Chapter 4

THE SHAPE OF CONTINENTS, AIR-SEA INTERACTION, AND THE RISING BRANCH OF THE HADLEY CIRCULATION

Shang-Ping Xie
International Pacific Research Center and Department of Meteorology, University of Hawaii, Honolulu, Hawaii 96822, U.S.A.
xie@hawaii.edu

Abstract This chapter begins with a brief history of Intertropical Convergence Zone (ITCZ) research. It then goes on to summarize recent progress in understanding why the ITCZ is locked in the Northern Hemisphere in the eastern Pacific and Atlantic Oceans, and how this northward-displaced ITCZ affects the space-time structure of tropical climate variability.

1. INTRODUCTION

The differential solar radiation in the meridional direction is the ultimate drive for the global Hadley circulation, dictating that its rising branch and heavy rainfall should be located near the equator. This solar forcing of the atmosphere is indirect, however, since most absorption of solar radiation takes place at the surface of earth. Over the tropical oceans, most of the absorbed solar energy is used for surface evaporation and the resultant water vapor is gathered by winds to fuel deep convection that is organized into zonally oriented rain bands. The ocean's effect on tropical convection and hence the rising branch of the Hadley circulation is obvious; tropical rain belts are anchored on the warmest waters, with spatial patterns that can

H.F. Diaz and R.S. Bradley (eds.), The Hadley Circulation: Present, Past and Future, 121–152.

markedly deviate from the distribution of insolation. In particular, the rain band over the eastern Pacific and Atlantic Oceans, called the Intertropical Convergence Zone (ITCZ), is mysteriously displaced to the north of the equator in the annual-mean climatology, a distribution inexplicable from solar forcing alone[1].

This chapter reviews the progress made in the past decade in understanding the coupled ocean-atmospheric dynamics that govern the rising branch of the Hadley circulation and places this progress in a historical perspective. This chapter focuses on the ITCZ over the eastern Pacific and Atlantic, while Webster (Chapter 1, "The Elementary Hadley Circulation," this volume) discusses convection in the Indo-western Pacific sector. Wang et al. (in press) is a global survey of air-sea interaction and its role in climate variability, including a comparative view for the three tropical oceans.

The rest of the chapter is organized as follows. Sections 2 and 3 give historical and observational background, respectively. Section 4 investigates ocean-atmosphere interactions that maintain the climatic asymmetry of the northward-displaced ITCZ, and Section 5 considers the effect of land-sea distribution. Section 6 discusses the climatic consequence of the northward-displaced ITCZ. Following a discussion of some remaining issues in Section 7, Section 8 summarizes the main results.

2. HISTORY OF THE STUDY OF TROPICAL WINDS AND RAINS

> "It is not the work of one, nor of few, but of a multitude of Observers, to bring together the experience required to compose a perfect and complete History of these Winds."
>
> Edmond Halley (1686)

Before the invention of steam engines, knowledge of the direction, speed and steadiness of sea surface winds was of vital importance for the navigation of sailing boats. By the late seventeenth century, the traffic between Europe and the New World had grown to such a level that Halley (1686) was able to compile a quite accurate map of surface-wind streamlines for the tropical Atlantic and Indian Oceans by gathering information from navigators. Figure 4-1 reproduces the Atlantic portion of Halley's wind map that depicts the steady trade winds in the Northern and Southern Hemi-

---

[1] The latitude of the sinking branch of the Hadley circulation is not directly determined by solar radiation either. Instead, it is determined by dynamic requirements like angular momentum conservation and baroclinic instability (Held and Hou 1980; Lindzen and Hou 1988).

spheres. Remarkably, the southeasterly and northeasterly trades meet north of, instead of on, the equator as one might expect from equatorial symmetry. The ITCZ—the modern term for the region where the trade winds meet—is displaced to the Northern Hemisphere in the annual mean. Halley wrote about the ITCZ: "it were improper to say there is any Trade Winds, or yet a Variable; for it seems condemned to perpetual Calms, attended with terrible Thunder and Lightning, and Rains so frequent, that our Navigators from thence call this part of the Sea the *Rains*".

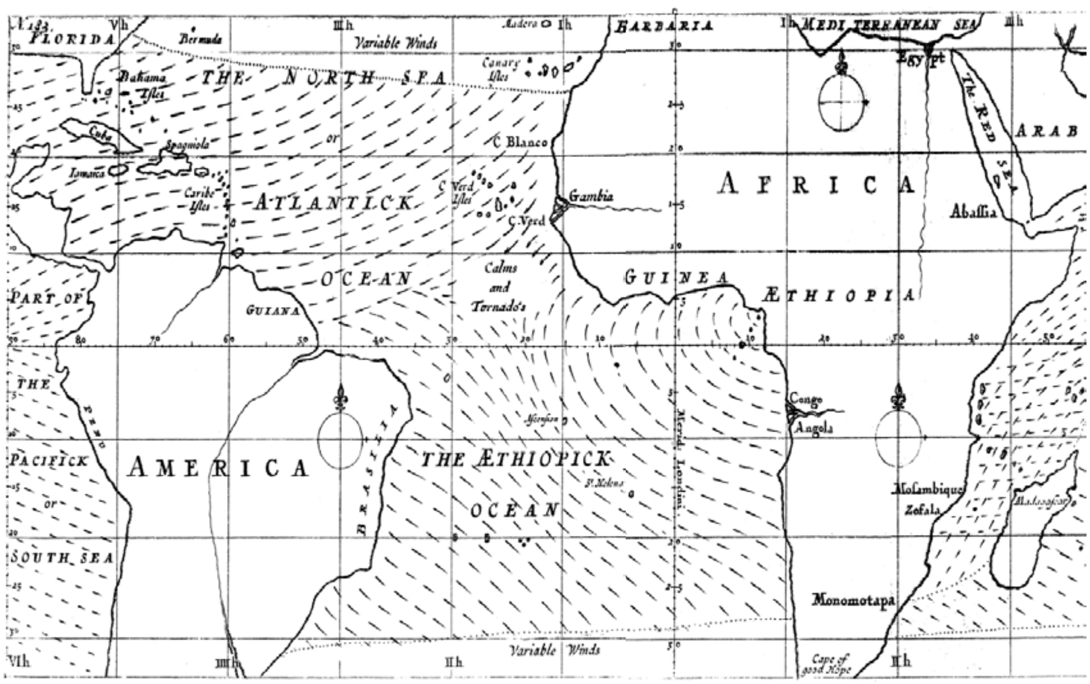

Figure 4-1. Halley's (1686) map of surface wind streamlines. The southeasterly trade winds are shown to converge onto the Northern Hemisphere.

In the ITCZ, surface air rises and in the process, the water vapor it carries condenses, resulting in the frequent rains and thunderstorms Halley noted and releasing a huge amount of latent heat that drives the Hadley and global circulation of the troposphere. Hereafter we will use the terms ITCZ,[2]

[2] More precisely, our definition of ITCZ refers to those surface convergence zones over warm oceans with sea surface temperatures (SSTs) greater than the convective threshold (26–27°C). There are surface convergence zones over cool sea surface that are not associated with deep convection and significant precipitation. Instead, they are associated with shallow boundary-layer circulation.

convective zone, and precipitation band interchangeably. The ITCZ resides in a zone of "perpetual calms" in Halley's words, and is now called the Equatorial Doldrums in textbooks. As will become clear in Section 4.1, the collocation of the Doldrums with the ITCZ is the key to the mystery of their northward displacement from the equator.

Before the late seventeenth century, the vast Pacific Ocean was much less navigated than the Atlantic and Halley had little information on its wind distribution other than accounts that "there is great conformity between the winds of this Sea and those of the Atlantic." This lack of observations forced Halley to draw an "analogy between" the Pacific winds "and those of the Atlantic." Interestingly, Halley did not draw a perfect analogy with the Atlantic winds; in his map, the Pacific trades converge on the geographical equator, rather than on northern latitudes as in the Atlantic. Perhaps Halley or his contemporaries had no reason to believe that the Pacific wind system should depart from equatorial symmetry. By the late nineteenth century, Köppen's (1899) atlas showed that the similarity between the Pacific and Atlantic is greater than Halley thought; as in the Atlantic, the Pacific trades also converge onto the Northern Hemisphere even in boreal winter.

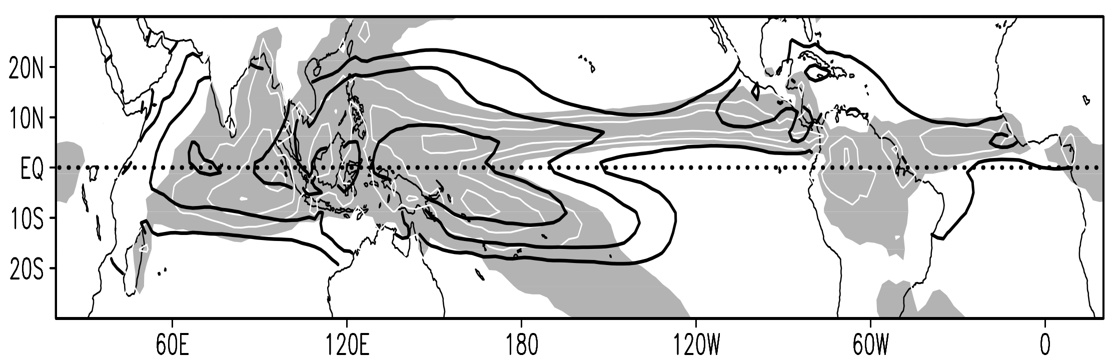

Figure 4-2. Annual-mean climatological precipitation (white contours at 2 mm/day intervals; shade > 4 mm/day), and SST (black contours at 1°C intervals; only contours for 27°C and above are plotted), based on the Climate Prediction Center Merged Analysis of Precipitation (CMAP; Xie and Arkin 1996) and the Reynolds and Smith (1994) data set, respectively.

A reliable precipitation climatology proves more difficult to obtain because of the sporadic nature of rains. In Bartholomew and Herbertson's (1899) map of annual rainfall, the Pacific Ocean was left blank. (One can nevertheless infer a strong equatorial asymmetry from the depicted rainfall on the Pacific coast that is over 160 inches/year in Colombia north of the equator but less than 10 inches/year on the Peruvian coast.) For the mid-twentieth century, Möller's (1951) map of annual mean rainfall is very similar to modern climatology (Fig. 4-2), showing that the ITCZ rain band is clearly displaced to the Northern Hemisphere over both the eastern Pacific and the Atlantic.

3. OBSERVATIONAL BACKGROUND

The advent of satellite remote sensing opened the door for global observations of clouds in the 1960s and, somewhat later, for observations of precipitation. In an early climatology of reflectivity (U.S. Air Force and U.S. Department of Commerce 1971), the Pacific and Atlantic ITCZ appears on the dark ocean background as a silver belt that is north of the equator in both boreal summer and winter and one of the most visible and striking features in such satellite images. Since 1979, outgoing long-wave radiation (OLR) measurements by satellite infrared sensors are often used as a proxy of precipitating deep convection that reaches great heights. A paradox arises: Over the eastern Pacific and Atlantic, the OLR-based estimate of rainfall is too low compared to ship reports, which indicate substantial precipitation accompanied by strong surface wind convergence there (Fig. 4-3). It turns out that this underestimation by the OLR-based method in the eastern Pacific and Atlantic ITCZ is due to the fact that the sea surface temperature (SST ~27°C) there is significantly lower than it is in the Indo-western Pacific warm pool (SST > 28°C). As a result, convection in the eastern Pacific and Atlantic does not reach as high as in the western Pacific, yielding higher OLR values (Thompson et al. 1979; J.M. Wallace 1994, personal communication.). More recent satellite microwave sensors, measuring quantities more directly related to precipitation than the infrared ones, observe similar rain rates in the eastern and in the western Pacific.

Figure 4-2 shows the annual mean precipitation climatology based on combined infrared and microwave satellite observations. Over the continents and the Indo-western Pacific sector, the annual mean precipitation distribution in the tropics is more or less symmetric about the equator, consistent with solar radiation distribution. On the seasonal time scale, the maximum rainfall in these regions moves back and forth across the equator following the sun (Mitchell and Wallace 1992). This solar control of tropical convection breaks down over the eastern half of the Pacific and entire Atlantic, where deep convection is confined to the ITCZ north of the equator. This climatic asymmetry persists for most of the year, even during boreal winter when the solar radiation south of the equator exceeds that to the north

(Fig. 4-4). Only for a brief period during March and April, a double ITCZ appears with a rain band on each side of the equator.

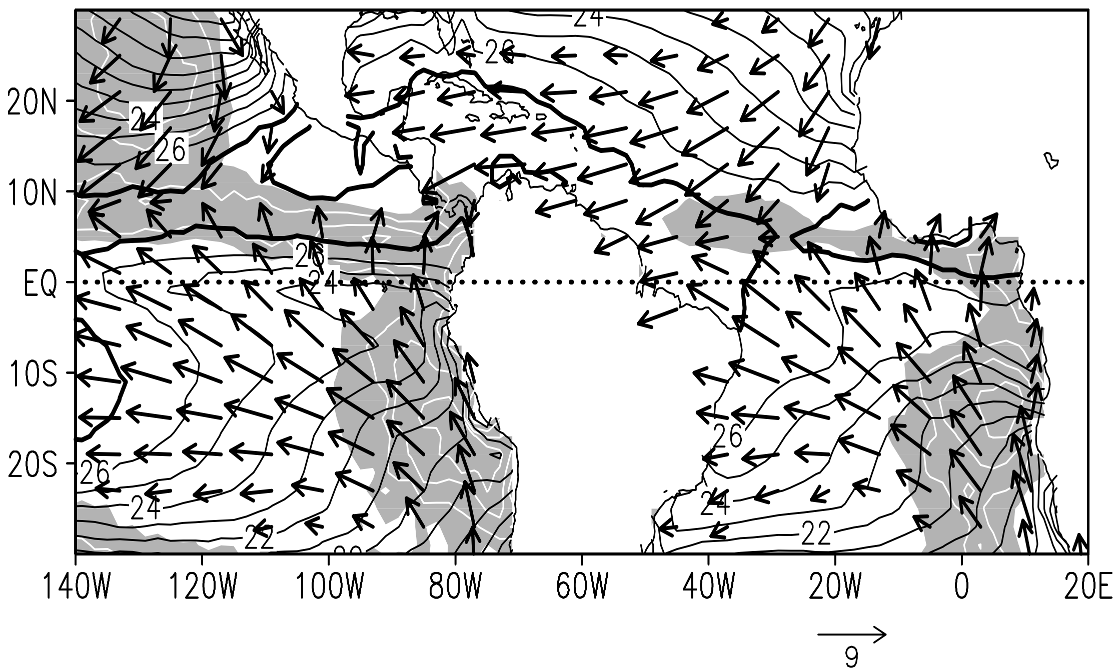

Figure 4-3. Climatological SST (contours in °C), surface wind vectors (m/s), and cloud cover (white contours at 5% intervals; shade > 60%), based on the Comprehensive Ocean-Atmospheric Data Set (COADS; Woodruff et al. 1987).

Located in the region where a great amount of latent heat is released to the atmosphere, the ITCZ is sometimes called the thermal equator. The peculiar location of the thermal equator in the eastern Pacific and Atlantic begs answers to the following questions. (1) Why is the ITCZ not on the equator where the annual mean solar radiation is the maximum? (2) Given that annual mean solar radiation is roughly symmetric about the equator, why is the ITCZ displaced north of the equator? and (3) What effect does this northward displacement of the thermal equator have on climate variability? Two schools of thought exist regarding the first two questions. One points to the strong hemispheric asymmetry in the landmass and its distribution and suggests that this continental asymmetry causes climatic asymmetry. The other camp proposes an SST control, countering the former with the fact that climatic asymmetry is weak in the Indian Ocean, yet the region has the greatest interhemispheric distribution in landmass. We will show that these schools of thought are not mutually exclusive and both are necessary for the complete answers. Let us first look at the arguments of SST control.

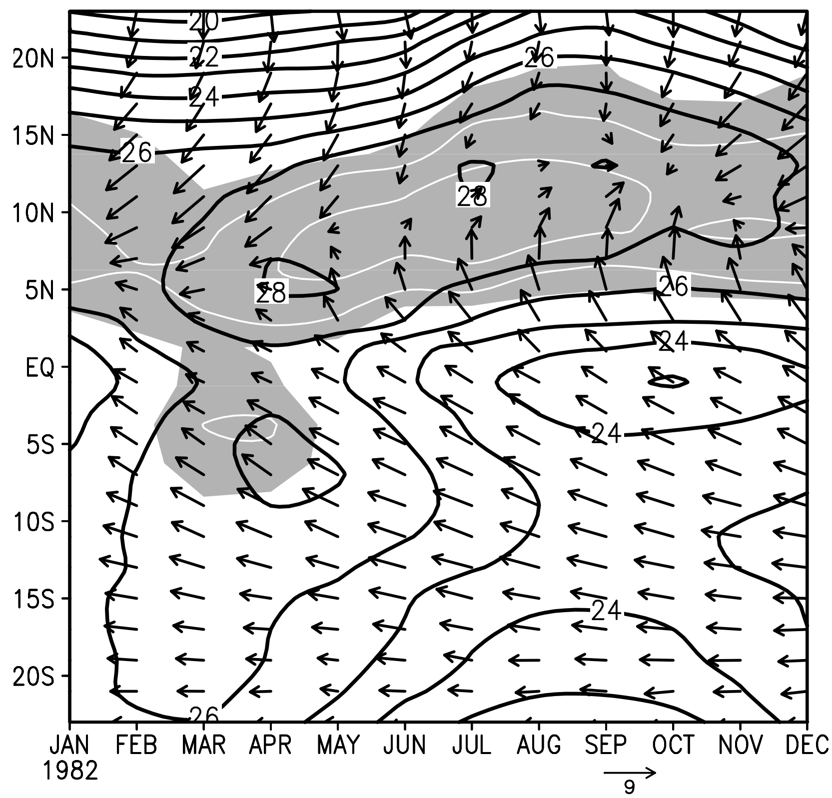

Figure 4-4. Time-latitude section of climatological SST (black contours in °C), surface wind vectors (m/s), based on COADS; and CMAP precipitation (white contours at 5 mm/d intervals; shade > 2.5 mm/d). All zonally averaged in 120°W–115°W.

Sea surface temperature affects tropical convection through its effect on moist static stability and its gradient that drives moisture-laden winds in the marine boundary layer (MBL). An empirical SST threshold for deep convection exists at 26°C–27°C (e.g., Waliser and Graham 1993). Indeed, major tropical precipitation is confined within the 27°C SST contours (Fig. 4-2). The correspondence between SST and tropical convection is not perfect, however. This is especially the case in the Indo-western Pacific region where, for example, SST has a broad equatorial maximum, yet the precipitation maximum occurs off the equator on either side. Atmospheric general circulation model (AGCM) experiments under aqua-planet conditions suggest that depending on the convection scheme used, precipitation shows a single maximum on the equator or a pair of maxima on either side of the equator in response to such a broad equatorial maximum of SST (Numaguti and Hayashi 1991). The physics of such subtlety is not well understood, and it remains unclear how SST affects convection over the Indo-western Pacific warm pool, where the SST gradient is weak.

The SST control over precipitation is much stronger in the eastern Pacific and in the Atlantic, where the SST gradient is strong. This SST control is perhaps best illustrated by its coevolution with convection and surface winds (Fig. 4-4). Most of the time, warm water with SSTs > 26°C is confined to north of the equator, and so is convection. From April to September, the precipitation maximum moves northward from 5°N to 10°N, apparently dragged by the northward shift of the SST maximum. Briefly in March and April, as the equatorial cold tongue relaxes, the meridional SST gradient weakens substantially between 10°S and 15°N, and SST south of the equator rises above the 26°C threshold, reaching as high as 27°C. Over this Southern Hemisphere warm water, considerable precipitation takes place and a double ITCZ symmetric about the equator is often observed during these months[3]. Surface wind convergence follows the same seasonal cycle of and is tightly coupled with SST and precipitation. While some details remain to be worked out, such as the equatorward displacement of the precipitation maximum from the SST maximum in the Northern Hemisphere (Hastenrath 1991), their joint seasonal cycle in Figure 4-4 illustrates the strong SST control of convection in this part of the world.

4. AIR-SEA FEEDBACK

The SST control mechanism offers a partial solution to the problem of climatic asymmetry. From such a meteorological point of view, the ITCZ remains north of the equator over the eastern Pacific and Atlantic because SST is higher north of the equator than south. From an oceanographic point of view, on the other hand, SST is higher north of the equator because the ITCZ stays in the Northern Hemisphere. This circular argument suggests that the northward-displaced ITCZ and high SST band are just two sides of the same coin and understanding both phenomena requires an air-sea interaction approach[4].

---

[3] The discovery of such a double ITCZ from satellite cloud imagery in March 1967 (Kornfield et al. 1967) led to a brief excitement that it vindicates Charney's (1971) then unpublished theory of Ekman CISK (conditional instability of the second kind) that predicts such a double ITCZ. Based on atmospheric GCM results, Manabe et al. (1974) show that SST effect is more important in controlling the eastern Pacific ITCZ.

[4] Pike (1971) used a coupled ocean-atmosphere model to study the meridional configuration of the ITCZ. His results answer the first question in the previous section. Namely, under the prevailing easterlies, wind-induced upwelling reduces SST on the equator to a level that deep convection is no longer possible. At the end of his 88-day integration, a single ITCZ forms away from the cold equator. Curiously, however, SST under the ITCZ is 0.5°C lower than on the other side of the equator, in contrary to the observed SST-precipitation relation (Fig. 4-3).

In recognition of their importance for the El Niño/Southern Oscillation (ENSO) and its global impact, the eastern Pacific ITCZ and equatorial annual cycle were extensively discussed at several of the National Oceanic and Atmospheric Administration's (NOAA) Equatorial Pacific Ocean Climate Studies (EPOCS) Program principal investigators meetings in the early 1990s. Stimulated by these discussions, investigators proposed several air-sea feedback mechanisms for maintaining the observed climatic asymmetry characterized by the northward-displaced ITCZ.

4.1. Wind-Evaporation-SST Feedback

Surface evaporation, a function of both SST and wind speed, is the major means for tropical oceans to balance incoming solar radiation. Surface wind speed reaches a minimum at the ITCZ in both the eastern Pacific and the Atlantic (Figs. 4-3 and 4-4), a fact Halley (1686) documented[5]. Based on this observation, Xie and Philander (1994) propose the following mechanism for breaking the equatorial symmetry set by solar radiation. Suppose that somehow SST north of the equator becomes slightly warmer than to the south (Fig. 4-5). The sea level pressure (SLP) gradient will drive southerly winds across the equator. The Coriolis force acts to turn these southerlies westward south and eastward north of the equator. Superimposed on the background easterly trades south of the equator, these anomalous southeasterlies increase surface wind speed and hence evaporative cooling. Conversely, north of the equator wind speed and surface evaporation decrease, amplifying the initial northward SST gradient. This wind-evaporation-SST (WES) feedback offers an explanation for the observed cross-equatorial differences in both wind speed and SST in Figure 4-3. If one assumes that everything else is the same at 10°N and 10°S, a 25% wind speed difference leads to an SST difference of 3°C according to the Clausius-Clapeyron equation for saturated water vapor content (for a typical wind speed of 7–8 m/s). To balance the net radiative flux, SST must rise (fall) under weak (strong) winds at 10°N (S).

Why Pike's model ITCZ chooses to form in the colder hemisphere is unclear, possibly because the integration is too short to filter out chaotic variability in the tropical convection. Manabe (1969) clearly recognizes the effect of equatorial upwelling on tropical convection, but somehow the oceanic ITCZ stays on the equator in his one-year integration of a coupled GCM.

[5] In the ITCZ, annual mean scalar wind speed is considerably greater than vector speed because of the rectification by westward-traveling easterly waves (e.g., Gu and Zhang 2002) and the seasonal march of the Doldrums (Section 7.1). In the 4-year (August 1999–July 2003) observations by the QuikSCAT satellite scatterometer, the mean scalar wind speed is 6–7 m/s in the ITCZ as opposed to ~8 m/s on the other side of the equator along 10°S.

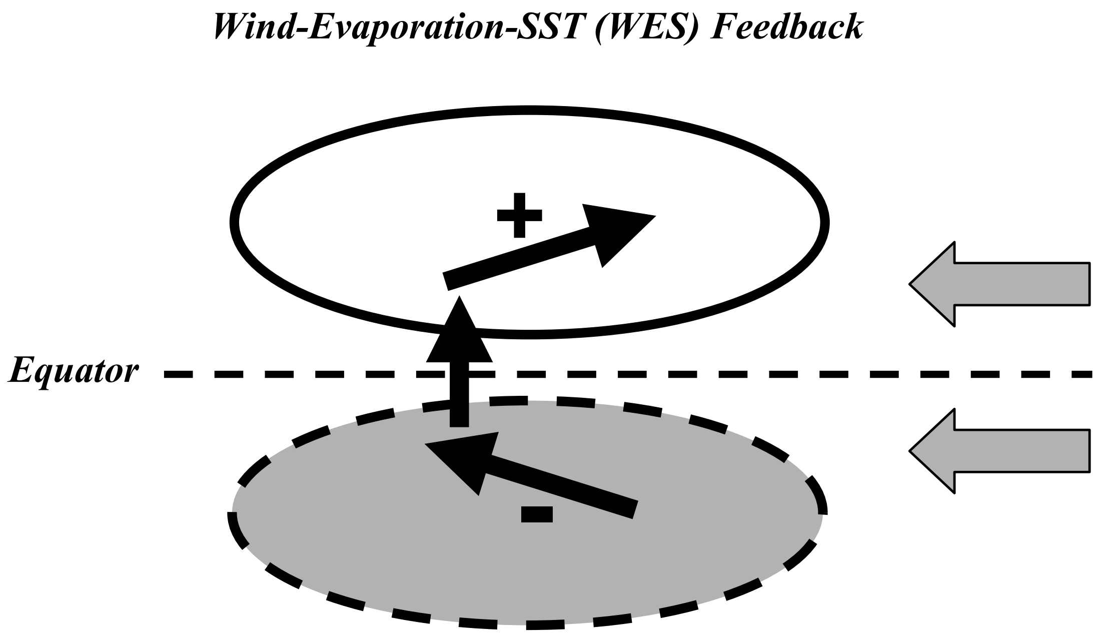

Figure 4-5. Schematic of the WES feedback: anomalies of SST in contours (negative dashed) and surface wind velocity in black vectors. The gray vectors on the right signify the background easterly trades.

Xie and Philander (1994) demonstrate the symmetry-breaking effect of this WES feedback with a zonally symmetric coupled model. The model convection is linearly proportional to SST above a threshold and vanishes when SST falls below it. The surface wind is computed based on a linear model. The model SST is computed based on a slab mixed-layer model cooled by upwelling centered on the equator. Because of the WES feedback, the symmetric solution with a double ITCZ becomes unstable and the model settles into an asymmetric steady state. Under forcing and boundary conditions that are perfectly symmetric, the coupled model develops a single ITCZ on one side of the equator that is collocated with the surface wind speed minimum and SST maximum (Fig. 4-6).

When equatorial upwelling is removed, SST reaches the maximum at the equator and so does the model convection. Only when the equator is kept colder than the convective threshold by ocean upwelling, the coupled system has a choice among the double-ITCZ symmetric solution and two asymmetric ones with a single off-equatorial ITCZ. Because of the WES feedback, the symmetric solution with a double ITCZ is unstable and the model settles into an asymmetric steady state. This necessary condition of equatorial upwelling for the development of climatic asymmetry is consistent with the observation that the ITCZ is nearly symmetric over the Indo-western Pacific warm pool but kept to the north of the equator over the eastern Pacific and the Atlantic where ocean upwelling maintains a cold equator.

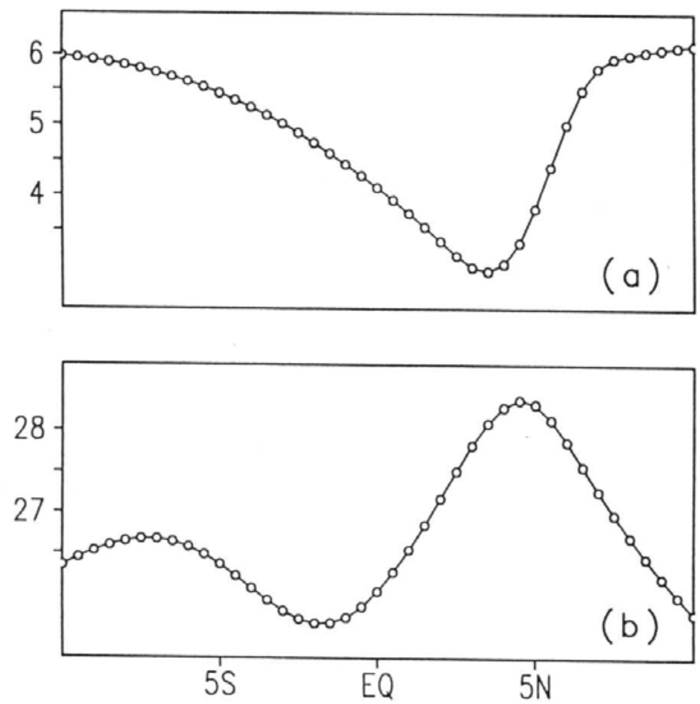

Figure 4-6. Asymmetric solution to the Xie and Philander (1994) model under equatorially symmetric conditions: (a) wind speed and (b) SST.

4.2. Stratus-SST Feedback

While solar radiation at the top of the atmosphere is nearly symmetric about the equator, its distribution at the sea surface is not, because of the reflection by clouds. Figure 4-3 shows the observed cloudiness climatology (shade). The narrow band of convective clouds in the ITCZ is a negative feedback on SST, reducing the climatic asymmetry. Toward the east end of both the Pacific and Atlantic basins, an extensive low cloud deck with annual mean cloud cover exceeding 60% helps cool the southern tropical oceans by reducing insolation at the sea surface. Klein and Hartmann (1993) report that the low-level stratus cloud cover is highly correlated and increases with atmospheric stability above the sea surface. The Peruvian cloud cover peaks in boreal fall when the local air temperature in the lower troposphere is at its minimum. Philander et al. (1996) suggest that this cloud cover peak off Peru results from seasonal cooling of local SST. They further propose a positive feedback between stratus clouds and SST: An initial SST cooling increases the atmospheric stability and hence the stratus cloud cover, which acts to amplify the initial sea surface cooling by reflecting solar radiation into space. This positive stratus-SST correlation is observed in interannual variations over the South Pacific (Klein and Hartmann 1993) and South Atlantic (Tanimoto and Xie 2002) stratus cloud regions. For the South Atlantic cloud deck, Tanimoto and Xie estimate that cloud cover increases by 10% in response to a 1°C drop in local SST, twice as much as Klein and Hartmann's estimate for the cloud cover response near the equatorial southeast Pacific. These different findings presumably arise because the authors focus on different SST and atmospheric circulation anomaly patterns—ENSO is an important player in South Pacific cloud cover variations while the anomalies of South Atlantic stratus clouds are associated with the meridional shift of the Atlantic ITCZ as well as local SST changes on the basin scale.

Using a coupled GCM, Philander et al. (1996) demonstrated the effect of this stratus-SST feedback on climatic asymmetry. They reported that climatic asymmetry is markedly strengthened if a stratus cloud parameterization based on Klein and Hartmann's (1993) observations is implemented, in which the cloud cover increases with static stability of the lower atmosphere. This stratus effect on the eastern Pacific ITCZ is found in other coupled GCMs (Ma et al. 1996; Kimoto and Shen 1997; Yu and Mechoso 1999; Gordon et al. 2000; Fu and Wang 2001).

Philander et al.'s (1996) original feedback concept considers only the indirect effect of stratus clouds on the ITCZ through SST. The intense upward long-wave radiation at the top of these clouds also cools the marine boundary layer, strengthening the South Pacific subtropical high and the cross-equatorial southerly winds that converge on the northern ITCZ (Nigam 1997). This direct effect on the atmosphere is confirmed in a full physics model but the effect is modest; a complete removal of the cloud radiative effect south of the equator leads to a 10%–20% decrease in the intensity of the eastern Pacific ITCZ (Wang et al., submitted).

4.3. Upwelling-SST Feedback

Southerly to southeasterly winds dominate the eastern Pacific and Atlantic equator. In response to these cross-equatorial winds, surface ocean currents flow downwind near the equator but become perpendicular to the wind direction 2°–3° away from the equator as the Coriolis effect becomes important. In response to a southerly wind forcing, this change in flow regime generates ocean upwelling south and downwelling north of the equator. The resultant northward SST gradient strengthens the cross-equatorial winds, completing a positive feedback loop (Chang and Philander 1994). This feedback explains the observation that the center of the equatorial cold tongue is consistently shifted south of the equator in the eastern Pacific and the Atlantic. Since the upwelling effect on SST becomes less important in the off-equatorial open ocean (say poleward of 3°), this mechanism is probably secondary in generating the broader latitudinal asymmetry between 10°S and 10°N.

The narrower meridional asymmetry, as characterized by a strong SST front at 2°N and weak SST gradients at 2°S, is receiving much attention lately. Satellite scatterometer measurements indicate a strong wind deceleration along the axis of the cold tongue and a strong acceleration as the air flows across the equator (Chelton et al. 2001). Attributed to SST-induced adjustment in vertical wind shear (Wallace et al. 1989), this deceleration of wind on the equator leads to strong wind curl that favors upwelling south of the equator (Chelton et al. 2001). It also maintains a surface wind convergence south of the equator (Liu and Xie 2002), which is not associated with deep convection because a strong temperature inversion caps the marine boundary layer except for a brief period during March–April when local SST exceeds 26°C (Fig. 4-4). Quasi-periodic (monthly) tropical instability waves produce spectacular meanders at the equatorial front centered at 1°N–2°N, inducing covariations in the atmospheric boundary layer. In particular, an increase in SST along the front is associated with an increase in boundary-layer cloud cover (Deser et al. 1993; Hashizume et al. 2001), an association opposite to the one observed over the stratus cloud decks west of South America and South Africa. The cross-equatorial flow in the MBL and its interaction with the ITCZ are foci of a recent field campaign over the eastern Pacific (Cronin et al. 2002; Raymond et al. 2004; Small et al. 2005).

5. CONTINENTAL FORCING AND ITS WESTWARD CONTROL

Air-sea feedbacks are important in keeping the ITCZ north of the equator. They do not fully explain, however, why the Northern, not the Southern Hemisphere is favored in the Pacific and Atlantic. A long-held belief is that hemispheric asymmetry in area, shape, and orography of continents ultimately gives rise to the climatic asymmetry. This must be generally true, for the northward-displaced ITCZ has existed for a long time—at least since Europeans began sailing in the Atlantic many hundred years ago. Unanswered until very recently have been the following specific questions. Which continental features and how do they move the ITCZ away from the equator? For the Pacific, is the shape of the Asian-Australian continents to the west or that of the Americas more important? Given that the direct atmospheric response to continental asymmetry is likely to be confined near the coast, what sustains the climatic asymmetry in the middle of the vast Pacific, far away from any continents?

5.1. Westward Control

Xie (1996a) develops a simple theory to address these questions. The nondimensional equation for SST difference between meridional SST maxima north and south of the equator may be cast as

$$\frac{\partial(T_N - T_S)}{\partial t} = \sigma V - (T_N - T_S)\,, \tag{1}$$

where V is the meridional wind velocity on the equator and σ is a positive quantity called the WES coefficient. The right-hand side (rhs) is obtained by linearizing the surface latent heat flux. The first term reflects its wind-speed dependence and represents the WES effect: Southerly cross-equatorial winds will reduce (increase) wind speed and evaporation north (south) of the equator (Fig. 4-5). The second term on the rhs reflects the SST dependence of surface evaporation and acts as Newtonian cooling. The local cloud-SST feedback may be absorbed in the second term. Time has been normalized with the resultant effective Newtonian cooling coefficient.

Cross-equatorial wind velocity is modeled with a quasi-steady Rossby wave equation that is forced by the meridional SST gradient

$$(1 - \frac{\partial}{\partial x})V = (T_N - T_S)\,, \tag{2}$$

where the east-west distance x has been normalized by the e-folding scale of the damped long Rossby wave. Combining (1) and (2) yields an equation for cross-equatorial wind velocity that serves as a measure of climatic asymmetry

$$(\frac{\partial}{\partial t} + 1)(1 - \frac{\partial}{\partial x})V = \sigma V\,. \tag{3}$$

For the axisymmetric case, (3) reduces to

$$\frac{\partial}{\partial t}V = (\sigma - 1)V\,. \tag{4}$$

If the WES effect is strong enough to overcome the Newtonian cooling, the solution to (4) becomes unstable and small latitudinal asymmetry grows into amplitudes large enough to push the ITCZ to one side of the equator and eliminate convection on the other, as reported in Xie and Philander (1994).

In general, Equation (3) may be solved by imposing an eastern boundary condition

$$V\,|_{x=0} = V_E\,. \tag{5}$$

V_E is positive in both the Pacific and Atlantic (Fig. 4-3) and as part of the westward-traveling Rossby wave, results solely from continental forcing to the east. The steady-state solution to (3) and (5) is

$$V = V_E e^{(1-\sigma)x}\,. \tag{6}$$

Namely, the oceanic climate asymmetry is controlled by continental asymmetry on the ocean's eastern boundary. Positive air-sea feedback increases the e-folding zonal scale $(1 - \sigma)^{-1}$, allowing the influence of continental asymmetry to penetrate far into the west, over nearly 10,000 km in the Pacific. Observed meridional wind on the equator peaks on the South American coast and decays westward, indicating that the real Pacific Ocean-atmosphere is subcritical ($\sigma < 1$).

The continent's westward control over oceanic climate stems from the fact that under the long-wave approximation, all asymmetric signals in the ocean and atmosphere have to propagate westward as Rossby waves. (The Kelvin wave can propagate eastward, but is symmetric about the equator.) Figure 4-7 is an explicit demonstration of this westward control in a coupled model in which a northern land bulge creates the latitudinal asymmetry. Basin-wide northward displacement of the ITCZ occurs only when the continental forcing is placed on the eastern continent. This westward co-propagation of ocean-atmospheric anomalies is consistent with the coupled GCM results that a localized radiative cooling off the Peruvian coast strengthens climatic asymmetries across the Pacific (Ma et al. 1996; Kimoto and Shen 1997; Yu and Mechoso 1999). An immediate implication of this westward control mechanism is that the search for the symmetry-breaking forces can be narrowed down to the eastern continent—the Americas for the Pacific and Africa for the Atlantic.

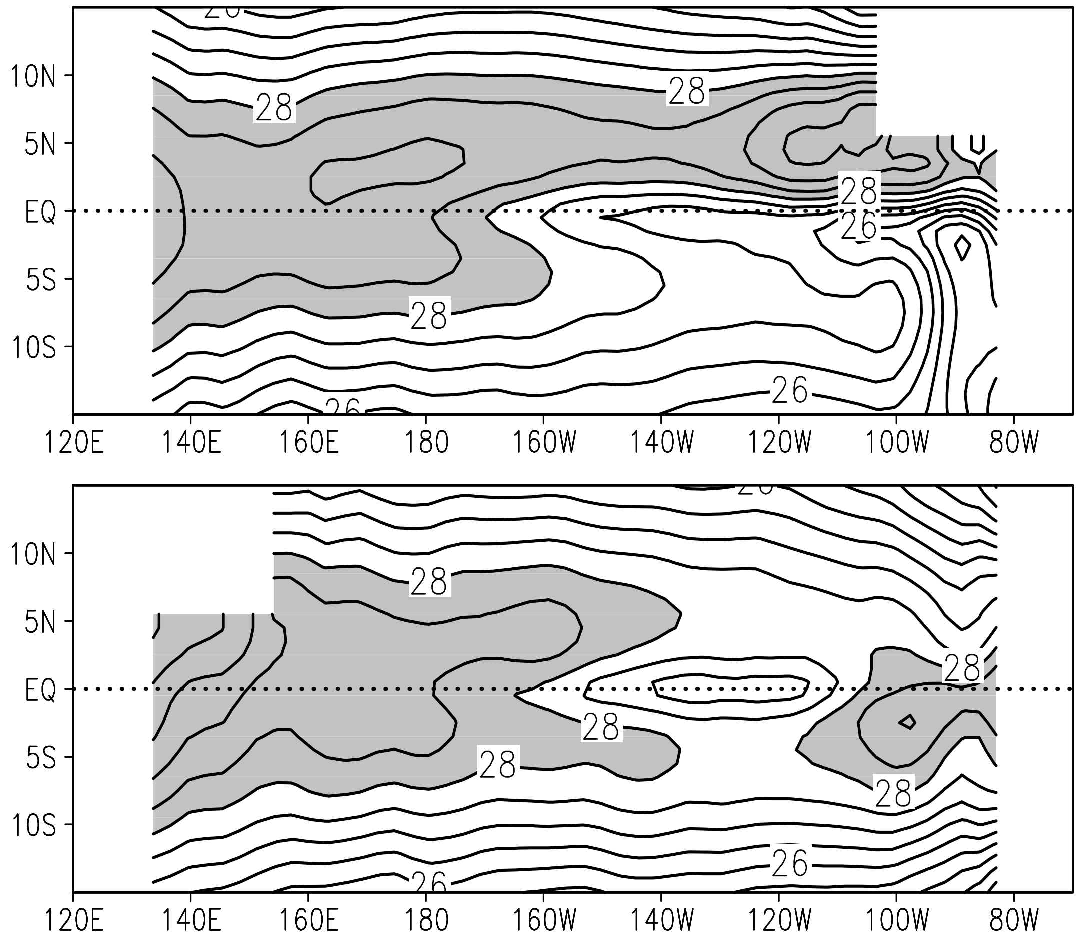

Figure 4-7. Time-mean SST in a coupled model where a northern land bulge is added to the eastern (upper panel) and western (lower panel) continent. From Xie and Saito (2001).

5.2. Continental Forcing and Basin-Wide Adjustment

Using a coupled GCM, Philander et al. (1996) were the first to demonstrate that the shape of the west coast of the Americas and Africa leads to basin-wide displacement of the ITCZ into the Northern Hemisphere in the Pacific and Atlantic, respectively. First, they run their atmospheric GCM under zonally uniform and latitudinally symmetric SST to isolate the effect of continental geometry on ocean winds. Regarding the Pacific, they point to the northwest tilt of the coast of the American continents. South of the equator, winds are roughly parallel to the coast, which would induce coastal upwelling and cool SSTs there (Fig. 4-8 [top]). North of the equator, by contrast, winds are nearly perpendicular to the coast with little effect on upwelling and as a result, SST remains high there. Then based on their coupled model runs, Philander et al. show that though confined near the coast, this initial hemispheric difference in coastal upwelling and SST leads to a basin-wide shift in Pacific climate when air-sea interaction, the stratus-SST feedback in particular, kicks in (Figs. 4-8 [middle], 4-8 [bottom]).

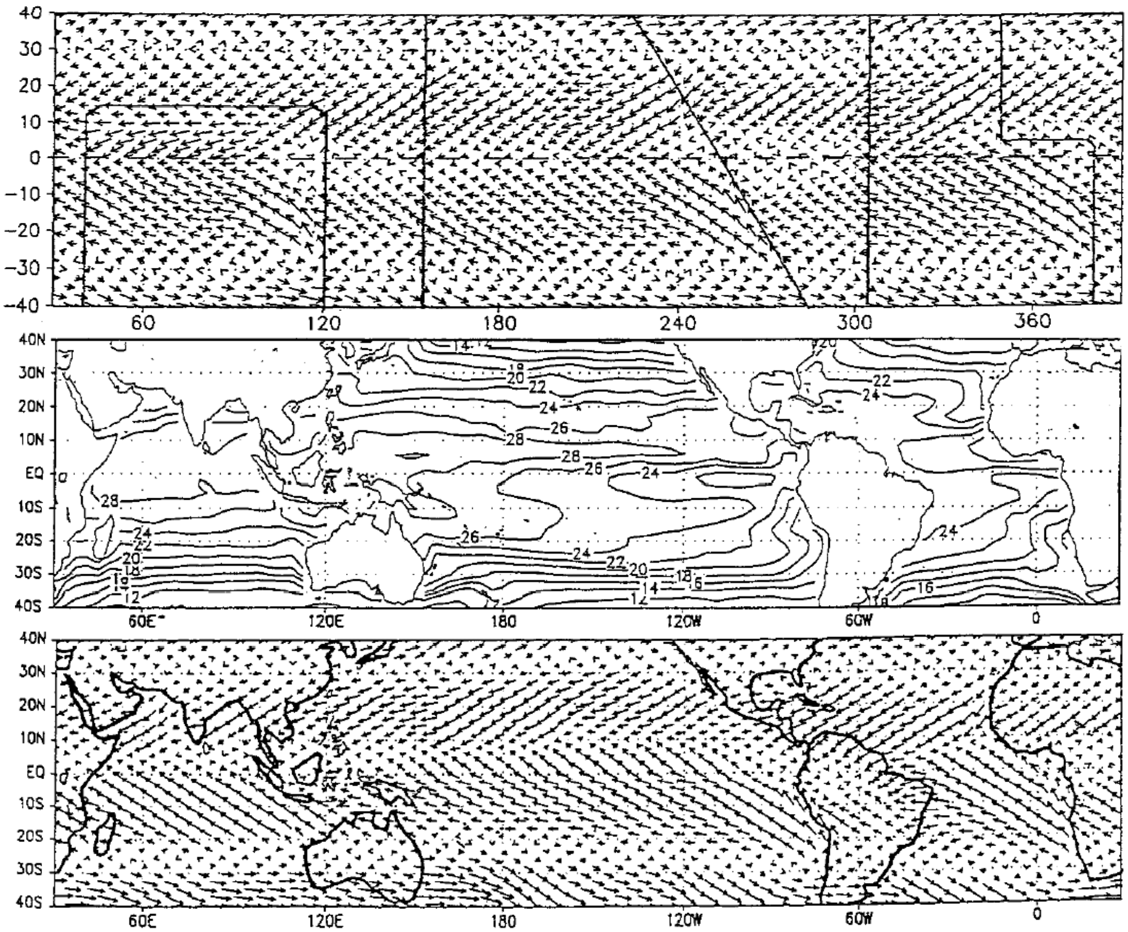

Figure 4-8. (top) Surface wind velocity in an atmospheric GCM forced with a specified SST distribution that is zonally uniform and equatorially symmetric. (middle) SST and (bottom) surface wind velocity in a coupled ocean-atmosphere GCM. The equinox insolation distribution is used and all the orography on land is removed. From Philander et al. (1996).

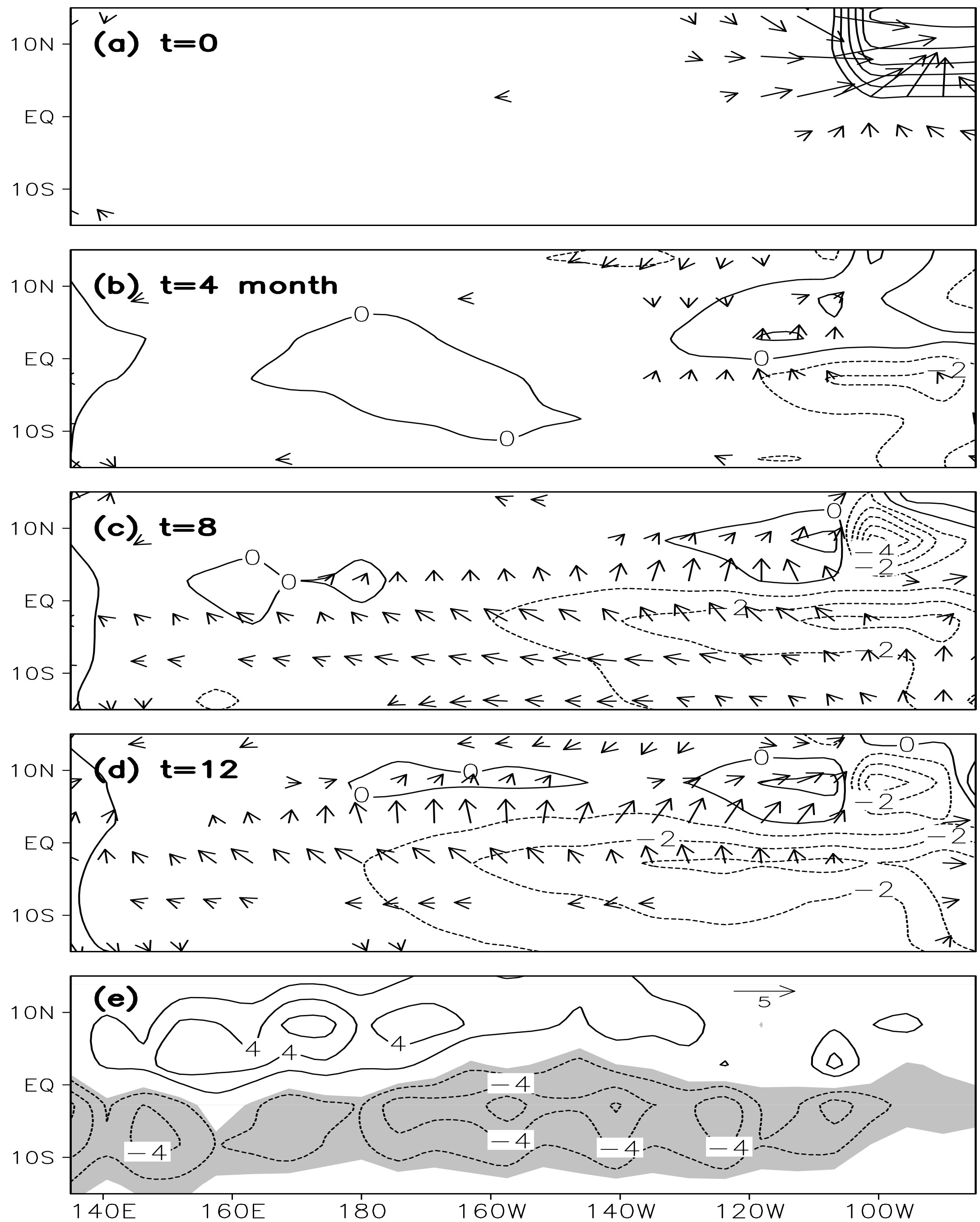

Figure 4-9. (a) Direct response of an atmospheric GCM to the addition of a northern land bulge: Anomalies of surface temperature (contours) and wind velocity. The evolution of surface temperature (contours in °C) and SST-induced surface wind velocity (m/s) changes 4, 8, and 12 months after the ocean model is turned on. (e) Changes in precipitation (mm/s) after the coupled model reaches a statistical steady state. From Xie and Saito (2001).

Regarding the Atlantic, the authors point to the bulge of West Africa, which is hotter than the ocean to the south and induces the southerly winds in the Gulf of Guinea. The winds in turn cause upwelling both off the South African coast and in the open ocean south of the equator. Xie and Saito (2001) suggest that the weakening of the northeasterly trades east of West Africa, in contrast to strong southeasterlies south of the equator, is an additional symmetry-breaking effect of the West African bulge by reducing local evaporation and hence warming the sea surface north of the equator.

Initial value problems shed further light on the role of air-sea interaction in establishing basin-wide climatic asymmetry. Figure 4-9 shows the coupled ocean-atmosphere adjustment to a sudden addition of a land bulge on a symmetric land strip north of the equator. The direct effect of this continental forcing is rather limited in space (Fig. 4-9a). When the ocean is allowed to interact with the atmosphere, however, it triggers a coupled wave front that moves quickly westward, pushing the ITCZ north of the equator on its way. In a matter of less than a year, strong climatic asymmetry resembling that observed in the Pacific and Atlantic is established basin-wide. Similar initial value problems are studied by Xie (1996a) with simpler models and by Ma et al. (1996) and Kimoto and Shen (1997) with coupled GCMs.

6. CONSEQUENCES OF ITCZ ASYMMETRY

6.1. Atmosphere and Ocean Circulation

The ITCZ supplies much of the heat that drives the global atmospheric circulation, and the latitude of its position is an important parameter for the global atmosphere and climate. A northward-displaced ITCZ induces hemispheric asymmetry in the Hadley circulation, with the southern cell being stronger than the northern one (Lindzen and Hou 1988). This difference leads to a stronger subtropical westerly jet in the Southern Hemisphere upper troposphere (Hou 1993; Chang 1995). This asymmetry in the subtropical jets furthermore results in temperature and storm activity differences.

In the Pacific, upper ocean circulation develops strong hemispheric asymmetries in response to wind curls associated with the northward-displaced ITCZ. The North Equatorial Countercurrent (NECC) is one such asymmetry, flowing eastward against the easterly trades and in geostrophic balance with a thermocline ridge beneath the ITCZ[6] (Wyrtki and Koblinsky 1984; Kessler and Taft 1987). Such an eastward countercurrent is not observed south of the equator.

Temperature stratification in the equatorial oceans is maintained by cold water subducted during the winter in the subtropics and transported into the equator along the thermocline by the so-called subtropical cells (STCs; see Liu and Philander [2001] for a review). The subtropical water reaches the equator under the surface via an interior pathway in the South Pacific, while it goes mostly through the western boundary current in the North Pacific (Lu and McCreary 1995). This asymmetry in STCs, along with that in precipitation, moreover gives rise to pronounced hemispheric differences in thermocline salinity distribution (Nonaka and Takeuchi 2001). Given strong mixing/dissipation in the western boundary current, this hemispheric difference in the thermocline water pathway may have important consequences for decadal variability (Gu and Philander 1997; Nonaka et al. 2002; McPhaden and Zhang 2002).

6.2. Equatorial Annual Cycle

The disparity between solar forcing and climatic response is also seen on the seasonal time scale, along the equator in the eastern Pacific and the Atlantic. Despite little seasonal change in insolation on the equator, SST displays a pronounced annual cycle at the Galapagos Islands (90°W, equator), rising to a maximum of 27°C in March and dropping to a minimum of below 22°C in September. For comparison, 20° to the north, near Hawaii, with much larger annual variations in insolation, the annual range in SST is only about half as large, between 24°C and 27°C.

The annual cycle of SSTs on the equator displays a pronounced westward propagation in the eastern Pacific (Horel 1982) and the Atlantic, which Mitchell and Wallace (1992) show is a result of its interaction with zonal wind. Xie (1994b) suggests that the northward displacement of the ITCZ is the ultimate cause of this peculiar annual cycle on the equator (see also Giese and Carton [1994]). In the eastern Pacific and Atlantic, the ITCZ stays north of the equator throughout the year, intensifying in boreal summer and weakening in winter in response to annual solar forcing off the equator (Fig. 4-4). As a result, the cross-equatorial southerlies that converge

[6] The collocation of the NECC with the atmospheric ITCZ led many investigators to speculate that its advection of warm water from the western Pacific is a major reason for the high-SST band in 5°N–10°N. NECC's advective effect turns out to be of secondary importance for SST asymmetry since the zonal SST gradient is very weak along it (Fig. 4-4; Xie 1994a).

onto the ITCZ strengthen in boreal summer and fall, causing equatorial SST to fall by enhancing the upwelling/entrainment of cold water and surface evaporation. Conversely, the relaxed southerlies in boreal spring warm up the equatorial ocean mixed layer by reducing cold upwelling and evaporation. If the ITCZ were displaced to the south of the equator, the equatorial seasonal variations would still be dominated by the annual cycle, but its phase would be opposite—warm in September and cold in March at the Galapagos.

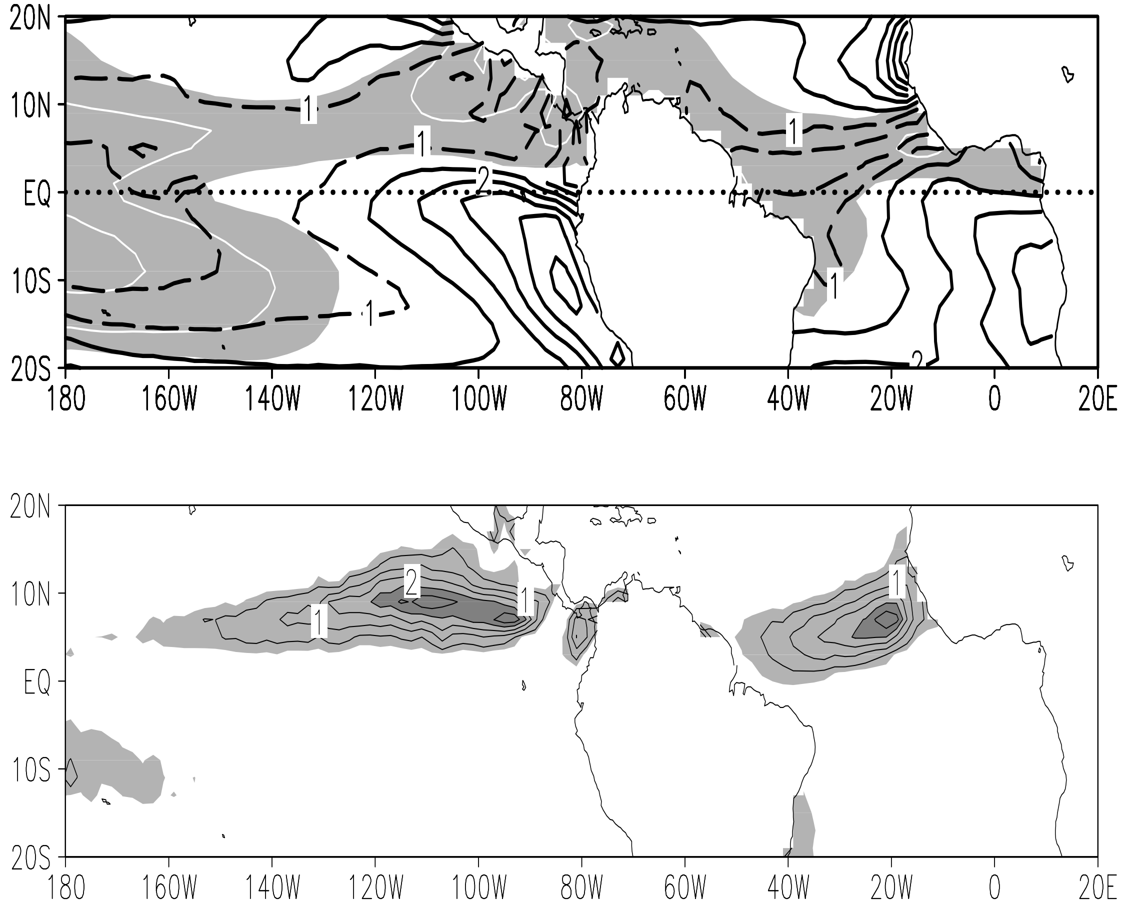

Figure 4-10. (top) Annual harmonic of SST (contours at 0.5°C intervals; values at or less than 1°C dashed), superimposed on annual mean SST (white contours at 1°C intervals; shade > 27°C). (bottom) Effect of seasonal cycle on scalar wind speed (contours at 0.5 m/s intervals; light shade > 0.25 m/s and dark shade >2 m/s). Based on COADS data.

Thus the ITCZ is the climatic equator from the standpoint of seasonal variations. South of the ITCZ the seasonal cycle bears the Southern Hemisphere characteristics of being warm in March and cold in September. To its north, SST peaks in September and is coldest in March. Figure 4-10a shows the annual harmonic of observed SST, with shaded areas indicating regions of high SST. Over most of the Atlantic and east of 140°W in the Pacific, where the ITCZ stays north of the equator and the annual mean southerlies are maintained on the equator, the SST annual harmonic reaches a local maximum on or slightly south of the equator because of the shallow thermocline and strong air-sea interaction there. The annual SST harmonic is at its minimum along the climatological ITCZ, consistent with the notion of its being the climatic equator. This collocation of the annual harmonic minimum and ITCZ is particularly conspicuous in the Atlantic. The negative feedback between convective clouds and local SST seems partly responsible for the weak annual variations in SST under the ITCZ over both the Pacific and the Atlantic.

6.3. Tropical Atlantic Variability

The position of the Atlantic ITCZ displays considerable variability in latitude on interannual to decadal time scales, resulting in droughts in some years and floods in other years in northeast Brazil (Hastenrath 1991). Empirically, such an anomalous meridional displacement of the ITCZ is known to be associated with anomalies in cross-equatorial gradients in SST and zonal wind (Nobre and Shukla 1996; Chiang, Chapter 16, "Present-Day Climate Variability in the Tropical Atlantic: A Model for Paleoclimatic Changes," this volume), very similar to the coupled pattern in Figure 4-5. Chang et al. (1997) show that air-sea interaction and the WES feedback in particular help to give rise to the observed association between anomalies of ITCZ rainfall, surface winds, and SST. Subsequent studies suggest that both the stratus-SST (Tanimoto and Xie 2002) and upwelling-SST (Xie and Saito 2001) feedbacks discussed in Section 4 also contribute. Results of Okumura et al.'s (2001) atmospheric GCM experiments suggest that an anomalous northward shift of the Atlantic ITCZ, forced by an SST pattern like the one in Figure 4-5, weakens the atmospheric Azores high in the subtropics by weakening the local Hadley cell over the North Atlantic.

The northward displacement of the climatological ITCZ over the Atlantic may affect the space-time structure of tropical variability. The WES feedback owes its positive sign to the sign change in the Coriolis parameter across the equator and is most effective under the north-south symmetric mean state. The departure of the mean ITCZ from the geographic equator weakens the WES feedback and hence the interhemispheric interaction (Okajima et al. 2003). This weakening of interaction is possibly responsible for the observed lack of significant correlation between tropical North and South Atlantic SSTs (Houghton and Tourre 1992).

7. CHALLENGES AHEAD FOR REALISTIC CLIMATE SIMULATION

Despite all of the progress outlined above, many state-of-the-art global coupled GCMs still have a problem keeping the Pacific and the Atlantic ITCZ north of the equator. (Mechoso et al. [1995] dub this deficiency as the double ITCZ syndrome.) In these GCMs, deep convection lingers for too long south of the equator, and in some models, the southern ITCZ persists throughout the year over the eastern Pacific, resulting in a double ITCZ in the annual mean climatology instead of the observed northward-displaced single ITCZ. This deficiency in simulating the observed climatic asymmetry has been identified as a major challenge at a recent workshop devoted to discussion of the tropical biases of GCMs (http://pod.tamu.edu/~bias/). These biases have important implications for climate simulation and prediction. For example, the failure to keep the eastern Pacific ITCZ north of the equator is certain to affect the simulation of the equatorial annual cycle, which is known to affect the properties of ENSO, most notably its seasonal phase-locking (Jin et al. 1994; Li and Hogan 1999). The properties of ENSO are highly sensitive to even slight changes in the mean state; the eastern equatorial ocean is considerably cooler with the ITCZ displaced to the north of the equator than with a double symmetric ITCZ because of the upwelling induced by the cross-equatorial southerlies (Fig. 4-7). The meridional configuration of the mean ITCZ also affects the WES feedback and the cross-equatorial mode of climate variability (Okajima et al. 2003) that is most pronounced in the Atlantic.

Ad hoc flux adjustments are often used in simulation and prediction models to prevent them from drifting quickly away from the realistic climate. Because the same air-sea feedbacks contribute both to the mean state and to causing temporal variability (e.g., the Bjerknes feedback for the Walker circulation/ENSO and the WES for climatic asymmetry/Atlantic cross-equatorial variability), the failure of a model to maintain a realistic mean climate indicates that it is representing these feedbacks poorly and it may be severely distorting variability. Dijkstra and Neelin (1999) show that by requiring a coupled ocean-atmosphere model to simulate a realistic cold tongue, they can narrow the parameter space for ENSO and reduce the ambiguity in parameter choice compared with a flux-adjusted version of the same model.

In light of the importance of the climatic asymmetry as discussed above and the difficulty maintaining it in GCMs, we discuss some remaining issues in this section.

7.1. Effect of Seasonal Forcing

We have implicitly assumed so far that latitudinal asymmetry of long-term annual mean climate is independent of seasonal variations, an assumption that is generally untrue. Here we consider the effect of seasonal cycle on scalar wind speed (W), which exerts a strong influence on SST via evaporation and is a nonlinear function of zonal and meridional wind velocities. The crudest estimate of W is obtained by using the annual mean climatological wind components (u_c, v_c), namely,

$$W_C = \sqrt{u_c^2 + v_c^2} \, ,$$

In the ITCZ in the far eastern Pacific and the Atlantic, both (u_c, v_c) and hence W_C approach zero, which would require a very high SST to balance the insolation. A better estimate of scalar wind speed results from using monthly climatological wind velocity (u_m, v_m) instead, with m denoting calendar month,

$$W_M = \sum_{m=1}^{12} \sqrt{u_m^2 + v_m^2} \, / 12 \, .$$

Figure 4-10b shows the difference between these two estimates, W_M–W_C, which measures the effect of seasonal-varying winds on scalar wind speed and hence SST. The effect of seasonal variations on scalar wind is large along the climatological ITCZ and mostly due to the seasonal migration of the ITCZ and its weak-wind zone, the latter being around the equator in March but moving to 10°N in August (Fig. 4-4).

Amounting to a 2 m/s increase in wind speed along the ITCZ north of the equator, this effect of seasonal forcing reduces cross-equatorial differences in wind speed and hence the latitudinal asymmetry of SST and ITCZ generated by the WES and stratus-SST feedbacks. This weakening of annual mean climatic asymmetry by seasonal forcing has been demonstrated in a simple model (Xie 1996b). While different views exist regarding the role of the annual cycle (Wang and Wang 1999), a realistic simulation of the seasonal migration of the ITCZ is very important for maintaining a realistic degree of climatic asymmetry and vice versa.

7.2. Stratus Cloud Deck

Several coupled GCM studies consistently point to the poor representation of marine boundary layer clouds in the South Pacific as the major cause of the double ITCZ syndrome (Philander et al. 1996; Ma et al. 1996; Kimoto and Shen 1997; Gordon et al. 2000). The loosely used term “stratus” in part of the air-sea interaction community (Klein and Hartmann 1993; Philander et al. 1996) includes at least two distinct cloud types: stratocumulus and trade cumulus. Stratocumulus clouds form in the surface mixed layer, which is tightly coupled with the sea surface, are more area extensive in space and more persistent in time, and hence record larger values of cloud cover than trade cumulus clouds, which rise higher but cover smaller areas and are of shorter duration. The trade cumulus clouds are cumulus clouds formed in the so-called decoupled boundary layer where a stable layer exists between the surface mixed layer and MBL top. Norris (1998) analyzes vertical soundings made at ocean weather stations and shows that the decrease in “stratus” cloud cover with increasing SST (or decreasing static stability of the lower atmosphere) in Klein and Hartmann (1993) is associated with the change in cloud type from stratocumulus over low SSTs to cumulus over higher SSTs (his Fig. 4). Because this cloud type transition is associated with a large change in cloud-layer coupling with the sea surface, Norris suggests that adequate modeling of the decoupled MBL is the key to simulating the sensitivity of cloud type and cloud cover to changes in SST. This sensitivity gives rise to the stratus-SST feedback discussed in Section 4.2.

Typical GCMs have only a few levels in the first 2 km above the sea surface, a resolution that is insufficient to represent the aforementioned transition in MBL and cloud types. Regional atmospheric models (RAMs), affording higher resolutions and more sophisticated physics for turbulence and clouds, can be a useful tool bridging global GCMs on one hand and field observations and large-eddy cloud simulations on the other. Figure 4-11 shows an example of an RAM simulation, with a vertical transect in the South Pacific. In general, MBL clouds are capped by a temperature inversion whose strength decreases westward. Over the cold water off the coast of Peru, the MBL is shallow and stratocumulus clouds form in the surface mixed layer. Both cloud base and top rise toward the west with SST increasing. West of 110°W, the MBL becomes decoupled from the sea surface, with cumulus clouds transporting moisture through the stable layer into a cloud layer below the inversion[7]. Sensitivity experiments show that detrain-

[7] Lasting for short time, cumuli do not leave a strong signal in the time-mean cloud water field.

ment of cloud water at the cloud top is important for maintaining the inversion-cloud couplet (McCaa and Bretherton 2003; Wang et al. 2005). In the above RAM, the intense radiative cooling at the cloud top also induces a strong downdraft that is confined to the cloud layer. The simulated westward transition from stratocumulus in a coupled MBL to cumulus in an uncoupled MBL is consistent with the few observations that exist for this region (Garreaud et al. 2001).

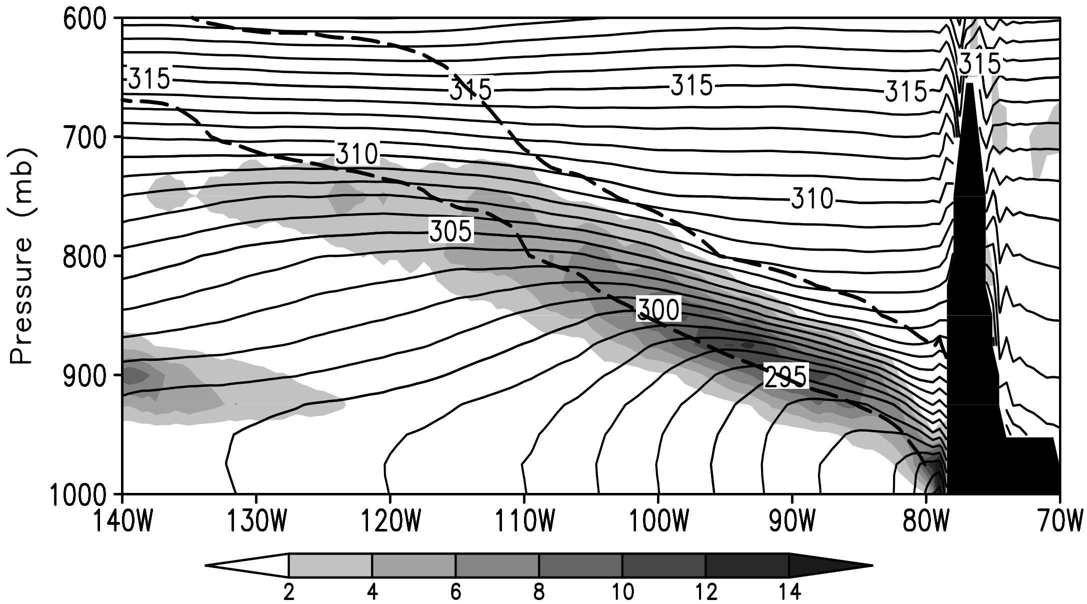

Figure 4-11. Simulated cloud liquid water content (shaded in 10^{-2} g/kg) and virtual potential temperature (solid contours in K) in Xu et al.'s (2004) regional atmospheric model, averaged for August–October 1999 along 10°S. Dashed lines denote the boundaries of the inversion layer. The Andes are shaded in black.

7.3. Andes

The steep and narrow Andes are an overlooked continental forcing. At the equator, the mountain range is only 200 km wide, rising from near sea level to 3.5 km high in less than 100 km. At typical resolutions of 300 km, global climate models severely smooth the Andean mountains to less than 1 km high between 10°S and 10°N, while in reality they are rarely lower than 3 km. Blocking the easterly winds, the Andes induce downward motion on the leeside and thereby help maintain the temperature inversion and the stratocumulus cloud deck off South America. In a RAM experiment that reduces the height of the Andes, the resultant anomalous convergence offshore weakens the inversion and reduces the stratocumulus cloud cover in the August–October season while prolonging the southern ITCZ in March and April (Xu et al. 2004). While these mountain effects on the atmosphere are confined to the coastal region, they can have a basin-wide impact on Pacific climate with the help of coupled ocean-atmosphere waves (Section 5). Thus, smoothing the Andes in global GCMs may contribute to the double-ITCZ syndrome and too weak climatic asymmetry.

8. SUMMARY

Halley recognized the effect of the West African bulge on the winds in the Gulf of Guinea. He wrote: "if a Country lying near the Sun . . . such as the Deserts of Lybia [sic] are usually reported to be, the heat occasioned by the reflection of the Sun's Beams, and the retention there of in the Sand, is incredible to those that have not felt it; whereby the Air being exceedingly rarified, it is necessary that this cooler and more dense Air should run thitherwards to restore the Equilibrium: This I take to be the cause, why near the Coast of Guinea the Wind always sets in upon the Land, blowing Westerly instead of Easterly." Amazingly, the wind map Halley drew 320 years ago contains many of the elements necessary to build a modern solution to the age-old riddle of climatic asymmetry. The bulge of West Africa causes cross-equatorial southerlies in the Gulf of Guinea, initiating an air-sea coupled wave front that pushes the southeast trades to cross the equator into the Northern Hemisphere where they meet the northeast trades. Over the warm waters where these two trade wind systems meet, convection takes place, producing the thunderstorms and rains Halley recorded. On average, winds are calm in the Doldrums of the ITCZ, allowing water to stay warm and maintaining the climatic asymmetry.

It is not the atmosphere, nor the ocean alone, not even their coupling, but the collective effort of the ocean, atmosphere, and the land that gives rise to the long silver band of clouds, stretching north of the equator by half the globe in the Pacific and Atlantic (Fig. 4-2). Like the ENSO phenomenon, this climatic asymmetry attests yet again to the importance of air-sea interaction in making the response of earth's climate deviate considerably in space and time from the pattern expected by the solar forcing alone. Given that the Pacific and Atlantic ITCZ has remained north of the equator for hundreds if not thousands of years, it is quite natural to suggest that the land-sea distribution is the ultimate cause. This answer, however, lacks crucial details such as which continental features are the cause and how they influence climate thousands of kilometers away. Recent studies of air-sea coupling have enabled us to narrow the search of continental forcing to the eastern side of an ocean basin. Furthermore, they have demonstrated that, perhaps to the surprise of many, the direct cause of the great climatic asymmetry over the Pacific and Atlantic is found in regional features of the continents such as the coastal line, not the overall distribution of landmass, such as Asia in the Northern Hemisphere versus Australia in the Southern Hemisphere. This is not to say that the global landmass distribution is of no consequence, but its role is rather indirect. The global distribution of land and seas, for example, is the reason that easterlies prevail in the equatorial Pacific and Atlantic and westerlies in the equatorial Indian Ocean (e.g., Philander et al. 1996), allowing the equatorial cold tongue and the northward-displaced ITCZ to develop in the former oceans, but not in the latter. Much needs to be studied as to what shapes and anchors the climatological rain bands over the Indian and western Pacific where SST gradients are weak and continental influences are strong.

On the geological time scale, continents constantly move and change their shape, orientation, and relative position. For example, in the middle Eocene 50 million years ago, the Andes were lower, and North America was more than 10° north of its present position and unattached to South America (www.scotese.com). Connected with the Pacific through the gap between North and South America, the Atlantic was narrower, with the south coast of West Africa closer to the equator. These changes in landmass distribution are certain to affect the meridional configuration of the ITCZ and other features of the tropical and global climate. Coupled dynamic modeling of continental effects as presented here offers a physical basis for interpreting paleoclimate records and a useful framework for inferring how earth's climate has evolved with drifting continents and changing landscapes.

9. ACKNOWLEDGMENTS

The author would like to thank J. Matsumoto for access to the library of the Department of Geography, University of Tokyo; C.-H. Chang for a literature search; and G. Speidel and H. Xu for helpful comments. This work is supported by NOAA, NSF, NASA, NSFC, and JASTEC. IPRC contribution #279 and SOEST contribution #6384.

10. REFERENCES

Bartholomew, J.G., and A.J. Herbertson. 1899. In, Buchan, A. (ed.). *Atlas of Meteorology*. Edinburgh: Royal Geographical Society.

Chang, E.K.M. 1995. The influence of Hadley circulation intensity changes on extratropical climate in an idealized model. *Journal of Atmospheric Science* 52: 2006–2024.

Chang, P., and S.G.H. Philander. 1994. A coupled ocean-atmosphere instability of relevance to the seasonal cycle. *Journal of Atmospheric Science* 51: 3627–3648.

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

Charney, J.G. 1971. Tropical cyclogenesis and the formation of the intertropical convergence zone. In, Reid, W.H. (ed.). *Mathematical Problems of Geophysical Fluid Dynamics.* Providence, Rhode Island: American Mathematics Society.,.

Chelton, D.B., S.K. Esbensen, M.G. Schlax, N. Thum, M.H. Freilich, F.J. Wentz, C.L. Gentemann, M.J. McPhaden, and P.S. Schoph. 2001. Observations of coupling between surface wind stress and sea surface temperature in the eastern tropical Pacific. *Journal of Climate* 14: 1479–1498.

Cronin, M.F., N. Bond, C. Fairall, J. Hare, M.J. McPhaden, and R.A. Weller. 2002. Enhanced oceanic and atmospheric monitoring underway in Eastern Pacific. *Eos Trans.* American Geophysical Union, 83(19): 205, 210–211.

Deser, C., J.J. Bates, and S. Wahl. 1993. The influence of sea surface temperature on stratiform cloudiness along the equatorial front in the Pacific Ocean. *Journal of Climate* 6: 1172–1180.

Dijkstra, H.A., and J.D. Neelin. 1999. Coupled processes and the tropical climatology. Part III: Instabilities of fully coupled climatology. *Journal of Climate* 12: 1630–1643.

Fu, X., and B. Wang. 2001. A coupled modeling study of the seasonal cycle of the Pacific cold tongue, Part I: Simulation and sensitivity experiments. *Journal of Climate* 14: 756–779.

Garreaud, R.D., J. Rutllant, J. Quintana, J. Carrasco, and P. Minnis. 2001. CIMAR-5: A snapshot of the lower troposphere over the subtropical Southeast Pacific. *Bulletin of the American Meteorological Society* 82: 2193–2207.

Giese, B.S., and J.A. Carton. 1994. The seasonal cycle in a coupled ocean-atmosphere model. *Journal of Climate* 7: 1208–1217.

Gordon, C.T., A. Rosati, and R. Gudgel. 2000. Tropical sensitivity of a coupled model to specified ISCCP low clouds. *Journal of Climate* 13: 2239–2260.

Gu, D., and S.G.H. Philander. 1997. Interdecadal climate fluctuations that depend on exchange between the tropics and extratropics. *Science* 275: 805–807.

Gu, G., and C. Zhang. 2002. Westward-propagating synoptic-scale disturbances and the ITCZ. *Journal of Atmospheric Science* 59: 1062–1075.

Halley, E. 1686. A historical account of the trade winds, and monsoons, observable in the seas between and near the Tropicks, with an attempt to assign the phisical cause of the said winds. *Philosophical Transactions of the Royal Society of London*, 16: 153–168.

Hashizume, H., S.-P. Xie, W.T. Liu, and K. Takeuchi. 2001. Local and remote atmospheric response to tropical instability waves: A global view from the space. *Journal of Geophysical Research-Atmosphere*, 106: 10173–10185.

Hastenrath, S. 1991. *Climate Dynamics of the Tropics*. Boston: Kluwer Academic, 488 pp.

Held, I.M., and A.Y. Hou. 1980. Nonlinear axially symmetric circulations in a nearly inviscid atmosphere. *Journal of Atmospheric Science* 37: 515–533.

Horel, J.D. 1982. On the annual cycle of the tropical Pacific atmosphere and Ocean. *Monthly Weather Review*, 110: 1863–1878.

Hou, A.Y. 1993. The influence of tropical heating displacement on the extratropical climate. *Journal of Atmospheric Science* 50: 3553–3570.

Houghton, R.W., and Y.M. Tourre. 1992. Characteristics of low-frequency sea surface temperature fluctuations in the tropical Atlantic. *Journal of Climate* 5: 765–771.

Jin, F.-F., J.D. Neelin, and M. Ghil. 1994. El Niño on the devil's staircase: Annual subharmonic steps to chaos. *Science* 264: 70–72.

Kessler, W.S., and B.A. Taft. 1987. Dynamic heights and zonal geostrophic transports in the central tropical Pacific during 1979–84. *Journal of Physical Oceanography* 17: 97–122.

Kimoto, M., and X. Shen. 1997. Climate variability studies using general circulation models. In, Sumi, A. (ed.). *The Frontiers of Climate Research II.* Center for Climate System Research, University of Tokyo, pp. 91–116.

Klein, S.A., and D.L. Hartmann. 1993. The seasonal cycle of low stratiform clouds. *Journal of Climate* 6: 1587–1606.

Köppen, W. 1899. Winds over the oceans. In, Buchan, A. (ed.). *Atlas of Meteorology*. Edinburgh: Royal Geographical Society, Plate 14.

Kornfield, J., A.F. Hasler, K.J. Hanson, and V.E. Suomi. 1967. Photographic cloud climatology from ESSA III and V computer produced mosaics. *Bulletin of the American Meteorological Society* 48: 878–883.

Li, T., and T.F. Hogan. 1999. The role of the annual-mean climate on seasonal and interannual variability of the tropical Pacific in a coupled GCM. *Journal of Climate* 12: 780–792.

Lindzen, R.S., and A.Y. Hou. 1988. Hadley circulation for zonally averaged heating centered off the equator. *Journal of Atmospheric Science* 45: 2416–2427.

Liu, W.T., and X. Xie. 2002. Double intertropical convergence zones—A new look using scatterometer. *Geophysical Research Letters* 29: 2072, doi:10.1029/2002GL015431.

Liu, Z., and G. Philander. 2001. Tropical-extratropical oceanic exchange pathways. In, Siedler, G., J. Church, and W.J. Gould (eds.). *Ocean Circulation and Climate: Observing and Modeling the Global Ocean*. San Diego: Academic Press, pp. 247–254.

Lu, P., and J.P. McCreary. 1995. Influence of the ITCZ on the flow of the thermocline water from the subtropical to the equatorial Pacific Ocean. *Journal of Physical Oceanography* 25: 3076–3088.

Ma, C.-C., C.R. Mechoso, A.W. Robertson, and A. Arakawa. 1996. Peruvian stratus clouds and the tropical Pacific circulation—A coupled ocean-atmosphere GCM study. *Journal of Climate* 9: 1635–1645.

Manabe, S. 1969. Climate and the ocean circulation. Part 2: The atmospheric circulation and the effects of heat transfer by ocean currents. *Monthly Weather Review* 97: 775–805.

Manabe, S., D.G. Hahn, and J.L. Holloway. 1974. The seasonal variation of the tropical circulation as simulated by a global model of the atmosphere. *Journal of Atmospheric Science* 31: 43–83.

McCaa, J.R., and C.S. Bretherton. 2003. A new parameterization for shallow cumulus convection and its application to marine subtropical cloud-topped boundary layers: Part II: Regional simulations of marine boundary layer clouds. *Monthly Weather Review* 132: 883–896.

McPhaden, M.J., and D. Zhang. 2002. Slowdown of the meridional overturning circulation in the upper Pacific Ocean. *Nature* 415: 603–608.

Mechoso, C.R., A.W. Robertson, and Coauthors. 1995. The seasonal cycle over the tropical Pacific in general circulation models. *Monthly Weather Review* 123: 2825–2838.

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

Möller, F. 1951. Viertel Jahrs Karten des Niederschlags fur die Ganze Erde. *Petermanns Geograph. Mitt.*, 95: 1–7.

Nigam, S. 1997. The annual warm to cold phase transition in the eastern equatorial Pacific: Diagnosis of the role of stratus cloud-top cooling. *Journal of Climate* 10: 2447–2467.

Nobre, P., and J. Shukla. 1996. Variations of sea surface temperature, wind stress, and rainfall over the tropical Atlantic and South America. *Journal of Climate* 9: 2464–2479.

Nonaka, M., and K. Takeuchi. 2001. Tropical subsurface salinity and tritium distributions in the Pacific: Their differences and formation mechanisms. *Journal of Physical Oceanography* 31: 1388–1395.

Nonaka, M., S.-P. Xie, and J.P McCreary. 2002. Decadal variations in the subtropical cells and equatorial Pacific SST. *Geophysical Research Letters* 29: 1116, doi: 10.1029/2001GL013676.

Norris, J.R. 1998. Low cloud type over the ocean from surface observations. Part I: Relationship to surface meteorology and the vertical distribution of temperature and moisture. *Journal of Climate* 11: 369–382.

Numaguti, A., and Y.-Y. Hayashi. 1991. Behaviors of cumulus activity and the structures of circulations in an aqua-planet model. *Journal of the Meteorological Society of Japan* 69: 563–579.

Okajima, H., S.-P. Xie, and A. Numaguti. 2003. Interhemispheric coherence of tropical climate variability: Effect of climatological ITCZ. *Journal of the Meteorological Society of Japan* 81: 1371–1386.

Okumura, Y., S.-P. Xie, A. Numaguti, and Y. Tanimoto. 2001. Tropical Atlantic air-sea interaction and its influence on the NAO. *Geophysical Research Letters* 28: 1507–1510.

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

Pike, A.C. 1971. Intertropical convergence zone studied with an interacting atmosphere and ocean model. *Monthly Weather Review* 99: 469–477.

Raymond, D.J., S.K. Esbensen, C. Paulson, M. Gregg, C.S. Bretherton, W.A. Petersen, R. Cifelli, L.K. Shay, C. Ohlmann, and P. Zuidema. 2004. EPIC2001 and the coupled ocean-atmosphere system of the tropical east Pacific. *Bulletin of the American Meteorological Society* 85: 1341–1354.

Reynolds, R.W., and T.M. Smith. 1994 Improved global sea surface temperature analyses using optimal interpolation. *Journal of Climate* 7: 929–948.

Small, R.J., S.-P. Xie, Y. Wang, S.K. Esbensen, and D. Vickers. 2005. Numerical simulation of boundary layer structure and cross-equatorial flow in the eastern Pacific. *Journal of Atmospheric Science* 62: in press.

Tanimoto, Y., and S.-P. Xie. 2002. Inter-hemispheric decadal variations in SST, surface wind, heat flux and cloud cover over the Atlantic Ocean. *Journal of the Meteorological Society of Japan* 80: 1199–1219.

Thompson, R.M., S.W. Payne, E.E. Recker, and R.J. Reed. 1979. Structure and properties of synoptic-scale wave disturbances in the intertropical convergence zone of the eastern Atlantic. *Journal of Atmospheric Science* 36: 53–72.

U.S. Air Force and U.S. Department of Commerce. 1971. *Global Atlas of Reflective Cloud Cover, 1967–1970*. Washington, D.C.

Waliser, D.E., and N.E. Graham. 1993. Convective cloud systems and warm-pool surface temperatures: Coupled interactions and self-regulation. *Journal of Geophysical Research* 98: 12881–12893.

Wallace, J.M., T.P. Mitchell, and C. Deser. 1989. The influence of sea surface temperature on surface wind in the eastern equatorial Pacific: Seasonal and interannual variability. *Journal of Climate* 2: 1492–1499.

Wang, B., and Y. Wang. 1999. Dynamics of the ITCZ-equatorial cold tongue complex and causes of the latitudinal asymmetry. *Journal of Climate* 12: 1830–1847.

Wang, C., S.-P. Xie, and J. A. Carton (eds.). 2004. *Earth Climate: The Ocean-Atmosphere Interaction*, Geophysical Monograph 147. Washington D.C.: American Geophysical Union, pp. 1-19.

Wang, Y., S.-P. Xie, B. Wang, and H. Xu. 2005. Large-scale atmospheric forcing by southeast Pacific boundary-layer clouds: A regional model study. *Journal of Climate* 18: 934-951.

Woodruff, S.D., R.J Slutz, R.L. Jenne, and P.M. Steurer. 1987. A comprehensive ocean-atmosphere dataset. *Bulletin of the American Meteorological Society* 68: 521–527.

Wyrtki, K., and B. Koblinsky. 1984. Mean water and current structure during the Hawaii to Tahiti shuttle experiment. *Journal of Physical Oceanography* 14: 242–254.

Xie, P., and P.A. Arkin. 1996. Analyses of global monthly precipitation using gauge observations, satellite estimates, and numerical model predictions. *Journal of Climate* 9: 840–858.

Xie, S.-P. 1994a. Oceanic response to the wind forcing associated with an ITCZ in the northern hemisphere. *Journal of Geophysical Research-Oceans* 99: 20393–20402.

Xie, S.-P. 1994b: On the genesis of the equatorial annual cycle. *Journal of Climate* 7: 2008–2013.

Xie, S.-P. 1996a. Westward propagation of latitudinal asymmetry in a coupled ocean-atmosphere model. *Journal of Atmospheric Science* 53: 3236–3250.

Xie, S.-P. 1996b. Effects of seasonal solar forcing on latitudinal asymmetry of the ITCZ. *Journal of Climate* 9: 2945–2950.

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

Xie, S.-P., and K. Saito. 2001. Formation and variability of a northerly ITCZ in a hybrid coupled AGCM: Continental forcing and ocean-atmospheric feedback. *Journal of Climate* 14: 1262–1276.

Xu, H., Y. Wang, and S.-P. Xie. 2004. Effects of the Andes on eastern Pacific climate: A regional atmospheric model study. *Journal of Climate* 17: 589–602.

Yu, J.-Y., and C.R. Mechoso. 1999. Links between annual variations of Peruvian stratocumulus clouds and of SST in the eastern equatorial Pacific. *Journal of Climate* 12: 3305–3318.