do observed SST and wind fields have a dominant annual (12-month) period in the equatorial eastern Pacific while the sun crosses the equator twice a year? What physical processes are involved in generating the annual cycle?
10. In Problem 9, what happens if the time–mean climate state is symmetric about the equator in the tropical eastern Pacific?

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