Chapter 4

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

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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

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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-

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[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*".

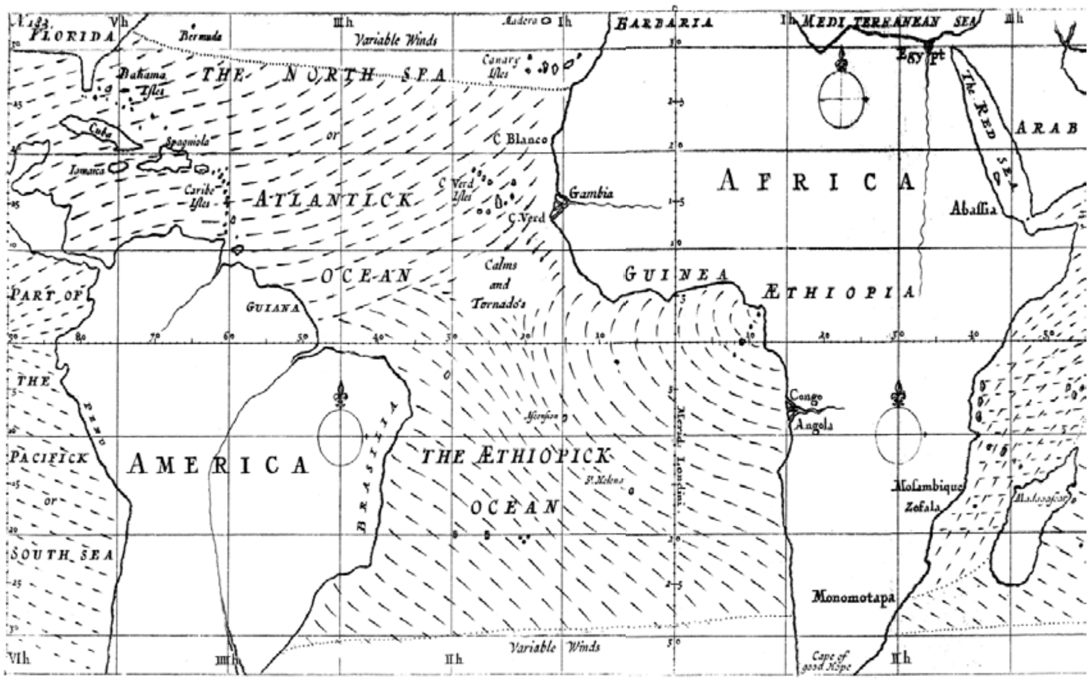

Figure 4-1. Halley's (1686) map of surface wind streamlines. The southeasterly trade winds are shown to converge onto the Northern Hemisphere.

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.

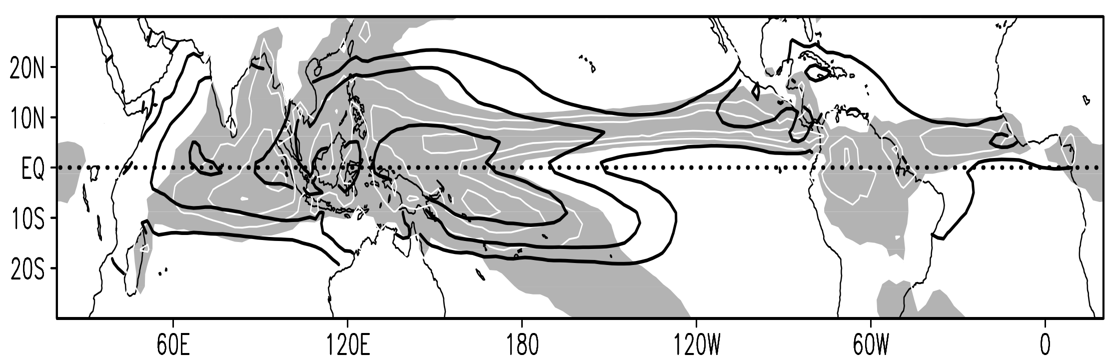

Figure 4-2. Annual-mean climatological precipitation (white contours at 2 mm/day intervals; shade > 4 mm/day), and SST (black contours at 1°C intervals; only contours for 27°C and above are plotted), based on the Climate Prediction Center Merged Analysis of Precipitation (CMAP; Xie and Arkin 1996) and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

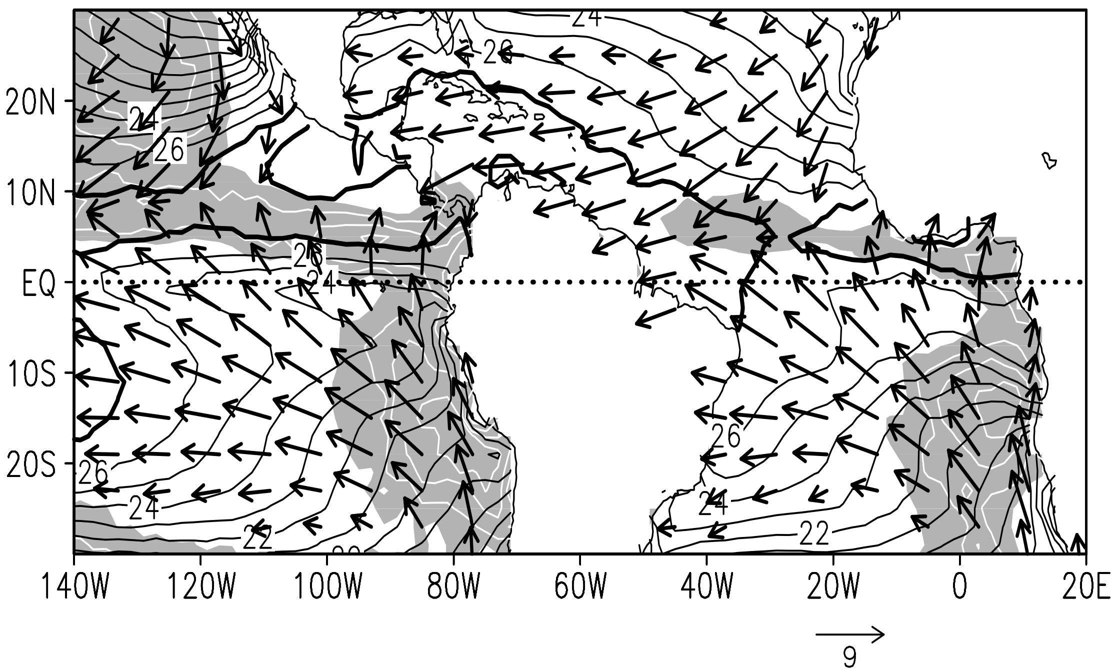

Figure 4-3. Climatological SST (contours in °C), surface wind vectors (m/s), and cloud cover (white contours at 5% intervals; shade > 60%), based on the Comprehensive Ocean-Atmospheric Data Set (COADS; Woodruff et al. 1987).

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.

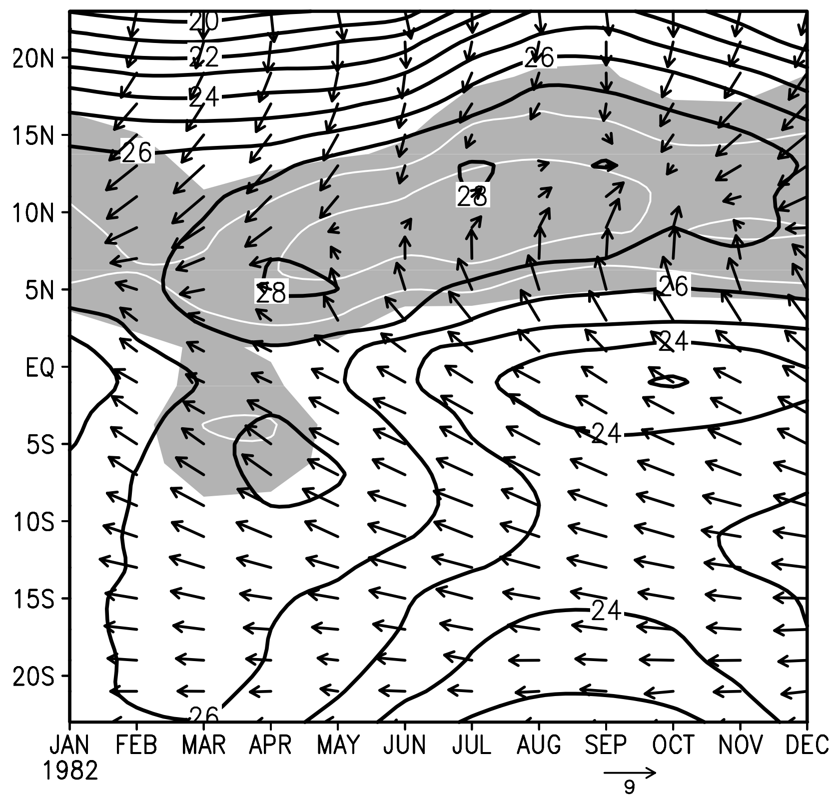

Figure 4-4. Time-latitude section of climatological SST (black contours in °C), surface wind vectors (m/s), based on COADS; and CMAP precipitation (white contours at 5 mm/d intervals; shade > 2.5 mm/d). All zonally averaged in 120°W–115°W.

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].

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[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.

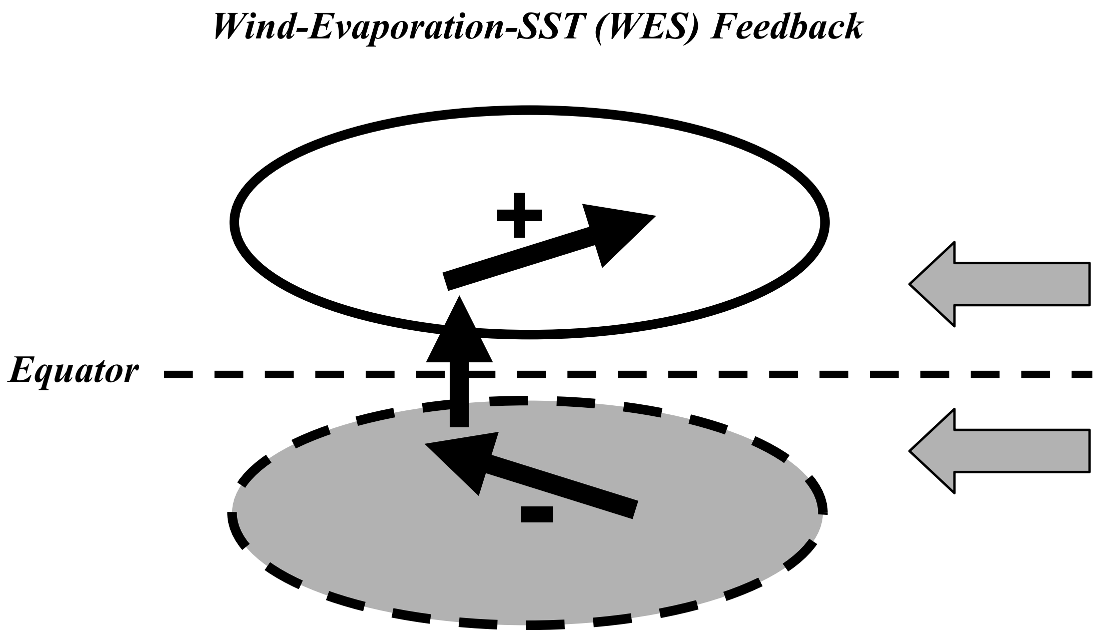

Figure 4-5. Schematic of the WES feedback: anomalies of SST in contours (negative dashed) and surface wind velocity in black vectors. The gray vectors on the right signify the background easterly trades.

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.

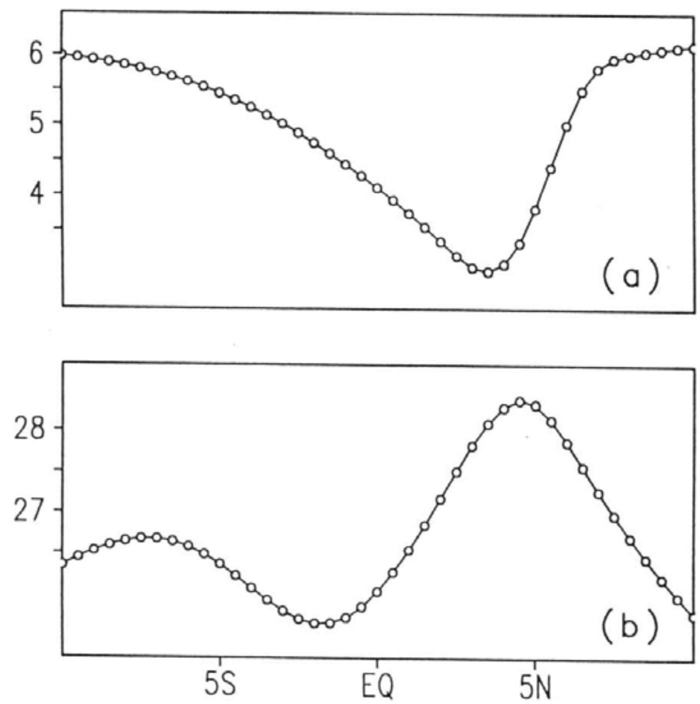

Figure 4-6. Asymmetric solution to the Xie and Philander (1994) model under equatorially symmetric conditions: (a) wind speed and (b) SST.

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.

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