The Annual Cycle in Equatorial Convection and Sea Surface Temperature

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(Manuscript received 5 December 1990, in final form 22 November 1991)

ABSTRACT

The coupled atmosphere–ocean system in the equatorial eastern Pacific and Atlantic exhibits a distinct annual cycle that is reflected in contrasting conditions at the times of the two equinoxes. The contrasts are so strong that they dominate the annual march of zonally averaged outgoing longwave radiation for the equatorial belt. The March equinox corresponds to the warm season when the equatorial cold tongues in the eastern Pacific and Atlantic are absent. With the onset of summer monsoon convection over Colombia, Central America, and West Africa in May–June, northward surface winds strengthen over the eastern Pacific and Atlantic, the equatorial cold tongues reappear, and the marine convection shifts from the equatorial belt to the intertropical convergence zones (ITCZs) along 8°N. As the northern summer progresses, the ITCZs remain strong and shift northward to near 10°N, while sea surface temperature (SST) continues to drop over the cold tongues and the southern tropics, perhaps in response to the expanding stratuscumulus cloud decks in the latter region. The cold tongue–ITCZ complex persists through the September equinox, which is characterized by suppressed convection, not only over the cold tongues, but also over much of equatorial South America.

On the basis of observational evidence concerning the timing and year-to-year regularity of the surface wind changes during the development of the cold tongues, it is argued that 1) the increase in the northward surface winds in response to the onset of the northern summer monsoon may be instrumental in reestablishing the cold tongues, and 2) positive feedbacks involving both the zonal and meridional wind components contribute to the remarkable robustness of the cold tongue–ITCZ complexes in both oceans.

1. Introduction

The annual march in insolation on the equator is forced by the semiannual cycle in solar declination and an annual cycle in earth–sun distance. Given the present tilt of the earth’s axis and the obliquity of its orbit, the former effect is about 24% larger than the latter. The climate classification scheme of Köppen (1884) was predicated on the expectation that the semiannual cycle would be dominant in the equatorial belt. It places considerable emphasis on the equinoctial wet seasons observed over parts of equatorial Africa and Indonesia, while virtually ignoring the more pronounced annual cycle observed over South America, the Atlantic, and the eastern Pacific.

A latitude–time section of climatological mean, zonally averaged outgoing longwave radiation (OLR) for the entire tropical belt is shown in Fig. 1. Consistent with Köppen’s classification scheme, the heaviest convection (indicated by the lowest values of OLR) shifts back and forth across the equator about one month behind the latitude of zero solar zenith angle. However, upon close inspection, it is evident that the equatorial wet season in March–April is much more pronounced than the one in September–October. The time series of zonally averaged, equatorial (6°N–6°S) OLR, shown in Fig. 2, exhibits a March–April minimum virtually every year. In this record, the amplitude of the annual cycle is three times as large as that of the semiannual cycle.

The dominance of the annual cycle is strongest in the half of the equatorial belt extending from 140°W eastward across South America, the Atlantic, and Africa to 40°E. A latitude–time section of climatological mean OLR for this “western” sector is shown in Fig. 3, together with the corresponding section for the “eastern” sector comprising the other half of the equatorial belt. The eastern sector is dominated by the annual march in solar declination angle, while the western sector exhibits an equally distinctive but entirely different annual march, with the band of heaviest convection taking shape in February in the southern tropics, migrating northward across the equator in March–April, and remaining in the 5°–10°N latitude belt from May almost until the end of the calendar year. Note that the western sector of the equatorial belt experiences a single, well-defined wet season in March–April. Table 1 shows climatological mean rainfall data for a selection

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of stations in the western sector. A single March–April wet season is evident at the Galapagos Islands and, to varying degrees, at all of the South American stations.

The eastern equatorial Pacific and the Atlantic oceans both exhibit a strong annual cycle in sea surface temperature (SST), with a March–April warm season [Wyrski (1965a,b), Hastenrath and Lamb (1977a), and Horel (1982) for the Pacific and Hastenrath and Lamb (1977a), Merle et al. (1980), and Picaut (1983) for the Atlantic]. It is evident from the analyses published in these works that portions of the equatorial belt experience an annual range of SSTs as large as the subtropical oceans. These regions correspond to the equatorial cold tongues, which are most pronounced during the peak of the cold season (August–October in the Pacific and July–September in the Atlantic). The remarkable regularity of the annual cycle from year to year is demonstrated in the scatterplots for the Atlantic and Pacific cold tongue regions shown in Fig. 4. Even in the Pacific, where the year-to-year variability is relatively large because of the El Niño-Southern Oscillation phenomenon, the annual range is about 2.5 times as large as the standard deviation of monthly means for individual calendar months.

Maps of the climatological mean SST, OLR, sea level pressure (SLP), and surface wind fields presented by Hastenrath and Lamb (1977a,b, 1978), Janowiak et al. (1985), Sadler et al. (1987a,b), Aceituno (1988), and others document strongly contrasting conditions during the two equinoctial seasons. In both ocean basins, March is the time of year when the convergence between the northeast and southeast trades is closest to the equator and when the cross-equatorial flow and the cold tongue are weakest. September is the time of largest equatorial asymmetry: the farthest northward penetration of the Southern Hemisphere trades and the strongest cross-equatorial flow. It also corresponds to the time when the equatorial cold tongue is strongest. In both oceans, the largest warm/cold season contrasts in SST are in the southern tropics, while the largest contrasts in OLR and surface winds are north of the equator. The seasonal contrasts are much more striking in the meridional wind component than in the zonal component.

The contrasting extreme season distributions of OLR and SST suggest a coupling of these two fields by way of the thermally forced atmospheric circulation and the wind-driven ocean circulation. The notion of the coupling of these fields was first proposed by Bjerknes (1966, 1969), who envisioned the existence of the cold tongue in the Pacific as a consequence of equatorial upwelling induced by a westward wind stress driven by convection over the warm waters of the western equatorial Pacific. Bjerknes hypothesized that the existence of the cold equatorial SSTs restricts the convection to the western portion of the basin and that variations in the strength and distribution of the convection influence the cold tongue temperatures through the equatorial zonal wind.

In this paper, we will attempt to draw some inferences concerning the nature of the atmosphere-ocean coupling in the annual cycle, based on observations of SST, surface wind, and OLR. We will argue that the meridional wind component associated with the continental monsoons and the oceanic intertropical convergence zones (ITCZs) is instrumental in this coupling and that the coupling, in turn, is responsible for the prominence of the annual cycle in the western sector of the equatorial belt. In section 3, we will examine the transition from the warm to the cold season in the "western" sector of the equatorial belt, with emphasis on OLR, SST, and surface winds over the oceans. Particular attention will be paid to the physical processes that cause the SST to drop. We will argue that dynamical forcing of the ocean is responsible for the onset of the cooling. We will speculate on the nature of this forcing (i.e., whether it is caused by an increase in zonal or meridional wind stress) in section 4. In the final section, we will offer a possible physical interpretation of the distinctive annual march observed in the western

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**Fig. 1.** Latitude–time section of zonal mean OLR for the entire tropical belt. Contour interval is 10 W m\(^{-2}\) for solid contours. The dashed vertical lines indicate the months of the equinoxes and solstices. A sun is plotted to indicate the month of the Northern Hemisphere summer solstice.

**Fig. 2.** Time series of equatorial band OLR (6°N–6°S) in W m\(^{-2}\).
sector of the equatorial belt. We will suggest that the marked differences between the two equinoctial seasons in this region might be a response of the coupled atmosphere–ocean system to the marked equatorial asymmetries in the shape of the eastern boundaries of both the Pacific and Atlantic, which permit the Northern Hemisphere summer continental monsoon rain belts to extend westward over regions of warm water and to interact with the equatorial ocean.

2. Data

The SST, sea level pressure (SLP), and surface vector wind and vector wind stress data used in this study are based on observations from ships-of-opportunity taken from the Comprehensive Ocean–Atmosphere Data Set (COADS). The individual observations that serve as input to the COADS have been grouped by 2° latitude–longitude boxes and averaged together to form arithmetic means for each year and month (Slutz et al. 1985).

The SST, SLP, and surface vector wind climatologies shown in this paper are based on a subjective analysis of the climatology for the period 1900 to 1979 by Sadler et al. (1987a, b). The major features in those analyses are in close agreement with an unsmoothed climatology derived from the COADS for the period 1946–85 (not shown). The surface vector wind-stress climatology shown in section 3 is based on data for the period 1946–85, the years of the highest sampling rate in the COADS.

OLR fields are from the TIROS-N and the NOAA-2, 3, ..., 10 polar-orbiting satellites, which span the period June 1974 to October 1987 with one gap of 10 months in 1978. They were obtained from the NOAA Climate Analysis Center in the form of monthly means on a 2.5° latitude–longitude grid. The changes in the instruments and their sampling characteristics are documented in Gruber and Krueger (1984), and Janowiak et al. (1985, 1987). Our albedo and absorbed shortwave radiation climatologies are derived from

**TABLE 1. Climatological rainfall in centimeters. The monthly rainfall amounts are taken from the World Monthly Surface Station Climatology, which is available from the National Center for Atmospheric Research.**

<table>
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<th>J</th>
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measurements taken on the same satellites as the OLR measurements. This dataset, which spans June 1974 through February 1978, was obtained from the NOAA National Environmental Satellite Data and Information Service. Both satellite datasets were interpolated onto a 2° latitude–longitude grid for compatibility with the COADS products.

3. Transition from the warm season to the cold season

The transition from the warm to the cold season can be viewed as taking place in two discrete steps, which will be referred to as Phase I and Phase II. The changes during Phase I (shown in Figs. 5 and 6 for the Pacific and Atlantic, respectively) are characterized by the following.

- The movement of the continental monsoon convection from the equatorial belt into the Northern Hemisphere, as evidenced by the marked decreases in OLR in Colombia and West Africa and the increases in the southern tropics. In both oceans, the OLR decreases extend westward into the warm pools along 9°N to form (or intensify) the oceanic ITCZs.
- An increase in the northward flow centered between the equator and 5°N, which feeds into the intensifying monsoon/ITCZ rain belts. This enhanced flow occurs in response to an increased northward-directed pressure gradient force, as shown in the lower panels of the figures. Note that the changes in zonal wind along the equator are quite small.
- Decreases in SST in the equatorial cold tongue regions and along the coasts of Peru and Angola. Note that the largest SST decreases are observed to the south of the belt in which the northward flow intensifies. In both oceans the SST decreases display a distinctive spatial signature suggestive of a response to remote forcing in the equatorial waveguide. The patterns are similar, in some respects, to the response to a steady, uniform northward wind stress in the numerical simulation of Philander and Pacanowski (1981). A similar pattern of SST decreases is also evident in maps of the month to month SST changes for both oceans (Wyrtki 1965b; Mitchell 1990; Shea et al. 1990).

The corresponding changes observed during Phase II of the transition are shown in Figs. 7 and 8. They are characterized by the following.

- A northward shift of the oceanic ITCZs, as evidenced by the further OLR decreases along 10°–12°N and the increases along 4°N.
- A tendency toward northward flow between 10°N and 12°N (somewhat to the north of the most pronounced tendency during Phase I). These changes are also linked to the northward migration of the ITCZs.
- Further SST decreases throughout the equatorial belt and the southern tropics. The distinctive equatorial signature of the Phase I pattern is not evident. The largest SST decreases are not concentrated in the upwelling regions along the equator and the coasts; instead, they are spread over broad, rather amorphous, regions ~10°S. A notable exception is the narrow region of large SST decreases along the northern coast of the Gulf of Guinea in Fig. 8. The annual march in this region has been discussed by Philander (1979) and Merle et al. (1980).

It seems unlikely that the large-scale features of the pattern of SST decreases observed during Phase II could be caused by either local or remotely forced upwelling. A possible alternative explanation is the expansion of the stratocumulus cloud decks observed during Phase II, as documented in Figs. 9 and 10. The regions of largest SST decreases during Phase II are marked by albedo increases of up to 15%, indicative of an increase in cloud cover within the stratocumulus cloud decks and some westward spreading of the decks. These increases in cloud cover are large enough to induce a substantial phase lag in the annual march of insolation reaching the sea surface. Figure 11 shows the annual march of albedo and net downward solar radiation at the top of the atmosphere, averaged over the stratocumulus cloud decks in the two oceans. Note that within these regions it is not until November, the month before the summer solstice, that the net insolation rises to values comparable to May. Differences between the values at the times of the two equinoxes are in excess of 50 W m⁻² (more than ten times as large as the differences in solar radiation incident upon the top of the atmosphere).

It is not clear what causes these increases in cloud cover during the Southern Hemisphere winter. They may be as much a result of the SST decreases as a cause
FIG. 5. The Pacific Phase I (March to May) difference in (a) OLR and vector surface wind, (b) SST and vector wind stress, and (c) pressure and vector surface wind. The OLR, SST, and pressure difference contour intervals are 10 W m\(^{-2}\), 0.5\(^\circ\)C, and 0.5 mb, respectively. The zero contour is thickened and the negative contours are dashed. Wind (wind stress) change vectors of magnitudes < 1 m s\(^{-1}\) (<10 m\(^2\) s\(^{-2}\)) are not plotted.
Fig. 6. As in Fig. 5 but for the Atlantic Phase I (April to June) difference.
Fig. 7. As in Fig. 5 but for the Pacific Phase II (May to August) difference.
FIG. 8. As in Fig. 5 but for the Atlantic Phase II (June to August) difference.
(i.e., a reflection of a positive feedback between changes in cloud cover and SST).

4. Forcing of the equatorial cooling

The results presented in the previous section suggest that the strengthening of the equatorial cold tongue during Phase I might be viewed as a remote response to wind-stress forcing. The signature of that forcing is more strongly evident in the meridional wind component than in the zonal component. The results of an analytical study by Cane and Sarachik (1977) suggest that changes in meridional wind stresses might be capable of inducing changes in equatorial SST. In particular, Cane and Sarachik showed that a distribution of meridional wind that is antisymmetric about the equator can produce a pycnocline displacement that is symmetric about the equator.

However, it does not necessarily follow that the cooling is a response to the increase in northward wind stress. Theoretical studies by Cane and Sarachik (1977, 1979), Moore and Philander (1977), and others indicate that changes in zonal wind stress on the equator are highly effective at forcing thermocline displacements at the eastern end of the equatorial waveguide, and there is observed, in fact, a weak signature in the zonal wind component on the equator (Katz et al. 1977; Meyers 1979; Wyrski 1981), particularly in the Atlantic. In addition, numerical integrations of reduced-gravity ocean models forced with observed wind fields (Moore et al. 1978; Busalacchi and O'Brien 1980, Busalacchi and Picaut 1983, Takeuchi 1988) indicate that the observed annual march of zonal wind stresses can account for many of the observed features of the annual march in the thermal structure in the eastern
Fig. 10. As in Fig. 9 but for the Atlantic (a) June to August albedo difference and (b) cold season (July–August–September) albedo.

Fig. 11. Climatological (a) albedo and (b) available and absorbed solar radiation (W m$^{-2}$) as measured by satellite for the southeastern Pacific (6°–16°S, 90°–80°W) and Atlantic (6°–16°S, 0°–10°E) oceans. The vertical lines in (b) correspond to the times of the equinoxes.
equatorial oceans. Therefore, it is worth examining, in somewhat more detail, the annual march of zonal wind on the equator and its possible role in accounting for the observed SST variations.

If changes in zonal wind stress to the west of the cold tongue are, in fact, responsible for the pronounced equatorial cooling signature during Phase I, an increase in easterly stress along the equator should be expected a few weeks in advance of the cooling. In order to document the equatorial wind-stress climatology, monthly averages of zonal wind stress and SST in the belt from 4°N–4°S are displayed in Fig. 12 in the form of a time-longitude section. Results are based on the years 1946–85 for the Pacific and 1951–75 for the Atlantic. The superimposed SST contours are based on the Sadler et al. climatology. Numerical values for selected longitudes are displayed in Table 2. In the Pacific, the annual cycle in zonal wind stress exhibits a variety of phases. If Phase I of the cooling is a response to changes in zonal wind stress, the westward stress should be expected to increase during or slightly in advance of Phase I of the cooling. Within a narrow range of longitudes around 120°W, the westward stress does, in fact, nearly double from March to May. Farther to the west, the increase in stress is observed too late in the year to account for the cooling during Phase I. A strong annual cycle in wind stress is observed in the western Atlantic, near 40°W, but the strongest increases in the westward stress are observed during Phase II of the cooling rather than during Phase I. The increases appear almost coincident with Phase I at 30°W (near St. Peter and Paul Rocks), but they are considerably weaker. Thus, for both oceans, the phase relationships are inconclusive.

Figure 13 shows corresponding results for the meridional wind stress. In both oceans, the increases in northward stress along 4°N, which are closely linked to the intensification of the ITCZ to the north, occur simultaneously with, or a month in advance of, the equatorial SST decreases in Phase I. This is the kind of relationship that was expected for the zonal wind stress.

The increase in wind stress should occur with a year-to-year regularity no less than that of the cooling that it is supposedly responsible for causing. Figure 4 demonstrated that the cooling is remarkably regular from year to year in both oceans. The upper panel of Fig. 14 shows (in a format analogous to Fig. 4) a scatterplot of the zonal wind stress in the equatorial Pacific near 120°W, the only longitude at which the stress increases early enough in the year to be considered as a possible cause of the cooling observed during Phase I. Data points for individual year-months do not cluster nearly as tightly as the corresponding points for equatorial Pacific SST, which is repeated from Fig. 4. The westward wind stress, \( -\tau^x \), increased from March to May in only 16 of the 21 years, and the year-to-year standard deviation was comparable to the mean increase (Table

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Table 2. Equatorial (4°N–4°S) Pacific and Atlantic zonal wind-stress (m² s⁻²) and cold tongue SST (°C) climatology, annual range, and standard deviation of the climatological mean monthly values about the annual mean. The Pacific cold tongue SST is calculated for the region 4°N–4°S, 104°–66°W, and the Atlantic SST is calculated for the region 4°N–4°S, 16°W–4°E.

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3). In contrast, SST in the region of the cold tongue decreased from March to May in all 31 years, and the year-to-year standard deviation was only half as large as the mean decrease.

It could be argued that the larger apparent year-to-year variability in the annual march of $-\tau^x$ over the equatorial Pacific is simply a reflection of the larger sampling variability inherent in that statistic. The number of observations in the vicinity of 120°W, where the wind stress was sampled in Fig. 14, is considerably smaller than in the region in which SST was sampled. In addition, wind stress exhibits much more day-to-
day variability than SST and is therefore subject to larger sampling fluctuations in monthly mean statistics, even if the number of observations is comparable. In order to address this concern, in the middle panel of Fig. 14 the corresponding data points for $\tau^y$ are plotted, averaged from the equator to $8^\circ$N and from $100^\circ$W to $120^\circ$W. In terms of latitude, this region corresponds to the strongest increases in $\tau^y$ during Phase I of the cooling. This belt of longitude has purposely been chosen so that the number of years sampled for $\tau^y$ is comparable to that of $\tau^x$. It is notable that an increase in $\tau^y$ was observed during Phase I in all 24 years of the record (Table 3). The standard deviation associated with the year-to-year variability of $\tau^y$ is as large as for $-\tau^x$, but it represents a much smaller fraction of the mean increase in $\tau^y$ (Table 3).

### Table 3. Wind stress and SST statistics during Phase I for selected regions. The individual monthly values are all based on at least 30 observations.

<table>
<thead>
<tr>
<th>Region</th>
<th>Climatological change</th>
<th>Year-to-year standard deviation of change</th>
<th>Number of years in which the change is in the expected sense</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pacific</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\tau^x$ 4°N–4°S, 130°–110°W</td>
<td>-8.5 m$^2$s$^{-2}$</td>
<td>8.7 m$^2$s$^{-2}$</td>
<td>16/21 (=76%)</td>
</tr>
<tr>
<td>$\tau^x$ 8°N–0°, 120°–100°W</td>
<td>25.6 m$^2$s$^{-2}$</td>
<td>7.4 m$^2$s$^{-2}$</td>
<td>24/24</td>
</tr>
<tr>
<td>SST 4°N–4°S, 104°–86°W</td>
<td>-1.4°C</td>
<td>0.8°C</td>
<td>31/31</td>
</tr>
<tr>
<td><strong>Atlantic</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\tau^x$ 4°N–4°S, 34°–26°W</td>
<td>-9.4 m$^2$s$^{-2}$</td>
<td>4.5 m$^2$s$^{-2}$</td>
<td>24/24</td>
</tr>
<tr>
<td>$\tau^x$ 6°N–0°, 16W–4°E</td>
<td>22.0 m$^2$s$^{-2}$</td>
<td>4.1 m$^2$s$^{-2}$</td>
<td>26/26</td>
</tr>
<tr>
<td>SST 4°N–4°S, 16°W–4°E</td>
<td>-2.7°C</td>
<td>0.7°C</td>
<td>26/26</td>
</tr>
</tbody>
</table>
Corresponding results for the Atlantic are presented in Fig. 15 and in the lower rows of Table 3. The areas over which $\tau^z$ and $\tau^x$ were averaged were selected on the basis of the patterns in Figs. 6, 12, and 13. As in the Pacific, $-\tau^z$ and $\tau^x$ exhibit comparable amounts of interannual variability, but the "signal-to-noise ratio" in the climatological mean annual march in $\tau^x$ is much higher than that for $-\tau^z$.

On the basis of these results, it is concluded that the remote response to the strengthening of the easterlies along the equatorial waveguide might contribute to the development of the cold tongue during Phase I. However, it is difficult to see how it could account for the relatively early timing and the remarkable regularity of the cooling of SST. The northward wind stress between the equator and 6°N exhibits an earlier and much more dependable increase by virtue of its more direct link to the developing Northern Hemisphere summer monsoon/ITCZ circulation discussed in the previous section. Even if it is capable of producing only a modest remote response in the thermocline depth and SST on the eastern side of the basin with a structure analogous to that simulated by Philander and Pacanowski (1981), it could conceivably play a crucial role in initiating the reappearance of the cold tongues. This argument will be developed further in the next and final section.

5. Discussion

We offer the following interpretation of the distinctive annual cycle in the eastern Pacific and Atlantic cold tongue/ITCZ complexes.

At the peak of the warm season in March–April, the distribution of precipitation, SST, and surface wind are much as one would expect based on sun–earth geometry. Precipitation is centered in the equatorial belt, the SST and wind fields are nearly symmetric about the equator, and only vestiges of the cold tongue are observed (and only in the Pacific), as shown in Fig. 16a.

The May–June period marks the onset of the Northern Hemisphere summer monsoon circulation over Colombia and Panama and over west Africa. In response to this stationary heat source, a pronounced northward low-level flow develops over the eastern Atlantic and Pacific, centered around 4°N. We hypothesize that the intensifying northward wind stress along the northern fringe of the equatorial waveguide induces a remote response, bringing colder water to the surface just to the south of the equator, as in the numerical experiment of Philander and Pacanowski (1981), to form (or intensify) the characteristic equatorial cold tongue. Cooling in the atmospheric planetary boundary layer above the cold tongue would cause sea level pressure to rise locally, increasing both the northward-di-
rected pressure gradient force immediately to the north of the cold tongue and the westward-directed pressure gradient force along the equator to the west of the cold tongue (Figs. 5, 6). These intensifying sea level pressure gradients would constitute positive feedbacks in the atmosphere-ocean coupled system; the former serving to further intensify the northward flow to the north of the cold tongue, and the latter intensifying the easterly

**Warm to Cold Season SST and Wind Stress Difference**

![Diagram](image)

Fig. 18. As in Fig. 5 but for the Atlantic warm season to cold season difference in SST and vector wind stress. The SST difference contour interval is 1°C.
wind component along the equator to the west of the cold tongue, both of which would promote upwelling in the cold tongue.

It is further hypothesized that in the presence of the pools of warm water off Central America and west Africa along 7°N, the enhanced northward flow out of the cold tongues might be capable of producing a westward extension of the continental convection, which becomes increasingly concentrated into oceanic ITCZs as the season wears on.

Once formed, the cold tongue–ITCZ complex tends to be self-sustaining, as suggested by Fig. 3 and documented more clearly in Fig. 17a, which shows the seasonal march of OLR over the segments of the western hemisphere that correspond to ocean in the northern tropics. The ITCZ persists through the September equinox and is still present while the convection over South America and Africa is migrating into the Southern Hemisphere, as shown in Figs. 16b and 17b. It is evident from the low-level flow pattern that the Atlantic ITCZ competes with the land convection over northeast Brazil for low-level moisture, suppressing the equinoctial rainy season that might otherwise be observed at this time. In support of this interpretation, we note that the October minus April SST difference pattern, shown in Fig. 18, is not unlike the SST anomaly patterns that Hastenrath and Heller (1977) and Moura and Shukla (1981) have identified as being associated with year-to-year variations in northeast Brazil rainfall. Moura and Shukla employed a similar spatial pattern of SST anomalies as a perturbation forcing for a general circulation model, with the resulting patterns of precipitation anomalies being remarkably similar to the October minus April rainfall patterns that can be inferred from Fig. 16 (i.e., the October climatology resembles the anomalous precipitation pattern in the “drought” experiment).

Not until well into the Southern summer monsoon season does the cold tongue–ITCZ complex, with its northward cross-equatorial flow weaken. This remarkable persistence suggests that the Southern Hemisphere continental monsoon belts may be too weak to overwhelm the meridional circulation associated with the residual ITCZ (Fig. 3a), perhaps because there are no preexisting pools of warm water in the eastern oceans along 10°S over which the land convection can easily spread westward to form an ITCZ. A weak Southern Hemisphere ITCZ has, on occasion, been observed in the eastern Pacific during the equatorial “warm season,” as in March 1989 (NOAA Climate Analysis Center Climate Diagnostics Bulletin).

The marked equatorial asymmetries in the annual mean SST distributions, with warmer water in the northern tropics, may be at least partially due to the preferred northwest–southeast orientation of the coastlines. In both oceans, the coastlines tend to parallel the trades in the Southern Hemisphere, favoring upwelling and equatorward advection of cold water, whereas they tend to block the trades in the Northern Hemisphere.

In emphasizing the influence of the American and African monsoon circulations upon the mean latitude and seasonal variations in the ITCZ, we do not wish to imply that other dynamical processes are unimportant. In this connection, it is worth noting that weak double ITCZs, analogous to those in the simulated in “aqua planet” models (e.g., Hayashi and Sumi 1986), have been observed in the central and eastern Pacific for brief periods during March and April, when the equatorial cold tongue is absent and the northward cross-equatorial flow is weakest (Kornfield et al. 1967; Hubert et al. 1969). In the central Pacific, there is little, if any, northward flow across the equator, yet the ITCZ is just as distinct as in the eastern Pacific, and it remains near 7°N year-round (Sadler et al. 1987b). At these latitudes, the belt of warm surface water associated with the north equatorial countercurrent (NECC) appears to be instrumental in anchoring the ITCZ, and the meridional profile of wind-stress curl associated with the ITCZ has, in turn, been invoked to explain the existence of the countercurrent (Wyrkki 1974; Wyrkki and Meyers 1976). Assuming that these phenomena are indeed coupled, it remains to be explained why they exhibit such a distinct preference for the Northern Hemisphere. In view of the strong zonal continuity of the ITCZ, the NECC, and the westward-propagating disturbances within them, it is not inconceivable that the equatorial asymmetries associated with the American coastline and monsoon system determine the choice of hemispheres for the ITCZ across the entire expanse of the Pacific.

The arguments put forth in this section constitute an interrelated set of hypotheses whose validity can be tested in atmospheric and oceanic general circulation models, and ultimately, in models of the coupled atmosphere–ocean system. In the atmospheric models, it will be of interest to clarify the role of the annual march of SST in forcing the annual cycle in 1) equatorial convection, 2) the latitude and intensity of the ITCZs, 3) rainfall over South America, and 4) the zonal and meridional surface wind components over the equatorial oceans. In the ocean models, further studies are needed to determine whether the observed increase in the northward surface winds in the Northern tropics during the onset of the summer monsoon might be capable of initiating the intensification of the cold tongues. Coupled models are the appropriate tool for investigating the time-dependent behavior and stability characteristics of the ITCZ/cold tongue complexes and the associated system of ocean currents.

Acknowledgments. The authors wish to thank Professors S. Hastenrath, S. G. H. Philander, and K. Wyrkki for their comments, which helped to focus the arguments in this paper. This work was supported by
the National Science Foundation under Grant GA 8822872.

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