Air–Sea Interaction in the Equatorial Atlantic Region*

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ABSTRACT

Using a dynamically motivated analysis of observations, and an intermediate-level coupled model, the interannual variability within the equatorial Atlantic is studied. It is found that a significant part of the observed variability can be described by an equatorial coupled mode akin to ENSO (El Niño–Southern Oscillation). The Atlantic mode signature is even more tightly focused on the equator and is situated proportionally farther to the west within the basin than its Pacific counterpart.

Model simulations capture the equatorial coupled mode in relatively pure form and, for what are thought to be the most realistic parameter choices, show interannual oscillations favoring a 4-year period, which are not self-sustaining. The simulated spatial patterns agree well with those extracted from observations, including those features that distinguish the Atlantic from the Pacific.

Sensitivity experiments show that the Atlantic coupled mode signal is less robust than the corresponding Pacific ENSO signal but is still well-defined qualitatively, within reasonable parameter ranges. The results demonstrate that the primary mechanisms of oscillation for the Atlantic and Pacific are the same but that differences in the zonal structure and strength of air-sea coupling and mean ocean stratification offset the large differences in basin size, allowing similar oscillation periods for the two basin modes. An explanation for the distinct spatial patterns of simulated Atlantic and Pacific anomalies is found in the differences in climatological mean fields and ocean basin configurations.

Together, the observational and model results present a picture of equatorial Atlantic variability in which coupled equatorial dynamics play an important but not exclusive role. It appears that the coupling is sufficiently strong to leave its imprint on the total variability but too weak to dictate it entirely, even at the equator.

1. Introduction

For at least two decades, large-scale air–sea interaction has been the subject of intense study in the context of the El Niño–Southern Oscillation (ENSO). As a result, much has been learned about the mechanisms of climate variability in the tropical Pacific, including the active contribution of upper-ocean dynamics. There have been many theories of how air–sea coupling in the unique setting of the tropical Pacific gives rise to ENSO-like interannual variability (e.g., Philander et al. 1984; Cane and Zebiak 1985; McCreary 1983; Suarez and Schopf 1988; Battisti and Hirst 1989; Cane et al. 1990; Jin and Neelin 1993). There are different emphases among these studies but they are more alike than different. For what is thought to be the most realistic parameter range, they all assign an important role to upper-ocean dynamics. The ocean accounts for much of the “inertia” of the coupled system and introduces temporal phase lags that help sustain interannual oscillations.

While ENSO accounts for the largest single contribution to interannual climate variability globally, it is not the only source, particularly for the extratropics and the Indian and Atlantic ocean sectors of the tropics. A number of studies have documented climatic anomalies in regions surrounding the tropical Atlantic. Specific areas that have been studied include the Caribbean (Hastenrath 1976), northwest Africa (Lamb and Pepler 1991), the Sahel (Lamb 1978a, 1978b; Lough 1986; Palmer 1986; Lamb et al. 1986; Folland et al. 1986; Wolter 1989; Lamb and Pepler 1991), the Gulf of Guinea coastal region (Wagner and da Silva 1993), the region in and around Angola (Hirst and Hastenrath 1983a, 1983b; Kousky et al. 1984), and northeast Brazil (Hastenrath and Heller 1977; Hastenrath 1978; Markham and McLain 1977; Moura and Shukla 1981; Hastenrath et al. 1984). Many of the above studies show clear relationships between regional climate anomalies and tropical basin-scale patterns of Atlantic SST, sea level pressure, and wind. In comparison, the associations with ENSO appear weaker (Hastenrath and Kaczmarszyk 1981; Lamb and Pepler 1991; Wright 1986), with the possible exception of Northeast Brazil (Covey and Hastenrath 1978; Kousky et al. 1984).

Many of the observed anomalies accompanying extreme precipitation events in the tropical Atlantic ap-

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pear as amplifications or temporal phase perturbations of the annual cycle (Hastenrath 1984). The same can be said of ENSO anomalies for the eastern Pacific region (Ramage 1977), though in the western Pacific the interannual variability is as large as the annual variability and exhibits some different features. The similarity between the two basins does not stop here, however. In some instances, such as during 1983/84, conditions in the tropical east Atlantic exhibit variations notably similar to the classical El Niño (Philander 1986).

One cannot help wondering whether the same physical mechanisms of air–sea interaction responsible for ENSO in the Pacific could give rise to the distinct climate variability in the Atlantic. Apart from the obvious differences in dimension and geometry of the two basins, there is nothing about the physics of tropical air–sea interaction that a priori singles out the Pacific basin. Of course, the differences in dimension and geometry are not minor and no doubt contribute to the distinct climatologies of the two regions. Considering only the east Pacific, however, even the climatologies bear close resemblance.

Hirst and Hastenrath (1983a,b) investigated some aspects of equatorial air–sea interaction in the Atlantic in accounting for rainfall anomalies in Angola. They found evidence for remote forcing of east Atlantic SST anomalies by zonal wind anomalies in the western basin. The importance of west Atlantic zonal winds was also highlighted by Servain et al. (1982) and McCreary et al. (1984). Such characteristics of remote forcing are of course well known for the Pacific in conjunction with ENSO (Wyrtki 1975), reinforcing the notion of dynamic similarity between the two basins. No previous study, however, has addressed Atlantic variability in the context of fully coupled dynamics.

The purpose of this paper is to explore precisely this issue for the equatorial Atlantic using marine observations and a simple model whose physics, in the setting of the tropical Pacific, describes the salient features of ENSO (Zebiak and Cane 1987; hereafter ZC). One of the questions to be pursued is, Can the same processes, operating in different settings, explain the similarities and the differences in observed variability? To this end, some relevant observational results will first be presented, followed by a description of the model, simulation results, discussion, and conclusions.

It must be stated that the present study is limited in that the dynamics of atmosphere–land surface interaction are excluded. While this may not seem a major shortcoming to studying the central Pacific, more than 5000 km from the nearest continent, the situation is clearly different for the Atlantic. Here, the potential exists for land processes to actively contribute to climate variability over the entire domain. The intention of this study is not to argue for or against such effects but rather to investigate the specific consequences of air–sea coupling in the region.

2. Data

In the following, reference will be made to observed SST and pseudostress anomalies for the tropical Pacific and Atlantic basins. Pacific SST data are derived from the Climate Analysis Center analysis (Reynolds 1988), obtained monthly on a 2° by 2° grid, spanning the period 1970–1991. Monthly anomalies are calculated using the climatological monthly means from the same source and are then subjected to a 1–2–1 filter in latitude, longitude, and time.

Pacific pseudostress data derive from The Florida State University analysis (Goldenberg and O'Brien 1981), also for a 2° by 2° grid, spanning the period 1961–1991. Monthly anomalies are subjected to smoothing and detrending as described in Cane et al. (1986).

Atlantic SST and pseudostress data derive from the analysis by Servain et al. (1987), available on a 2° by 2° grid for the period 1964–1988. Monthly SST and pseudostress anomalies are subjected to the same processing as the corresponding Pacific fields.

3. Observational results

Several authors have analyzed the variability of tropical Atlantic sea surface temperature (e.g., Weare 1977; Hastenrath 1978; Lough 1986; Servain and Legler 1986; Houghton and Tourre 1992) in terms of principal components. Time series from any of these studies reveal a moderate amplitude interannual signal. Though much smaller than observed in the tropical Pacific, this SST variability appears significant, with systematic links to regional climate anomalies (Moura and Shukla 1981; Lamb et al. 1986; Lough 1986; Palmer 1986; Wolter 1989). However, different analyses have produced different interpretations of Atlantic variability. Some, for example, have characterized a dominant mode of SST variability as a north–south dipole, while others suggest two distinct, uncorrelated modes (see Houghton and Tourre 1992; Servain 1991). Most principal components studies highlight regions of the higher-latitude tropics and show considerable variability on time scales of decades. Neither is true for the Pacific; the dominant ENSO signature is clearly focused on the equator, and its temporal variability is strongly focused at 3–5-year time scales. It is perhaps for this reason that little attention has been paid to identifying common modes of variability between the two basins.

Our understanding of the dynamics of ENSO, resulting from numerous observational and modeling studies, offers an alternative means of identifying coupled-mode signals in the Atlantic (or elsewhere). The active role played by equatorial upwelling dictates that the SST signature is largest at the equator (apart from dynamically distinct coastal regions). The thermally direct and local nature of atmospheric response near
the equator determines a well-defined circulation pattern accompanying positive and negative SST anomalies. The question of whether there exists Atlantic variability akin to ENSO can be reduced to whether these particular signatures are present.

The dynamics suggest near-equatorial SST as a good index of coupled-mode variability. We examine this first in the Pacific setting, by computing temporal correlations between the index NINO3 (area-averaged SST anomaly, 5°N–5°S, 150°W–90°W) and SST and pseudostress anomalies at all tropical Pacific locations. The particular index NINO3 is chosen because it includes the interior basin area with largest interannual variance. Figure 1 displays the spatial structure of the correlations for SST, and zonal and meridional pseudostress fields, computed over the period 1970–1991. In all instances, the familiar mature ENSO signatures are captured unambiguously. Notable features include warming of the entire eastern and central tropical Pacific, weak cooling of the west Pacific, strong central Pacific westerly anomalies, equatorward anomalies associated with displacements of both the South Pacific convergence zone (SPCZ) and intertropical convergence zone (ITCZ), and cyclonic circulation flanking the southeast Pacific pole of the Southern Oscillation. Even the vector field made up of the pseudostress component correlations is meaningful in terms of actual circulation composites or vector wind EOFs. This verifies that the analysis method is capable of achieving its goal—isolating equatorial coupled-mode signatures.

Turning next to the Atlantic, we apply the same analysis, based on this case on the period 1967–1988 (a segment of equal length to that for the Pacific analysis but slightly nonoverlapping due to the different periods of the datasets—see section 2). We define an index, ATL3, as the area-averaged SST anomaly over 3°N–3°S, 20°W–0°, again chosen on the basis of variance amplitudes. Correlations of SST and pseudostress anomalies against ATL3 are shown in Fig. 2. The similarities to ENSO patterns are striking and unmistakable. Not only is there a well-defined equatorial signature in SST and winds, but one that conforms to the same dynamical constraints as its Pacific counterpart in terms of SST structure and the relative circulation features. We interpret this as strong evidence of an Atlantic equatorial coupled mode, dynamically akin to ENSO. There are, however, some distinctions. First, notice that the correlation amplitudes are lower in the Atlantic analysis, on average by about 30%. Whereas the ENSO mode explains a large percentage of the total variability, its Atlantic counterpart explains a more modest amount. This implies relatively larger contributions of remotely forced variability or dynamically distinct modes in the Atlantic.

The Atlantic SST structure differs from the Pacific in extending across the entire basin at the equator, with no sign reversal, and in extending less far poleward, especially in the Northern Hemisphere. Note also that the coherence between equatorial and eastern coastal SST is much less in the Atlantic. Consistent with the SST differences, the Atlantic circulation features are displaced proportionally farther to the west in the basin than in the Pacific and are slightly more focused on the equator. An intriguing similarity is the cyclonic circulation in the southern latitudes of each basin, suggesting an Atlantic pressure oscillation analogous to

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**Fig. 1.** Correlations between Pacific SST and pseudostress anomalies and the index NINO3. Lower-right panel presents the zonal and meridional component correlations as vectors.
the Southern Oscillation that is linked to the equatorial coupled mode.

The time series of NINO3 and ATL3 are shown in Fig. 3. The standard deviation of ATL3 is half that of NINO3—a considerably smaller, but hardly insignificant interannual signal. ATL3 shows very little decadal variability; in fact, its variability characteristics are not decidedly distinct from NINO3, apart from slightly larger contributions from annual and seasonal time scales. Note, however, that NINO3 and ATL3 are unrelated; their correlation is −.07. The Atlantic coupled variability associated with ATL3 is not simply an extension of ENSO. This could be anticipated from the spatial structures, which by analogy with ENSO patterns are indicative of local interaction, and not merely remote forcing. The Atlantic variability described by Fig. 2 and the ATL3 index appear quite different from the more antisymmetric patterns that have been highlighted in many previous studies (e.g., Hastenrath 1978; Lamb 1978a; Servain 1991). The two are not entirely distinct, however, since the node of the so-called dipole as found in all studies is well north of the equator, in the climatological ITCZ region. Indeed, the SST pattern of Fig. 2 coincides well with the southern component of the dipole mode defined by Servain (1991). Not surprisingly, then, many of the year-to-year fluctuations of Servain’s dipole index coincide with those of ATL3. In marked contrast, however, is the substantial decadal variability of the dipole index, as compared with ATL3. The lower-frequency contributions arise largely from the Northern Hemisphere sector (north of 5°N; Servain 1991), with possibly some contribution from the equatorial Southern Hemisphere region.

Looking at individual maps of SST and wind anomalies (e.g., in Picaut et al. 1985), it is easy to find instances in which an off-equatorial monopole or a dipole pattern is dominant, and other instances in which an equatorial pattern like Fig. 2 is dominant (especially during the northern summer months). The years 1967 and 1968, associated with extremes in Sahel precipitation, primarily feature the equatorial pattern during northern summer (Lamb 1978b). A recent study by Wagner and de Silva (1993) shows that interannual variations in precipitation over the Gulf of Guinea coastal region are correlated with an SST and wind pattern almost identical to Fig. 2. Thus, the equatorial coupled mode, though not a complete descriptor, is an identifiable and potentially important component of Atlantic variability. For the present purposes of investigating equatorial air–sea interaction, it is this mode that will be the focus henceforth.

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Fig. 2. Correlations between Atlantic SST and pseudostress anomalies, and the index ATL3. Lower-right panel presents the zonal and meridional component correlations as vectors.

Fig. 3. Time series of observed NINO3 (solid line) (averaged Pacific SST anomaly, 90°W–150°W, 5°N–5°S) and ATL3 (dashed line) (averaged Atlantic SST anomaly, 20°W–0°W, 3°N–3°S).
How well do the Atlantic and Pacific coupled-mode structures account for the full variability specifically at the equator? Figure 4 shows equatorial SST anomalies for the western, central, and eastern longitudes of each basin. In the Pacific, the eastern and central regions tend to vary out of phase with the western region, whereas in the Atlantic, all longitudes tend to vary together. Pacific anomalies are small in the west, and amplified toward the east, whereas Atlantic anomalies are more nearly equal at all longitudes. These features are completely consistent with the coupled-mode signatures of Figs. 1 and 2. As noted above for ATL3, the Atlantic has more high-frequency variability. This takes the form of seasonal time scale “events” (as in 1984) superimposed on interannual oscillations. Also, extreme values tend to occur at different times in the two basins; during midyear for the Atlantic, and near the end of the year for the Pacific.

An issue not yet addressed is the evolution of “events.” By construction, our correlation analysis tends to depict the structure of mature anomalies. It cannot describe any evolution in structure—in particular, propagation characteristics. A representative picture of this aspect of coupled-mode variability is provided by equatorial longitude–time sections, shown in Figs. 5 and 6. It can be seen that east Pacific SST warmings are generally preceded by weak positive anomalies in the western region, and sometimes show slow eastward migration, as in 1982. During other events, such as 1972 and 1976, positive anomalies appear to migrate westward from the eastern boundary, so that the central Pacific warming lags that in the east; this is also apparent for the cold episode of 1988. Thus, there is no single evolution pattern with respect to longitudinal phase relations. Looking to the Atlantic, individual events also differ in structure. Several of the larger events affecting the central region (as in 1967, 1968, 1979, 1981) have little amplitude in the east, in accord with the generic coupled-mode signature of Fig. 2. There are exceptions, however—notably in 1984, when an event more like the classical Pacific El Niño occurred [though much shorter lived; see Philander (1986) and references therein]. Overall, one sees little evidence of systematic anomaly migration in the Atlantic. Rather, the variability appears more as quasi-stationary patterns, preferentially but not exclusively focused in midbasin.

Looking next to zonal wind stress (Fig. 6), eastward migration is conspicuous in the Pacific, though strong amplification near the date line introduces a large stationary component to the variability. There is comparatively very little signal in the east. This last point is equally true for the Atlantic, but otherwise there are striking differences. First, there is no migratory component to Atlantic zonal wind stress variability; the structure is stationary. Moreover, the strongest variability is situated in the western portion of the basin, as opposed to the interior basin in the Pacific case.

Fig. 4. Upper panel: equatorial Pacific SSTA at 145°E (solid line), 165°W (dotted line), and 115°W (dashed line). Lower panel: equatorial Atlantic SSTA at 40°W (solid line), 20°W (dotted line), and 0°W (dashed line).

Fig. 5. Observed equatorial SSTA as a function of longitude and time for the Pacific and Atlantic basins. Units are degrees Celsius.
Notice that despite the longitudinal differences in SST and zonal wind structure between the two basins there is a common overall SST-zonal wind relationship. For either ocean, the appearance of coherent warm (cold) SST anomalies coincides with westerly (easterly) changes in zonal wind immediately to the west. For the Pacific, however, the SST and wind patterns are strongest in the eastern and central regions, respectively, whereas in the Atlantic the SST and wind patterns are strongest in the central and western regions, respectively. The similarity in the SST-zonal wind relationship is suggestive of common dynamics, and not surprisingly, is well captured by the coupled-mode patterns derived above.

Collectively, the observational results reinforce the notion of dynamically similar coupled modes operating in the two oceans but also present interesting questions concerning the differences in positioning, time scale, and strength of interannual variability. A logical next step in addressing these issues is the application of a model embodying the hypothesized common physics. It is to this model that we turn next.

4. Model description

The model formulation parallels that of the ZC ENSO model exactly; only a brief summary and listing of specific modifications for the Atlantic will be given here. The governing equations are identical to ZC: they describe oceanic and atmospheric anomalies relative to prescribed monthly varying climatological conditions. Oceanic dynamics are represented by the time dependent, linear shallow-water equations and are augmented by a diagnostic, fixed-depth surface layer in which the entire Ekman transport is realized. Within this surface layer, a relatively complete temperature anomaly equation is carried, including horizontal and vertical advection terms for both the mean and anomalous circulation, and a simple Newtonian damping parameterization for anomalous surface heat flux. The monthly climatological surface temperature is specified, as is the climatological vertical temperature gradient and thermocline depth. The temperature anomaly of water entrained into the surface layer is parameterized in terms of thermocline displacements (identified with upper-layer thickness anomalies). Only when there is net upwelling does the entrainment temperature impact the surface-layer thermodynamics. Climatological mean surface currents (including upwelling) are derived from the same dynamical model driven by climatological surface wind stress.

The atmosphere model dynamics are described by steady, linear, shallow-water equations as in Gill (1980). The forcing is parameterized as the sum of two contributions, one relating directly to SST anomalies and another depending on the low-level convergence. The latter effect is intended to represent the feedback of moisture convergence and enhanced convective heating on the circulation (Zebiak 1986); it is nonlinear, since only when the total wind field becomes convergent does this process become active. The climatological surface wind convergence is specified from observations.

Coupling is achieved in the following manner. The ocean affects the atmosphere only through the SST anomaly field, which comprises part of the atmospheric heating as described above. The atmosphere affects the ocean only through surface wind anomalies, which are converted to stress anomalies using a quadratic bulk formula, and utilizing the specified climatological mean wind.

For the present study, the atmosphere model parameters are chosen identical to the standard case in ZC, as there is no physical basis for doing otherwise. The specified climatological mean winds are obtained from the Atlantic Hellerman and Rosenstein (1983) climatological stress by inverting the relation $\tau = \rho C_D \| \mathbf{u} \| \mathbf{u}$ and taking $\rho C_D = 0.0035 \text{ kg m}^{-3}$. The unusually large effective drag is meant to compensate for representing mean stresses in terms of long-term mean winds (i.e., to compensate for the loss of the variance contribution).

Several modifications are made to the ZC ocean model for adaptation to the Atlantic. First, the basin geometry is changed; whereas the Pacific model basin
where $T'$ is the surface temperature anomaly, $T_d'$ is a subsurface temperature anomaly, and $H_T$ is the mean depth at which $T_d'$ is evaluated. For the Pacific, $H_T$ was taken to be 100 m, corresponding to a depth slightly greater than that of the basin-mean isothermal layer. A much smaller $H_T$ is appropriate for the Atlantic; for the standard case we choose $H_T = 40$ m. As a parameterization for $T_d'$, the model uses the form

$$T_d' = \begin{cases} A_1 \left( \tanh \left( \frac{\tilde{h} + \tilde{h}}{h_1} \right) - \tanh \left( \frac{\tilde{h}}{h_1} \right) \right) & \text{if } h \geq 0 \\ A_2 \left( \tanh \left( \frac{\tilde{h} - \tilde{h}}{h_2} \right) - \tanh \left( \frac{\tilde{h}}{h_2} \right) \right) & \text{if } h < 0, \end{cases}$$

(2)

where $\tilde{h}$ is the calculated upper-layer thickness (thermocline depth) anomaly; $h_1, h_2$ are vertical scales for thermocline temperature changes; $A_1, A_2$ are temperature scales; and $\tilde{h}$ a specified mean thermocline depth (as a function of longitude). The values of $\tilde{h}$ range between 120 and 40 m for the Atlantic, as suggested in Fig. 7. This is in the same range as for the eastern half of the Pacific but considerably smaller than the 175-m maximum in the warm pool region. Table 1 lists the values of $\tilde{T}_2$ and $\tilde{h}$ for both the Pacific and Atlantic models.

The Pacific specifications for $h_1$ and $h_2$ are 80 and 33 m, the latter giving a stronger dependence of $T_2'$ on $\tilde{h}$. For the Atlantic, we take $h_1 = 80$ m and $h_2 = 40$ m as standard specifications but also examine other choices in sensitivity experiments. The values for $A_1$ and $A_2$ are set to give the same $T_d'$ for a given $\tilde{h}$ at the location where $\tilde{h} = H_T$ in each basin. The Atlantic (Pacific) values are +18 (+28) and -20 (-40). This specification amounts to assuming similar thermocline structures, despite differences in the mean position. The choice is somewhat arbitrary but reasonable; again, alternatives are examined in section 6.

Among the differences between the two models, the most significant are the equivalent depth and the surface-layer depth (and related $H_T$). Whereas the equivalent depth strongly affects the upper-ocean dynamic adjustment time, the surface-layer depth strongly affects the surface-layer thermodynamic response time. In comparison to their Pacific counterparts, the Atlantic

| TABLE 1. Specifications of mean thermocline depth ($\tilde{h}$; units in meters) and mean vertical temperature gradient ($\tilde{T}_2'$; units in °C m⁻¹) for the standard Pacific and Atlantic models, given at ten equally spaced longitudes spanning the respective basins. |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| Pacific $\tilde{h}$ | 165 | 165 | 165 | 168 | 175 | 125 | 90 | 71 | 53 | 50 |
| Atlantic $\tilde{h}$ | 120 | 100 | 85 | 75 | 65 | 55 | 50 | 45 | 45 | 45 |
| Pacific $\tilde{T}_2'$ | 0.020 | 0.020 | 0.020 | 0.018 | 0.013 | 0.029 | 0.043 | 0.055 | 0.055 | 0.055 |
| Atlantic $\tilde{T}_2'$ | 0.020 | 0.025 | 0.030 | 0.035 | 0.040 | 0.045 | 0.048 | 0.050 | 0.053 | 0.055 |

Fig. 7. Climatological annual mean temperature at the equator as a function of longitude and depth (in meters).
specifications result in a slower dynamic adjustment time but a faster thermodynamic response time. These issues will be pursued further below.

5. Results from the standard model

A coupled model run was initiated with an imposed easterly stress anomaly in the western equatorial Atlantic, held for four months and then removed. The resulting simulated ATL3 shows that rather weak interannual oscillations occur for a period of about 20 years and subsequently decay (Fig. 8). A parallel experiment with the Pacific model reveals the difference in interannual variability. As described in ZC, the Pacific model produces self-sustaining aperiodic oscillations of rather large amplitude (on average somewhat larger than observed), with a favored period of 4 years. In comparison, the Atlantic model variability, even initially (while under the influence of initial conditions), is two to three times smaller; however, the characteristic oscillation time is very similar. Experiments with numerous other initial conditions confirm these findings.

Also noticeable in Fig. 8 is that the simulated Atlantic anomalies are relatively more modulated by the annual cycle than their Pacific counterparts. This has the effect of focusing what can be identified as individual warming and cooling events into periods of duration one to two seasons, centered on northern summer/fall. In addition to being longer-lived, the Pacific “events” tend to reach maxima around the end of the year. Similar contrasting characteristics were found in observations (Fig. 4).

Additional detail of the model simulations is provided by longitude–time sections of zonal wind stress and SST at the equator (Figs. 9 and 10). The most striking contrast between the two simulations is the relative longitudinal displacement of anomalies. Whereas the model Pacific wind stress and SST anom-
Fig. 11. SSTA and wind anomalies during a cold event (upper panels) and a warm event (lower panels) of the standard case Atlantic simulation.

Fig. 12. As in Fig. 11 except for the standard case Pacific simulation.
the differences are the zonal positioning and the degree of hemispheric asymmetry. For both warm and cold episodes, simulated Atlantic SST anomalies extend only slightly north of the equator, with no corresponding reversal to the south. In contrast, Pacific anomalies are more symmetric, extending to the higher-latitude tropics of either hemisphere. The simulated Atlantic patterns are somewhat more asymmetric than the observed (Fig. 2), making the distinction between Atlantic and Pacific patterns even more pronounced than in nature. Note that the Atlantic simulations do not show strong (and unrealistic) wind reversals in the east that are characteristic of Pacific simulations. From the study of Zebiak (1990) one can deduce that the absence of appreciable eastern Atlantic easterlies (westerlies) during warm (cool) events is due to the SST anomalies being more equatorially confined. This results in a smaller excitation of atmospheric Kelvin waves relative to Rossby waves and a lesser zonal wind response to the east of the strongest forcing region.

Another important component of the coupled variability is upper-ocean heat content. As shown in many studies (e.g., ZC; Harrison 1989; Battisti 1988; Zebiak 1989; Graham and White 1988; Chao and Philander 1993; Jin and Neelin 1993; Neelin et al. 1992), fluctuations in upper-ocean heat content are both systematic and significant in the evolution of ENSO. In Figs. 13 and 14 we show time-longitude sections of upper-

![Fig. 14. As in Fig. 13 except for the standard Atlantic simulation.](image)

layer thickness (i.e., heat content) anomalies at three different latitudes (8°S, the equator, 8°N) for the Pacific and Atlantic simulations. The Pacific results are consistent with previous studies; there is evidence of westward propagation associated with Rossby waves at 8°S and 8°N, and a tendency for slow eastward propagation in the central Pacific at the equator. At all three latitudes, the signal has a strongly forced character, indicated by the phase reversals in the basin interior. This is the result of strong wind forcing in the central Pacific associated with the extremes of ENSO. There is no direct evidence of free-wave signals at the equator, the free-wave Kelvin speed being more than ten times faster than the eastward migration speed of coupled anomalies.

The Atlantic simulation shows general characteristics very similar to the Pacific; namely, the westward propagation at 8°S and 8°N, and eastward migration tendency at the equator. At 8°S and the equator, there is a distinctive stationary component to the variability, as is true for the Pacific, but the phase reversals tend to be situated proportionally farther west in the basin. This is of course consistent with the differences in positioning of zonal wind stress, as described above. Perhaps the most striking difference between the two simulations is seen at 8°N. Whereas for the Pacific the variability at 8°N closely parallels that at 8°S (a nearly symmetric pattern); in the Atlantic case the variability at 8°N is quite distinctive. In fact, the Atlantic 8°N signal has nearly a free Rossby wave character across most of the basin. This marked asymmetry between
north and south derives primarily from the structure of the background climatological mean winds, which are generally much weaker at 8°N than at 8°S over most of the basin. The mean wind structure in turn imposes an asymmetry on the wind stress anomalies (and the wind stress curl anomalies) that force fluctuations in upper-layer thickness.

Analysis of ocean mixed-layer thermodynamics provides still further insight into the similarities and contrasts between the two models. Based on the simulated SST anomalies (exemplified in Figs. 11–12), selected points in each basin are chosen for analysis. At or near the equator, three locations representing western, central, and eastern regions of the basin are analyzed in Fig. 15. For reference, we define these points as WP (the equator, 170°E), CP (the equator, 140°W), EP (1°S, 100°W), WA (the equator, 32°W), CA (1°S, 16°W), and EA (1°S, 0°). At each point the SST tendency is shown for zonal advection, meridional advection, advection by mean upwelling, advection by anomalous upwelling, and anomalous heat flux, over a six-year segment. In these models, the anomalous heat flux is proportional to the negative of the SST anomaly, so that the flux anomaly can conveniently be used to infer the SST anomaly. Considering first the west Pacific (WP), it is clear that by far the most important term is zonal advection (the approximate quadrature relationship between zonal advection and SST anomaly or heat flux shows that zonal advection is forcing the temperature changes). The simulated western Atlantic (WA) is in marked contrast, with large contributions from both mean and anomalous upwelling advection terms, as well as zonal advection. It is noteworthy that mean upwelling advection ($\bar{w}T^*_z$) acts effectively as a damping for both WP and WA, but with proportionally larger amplitude in the Atlantic case.

The analyses of CP and CA are generally similar, showing large contributions from both upwelling terms, and significant contributions from zonal advection. In both simulations, anomalous upwelling is a strong forcing of SST, and mean upwelling acts primarily as a damping but again with proportionally greater magnitude in the Atlantic.

The variability of EP is forced primarily by mean and anomalous upwelling. Zonal and meridional advection are important only occasionally, during the rapid decay of a warm event. As opposed to regions farther west, the SST variability in the EP region is significantly forced by the mean upwelling term (which depends on the anomalous temperature gradient; see Eqs. (1)–(2)) and therefore involves upper-ocean heat content and large-scale ocean dynamics. The forcing of EA also primarily involves the upwelling terms, with a noticeable contribution from meridional advection (due to the maximum mean upwelling being slightly south of EA; this term becomes very small at the upwelling maximum). As for EP, the mean upwelling term for EA is such as to force SST changes; that is, it has a systematic phase difference with SST. The magnitude of this forcing, however, is proportionally much smaller at EA. The forcing due to anomalous upwelling is much more dominant for EA than for EP.

For each basin, two other regions are examined, one in the central region north of the equator (NP at 8°N, 150°W; NA at 6°N, 28°W) and another in the eastern basin south of the equator (SP at 8°S, 110°W; SA at 8°S, 10°W). As noted above, the Atlantic SST patterns show greater asymmetry about the equator than do the Pacific patterns. Looking first at NP (Fig. 16), we find a rather simple balance in which meridional advection is overwhelmingly dominant. This reflects the mean northward Ekman drift advecting SST anomalies northward from the upwelling source regions near the equator. The situation for NA shows a significantly different balance, in which a similar meridional advection forcing is almost exactly offset by zonal advection forcing. Resultant SST change is very small (and therefore not coherent with variability at the equator).

The two southern points, SP and SA, show very similar balances, with meridional advection the dominant term, followed by anomalous upwelling and zonal advection. The similarity of the balances south of the equator is consistent with the general similarity in SST patterns for Pacific and Atlantic simulations.

6. Parameter sensitivities

Many of the standard model specifications are carried over directly from the Pacific model, and the success of the latter in simulating and predicting ENSO (see Cane et al. 1986; Barnett et al. 1988) can be taken as adequate evidence in support of them. Included in this group are the atmospheric model parameters, surface momentum and heat exchange coefficients, and background dissipation within the upper ocean. Other parameters have been assigned different values to account for the considerable differences in oceanic mean stratification between the two basins. The exact choices for some parameters must be considered arbitrary; that is, there is no rigorous means to justify a particular choice within a range of uncertainty. The question then becomes how sensitive the results are to such parameters, within their uncertainties. For the Pacific version of the model, it is the case that the simulated ENSO characteristics are very robust with respect to modest changes in model parameters (ZC; Zebiak and Cane 1989). The following set of experiments was designed to address the same question for the Atlantic model.

The first experiments examine not the sensitivity to parameters but the sensitivity to initial conditions. To what extent is the variability shown in Fig. 8 representative of the possible range of behaviors? This is indicated by a set of four simulations with the standard model, beginning from different initial states (Fig. 17).
The initial ocean and atmosphere states were obtained by forcing the ocean component of the model with observed wind stresses from 1967 to some chosen time, and separately forcing the atmospheric component with the resulting SST simulations over the same interval. Modest to small amplitude oscillations of period 3–5 years are present in each of the simulations, as in the standard case. The largest difference among them is the time period over which the oscillations remain at finite amplitude; this ranges from less than 10 to about 40 years. Clearly, such behavior does not represent the pure exponential decay of a linear, homogeneous system. It may, however, depend in part on the linear, nonhomogeneous terms associated with the seasonally
varying background state acting on the anomaly fields; certain initial anomaly states may be more favorably configured to extract energy from the time-varying background state than others. The spatial structures of anomalies in all cases are indistinguishable from those shown previously. Except where noted, this is also true for the parameter sensitivity experiments, and we will hereafter restrict attention to the temporal domain.

Among the more poorly determined parameters is $H_T$, the depth over which the near-surface vertical temperature gradients are evaluated. In a test experiment, the standard model value of 40 m was increased to 65 m (Fig. 18). The result (cf. Fig. 8) is a significant initial increase in amplitude but negligible change in frequency of interannual oscillations. Despite the initially larger amplitude of variability, finite amplitudes still cannot be sustained indefinitely; in this instance, the oscillations fade out after about 30 years. It is instructive to compare with the Pacific model sensitivity to the same parameter. Figure 19 shows results from a Pacific simulation in which the standard model $H_T$ was reduced from 100 to 71.4 m (i.e., a change in the same proportion as in the Atlantic experiments). Comparing with Fig. 8, one sees that the reduction in $H_T$ (cf. Fig. 8) leads to an increase in amplitude, no change in preferred frequency, and possibly a greater degree of periodicity. The strongest response to changing $H_T$ in either model is the change in amplitude, but the sense of the response is opposite. Though perhaps counterintuitive, this result is not entirely a surprise in light of the thermodynamic balances presented earlier.

It was found that for the western and central sectors, the SST tendency associated with the mean upwelling term (in which $H_T$ comes into play) was much more strongly damping in the Atlantic setting compared to the Pacific. Moreover, in the eastern basin, the mean upwelling tendency strongly forces SST in the Pacific, whereas it does so only very weakly in the Atlantic. The effect of increasing (decreasing) $H_T$ is to reduce (increase) the tendency associated with the mean upwelling [Eq. (1)]. For the Atlantic, the reduction of a primarily negative feedback results in an increase in the amplitude of variability. For the Pacific, the reduction (increase) of the strong eastern basin forcing evidently overcomes the reduction (increase) in damping in the west, thus weakening (strengthening) the oscillation. This is more a consistency argument than an explanation, but it serves to shift the focus to
Since the mean thermocline depths ($\bar{h}$) are less [see Eq. (2)]. Thus, this experiment imposes a very large dependence of $T_d$ on $h$. The results show that interannual variability of moderate amplitude can in this case be sustained (Fig. 20a). The favored period is 4 years, as in the standard model. Also, the anomalies tend to be shifted slightly toward the east (not shown).

Fig. 17. Simulated ATL3 from the standard Atlantic model but with alternate initial conditions.

Fig. 18. Model ATL3 from an experiment identical to the Atlantic standard case except for 60% increase in $H_T$ (see text).

Fig. 19. Model NINO3 from an experiment identical to the Pacific standard case, except for 60% decrease in $H_T$ (see text).

Fig. 20. ATL3 from Atlantic test experiments in which (a) $T_d$ was made identical to the Pacific standard case, (b) the parameter $A_k$ was changed from 20 to 26, (c) the equivalent depth was increased from 40 cm to 57 cm, and (d) all upwelling terms in the ocean surface-layer temperature equation were decreased by 20%.
The second experiment is less extreme: only the coefficient $A_2$ is increased, from 20 to 26. The results show no obvious differences with the standard model (Fig. 20b). The conclusion is that modest changes in $T_d$ parameters produce little qualitative change in the results and that only very large adjustments—seemingly beyond what is reasonable—allow for self-sustaining interannual oscillations.

Another experiment examines the dependence on oceanic equivalent depth. Previous work assigns a critical role to wave dynamics in accounting for the existence and the time scale of ENSO. Zebiak and Cane (1987) showed with the present Pacific model that both the amplitude and period of interannual oscillations increase (decrease) as the equivalent depth decreases (increases), though not linearly. For the Atlantic, we perform a similar experiment by increasing the equivalent depth from 40 to 57 cm. The latter value corresponds to estimates for the first internal mode, whereas the standard value is intermediate between the first and second mode. The result of an increased equivalent depth (Fig. 20c) is completely consistent with the Pacific experiments, showing a reduction in amplitude and a decrease in preferred period from 4 to 3 years. Not surprisingly, the oscillations diminish more rapidly than in the standard case. These results suggest that ocean wave dynamics play the same important role in generating and regulating oscillatory behavior in the Atlantic setting as in the Pacific. The thermodynamic balances for the EP and EA regions suggested likewise.

An experiment was run in which all upwelling-related terms in the surface-layer temperature equation are reduced by 20% (Fig. 20d). The sensitivity is very low; there is a hint of a reduction in amplitude but no other detectable change.

Finally, some experiments were run to test the sensitivity to the specified surface-layer depth. While the choice of 30 m was well motivated by the climatological mean thermal structure, the value is by no means precisely determined. In one experiment, the depth was increased to 40 m (somewhat closer to the standard Pacific value of 50 m). The climatological mean surface currents were first recomputed for a 40-m-layer depth and then the corresponding change was made in the anomaly calculation. Increasing the surface-layer depth had the effect of weakening the coupled variability by roughly a factor of 2 (not shown), but it did not change the time scale of variability.

7. Discussion

In both observations and model simulations, systematic differences are found in the spatial structure of Atlantic and Pacific anomalies. Near the equator, Atlantic anomalies appear shifted relatively toward the west as compared to their Pacific counterparts. The model provides a framework for understanding this Atlantic–Pacific contrast. Consider first the mean surface wind (or wind stress) fields (Fig. 21). In the Pacific, equatorial easterlies are strong throughout the middle third of the basin, whereas in the Atlantic, they are strongest in the western third. As equatorial upwelling is sensitive to zonal wind stress, the pattern of Atlantic upwelling is similarly displaced toward the west of the basin (the present model’s estimate is shown in Fig. 22). The differences in near-equatorial thermocline tilt (Fig. 7) also reflect the distinct zonal wind structures: the Atlantic thermocline slopes most in the western basin, and the Pacific thermocline slopes most in the eastern central basin. SST structures are similar in the eastern basins but feature the largest zonal gradients relatively farther west in the Atlantic (Fig. 23). Note also the absence of a very warm pool in the western Atlantic, consistent with the stronger upwelling, stronger winds (evaporation), and shallower thermocline that characterize the western part of this basin, relative to the Pacific.

Stronger mean wind implies a larger stress anomaly for a given wind anomaly. Stronger upwelling implies

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**Fig. 21.** Climatological annual mean pseudostress for the Pacific and Atlantic basins.
a larger SST anomaly for a given subsurface temperature anomaly. Sharper SST gradients imply a larger SST anomaly for a given surface current anomaly. All of these contribute to the net strength of coupling between atmospheric and oceanic anomalies; it follows that the Atlantic configuration of mean fields favors strong coupling relatively farther west within the basin. This accounts for the different zonal structure of anomalies in the model and, by implication, in nature.

Such differences in the locations of strong coupling have wider implications for coupled dynamics. The basis for the delay in the “delayed oscillator” mechanism proposed for the Pacific is the time it takes for upper-ocean thickness anomalies generated in the central basin to propagate to the western boundary and reflect as equatorial Kelvin waves (e.g., Battisti 1988; Schopf and Suarez 1988; Cane et al. 1990). If the generation region for such anomalies is already near the western boundary, as for the Atlantic, then the expected delay will be shorter. One might expect in this case that the preferred oscillation period would decrease. Even accounting for the smaller equivalent depth appropriate for the Atlantic, consideration of linear ocean dynamics alone would suggest much shorter oscillation periods for Atlantic variability. However, as pointed out by several authors, the characteristics of coupled modes are not necessarily simply related to the free-mode characteristics of either the atmosphere or ocean. In this instance, the quicker transition from positive to negative feedback associated with thermocline effects must be balanced against larger positive feedbacks from anomalous upwelling/downwelling or zonal advection. In fact, as seen above, the overall balance of terms in the Atlantic setting results in a preferred oscillation period similar to the Pacific.

The thermodynamic analyses presented in Figs. 15 and 16 reinforce several of these points. The largest single difference between Atlantic and Pacific simulations was found in the western basins (WP and WA). The stronger variability in the western Atlantic can now be understood in terms of the stronger mean winds, SST gradients, upwelling, and shallower mean thermocline depth that characterize that basin. Notice that the SST forcing associated with anomalous up-
welling is proportionally stronger in both the western and eastern sectors of the Atlantic, compared with corresponding regions of the Pacific. This indeed implies stronger positive feedbacks associated with these terms. The damping associated with mean upwelling is also stronger in the Atlantic (western and central regions), primarily because of the shallower mean thermocline, which gives rise to larger subsurface temperature anomalies ($T_\sigma$) associated with a given thermocline depth or upper-layer thickness anomaly [Eq. (2)].

In view of the differences just described, it is important to emphasize that ocean wave dynamics still play a crucial role in supporting interannual oscillations in the Atlantic, as in the Pacific: wave dynamics provide the primary source of phase leads/lags that characterize the oscillations and, in particular, the means of transition between extreme states. These common aspects of Atlantic and Pacific variability can be seen by comparing indices of zonally integrated equatorial heat content and SST anomalies for each basin (Fig. 24). The zonal integrals average over those features that distinguish the two basins, thereby emphasizing their common features. For either basin, time series of equatorial SST, zonal wind, thermocline tilt, and most other variables are very closely aligned (not shown), in accordance with a positive feedback relationship. The presence of an oscillation demands, however, that at least one component of the system exhibit different phase. That component, for either Pacific or Atlantic, is found in the integrated equatorial heat content, which is controlled by ocean wave dynamics. Note that for each basin, this index of heat content is approximately in quadrature with zonal mean SST anomaly (and all the other variables synchronized to it), as would be expected for the two components of the simplest linear oscillator.

As shown by Neelin (1991) and Jin and Neelin (1993), coupled oscillations can be sustained without the influence of ocean wave dynamics, in the form of what they define as an SST mode. These modes depend on the time derivative of the SST equation and in pure form require propagation in order to give oscillation. Jin and Neelin (1993) suggest that the most realistic parameter range is one in which contributions from both SST modes and ocean dynamics are important. To test this for the present model, we followed the procedure devised by Neelin (1991), in which the time derivative of the ocean dynamics equations is artificially increased. This moves the system toward the so-called fast-wave limit, in which pure SST modes exist (while maintaining the same oceanic equilibrium response to a fixed forcing). To complement this procedure, we also conducted experiments in which the time derivative of the SST equation was artificially increased—moving the system toward the fast-SST limit that defines the pure delayed oscillator (e.g., Cane et al. 1990). Figure 25 shows 20-year simulations of ATL3 for the standard Atlantic simulation, a $4 \times$ standard wave speed experiment, and a $4 \times$ standard SST tendency experiment. These and other similar calculations show that both wave dynamics and SST mode physics contribute to the overall time scale; the time scales are decreased in both distorted physics experiments. How-

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**FIG. 24.** Indices of SST (solid lines) and upper-ocean heat content anomalies (dashed lines) from the standard Atlantic and Pacific simulations. For either ocean basin, indices represent averages between $3^\circ$N and $3^\circ$S and over all longitudes.

**FIG. 25.** Time series of ATL3 for years 1–20 of (a) the standard Atlantic simulation; (b) a simulation with the oceanic wave speed (i.e., the time derivative on ocean dynamics equation) artificially increased by a factor of 4; and (c) a simulation with the time derivative of SST artificially increased by a factor of 4 (see text).
ever, the effect is larger for the fast-wave speed experiments, showing that the standard model is somewhat closer to the fast-SST (delayed oscillator) limit than the fast-wave limit.

Turning next to the issue of meridional structure, it was shown earlier that the more equatorially asymmetric structure of simulated Atlantic SST anomalies, relative to the their Pacific counterparts, results from zonal advection offsetting the tendency due to meridional advection at around 6°N (Fig. 16). For the Pacific, the zonal advection term is much less important, allowing meridional advection associated with the mean Ekman drift to spread equatorial anomalies both northward and southward. The structure of the climatological wind stress and the mean ocean stratification are significant factors, in that together they give rise to a stronger simulated north equatorial countercurrent in the Atlantic (not shown). This strong eastward mean current tends to disrupt SST anomalies originating by other means, in particular, those advected northward from the equator in the basin interior. The differences in mean wind stress and other fields to which we have appealed, here and elsewhere, pose very interesting questions of their own. This issue is beyond the scope of the present study, but one would presume the external parameters of basin size and land/sea configuration to be significant factors.

The simulated Atlantic variability in what we believe are the most realistic parameter ranges is not self-sustaining, yet it can persist at finite amplitude for as much as a few decades from a single initial perturbation. What are the implications for interpreting the observed variability? If the model results are indicative of reality, then it means the observed variability represents a synthesis of local equatorial coupled dynamics and remotely or separately induced forcing. It is evident that very modest external perturbations would be sufficient to ensure a realistic level of interannual variability via the coupled dynamics explicitly described by the model. There is ample observational evidence of such external effects. First, the correlations between Atlantic variables and equatorial Atlantic SST are lower than similar correlations among Pacific variables, suggesting that Atlantic variability represents less of a pure equatorial coupled mode; that is, that there is more variability of separate origin. Second, several studies have identified basin-scale, very-low-frequency (i.e., decadal) patterns that, by their geographical extent and focus in the higher-latitude tropics, are necessarily distinct from the equatorial modes explicitly described here. Third, there is evidence of weak but finite remote forcing due to ENSO. Finally, there are the potential effects of interactions with land surfaces, which for the Atlantic can hardly be ruled out. The presence of any or all of these effects could account for the differences between observed variability and what can be attributed to equatorial coupled dynamics, or what is simulated by the present model.

8. Summary and concluding remarks

The basic characteristics of ENSO are recognized as the expression of an equatorial coupled mode of variability whose existence and strength derive from the conditions particular to the tropical Pacific. Motivated by the general similarities between the tropical Atlantic and eastern Pacific, we have examined the interannual variability of the Atlantic with an eye toward coupled dynamics. Both observational and model results argue for the existence of an equatorial coupled mode in the Atlantic, akin to ENSO, though much weaker and with a few distinguishing features.

The key to isolating an Atlantic equatorial coupled-mode signal observationally is to utilize the known structure of such modes, which features a zonally varying, equatorially focused SST anomaly pattern. Knowledge of the form allows the straightforward construction of a modal index. Then, simple correlations between the index and wind or SST anomalies at all locations reveal the full coupled-mode signature. Not surprisingly, this analysis recovers the familiar ENSO structure when applied to Pacific data. However, when applied to Atlantic data, a strikingly similar pattern emerges. The temporal variability of the Atlantic modal index captures much of the activity within the equatorial Atlantic and is primarily interannual. It is distinct from the decadal-scale variability that has been identified in some studies (e.g., Servain 1991), especially in not extending to the higher-latitude tropics.

The equatorial pattern has been independently discovered by Wagner and da Silva (1993), in conjunction with precipitation extremes in the Gulf of Guinea coastal region. SST patterns shown by Lamb (1978b) for the particular wet and dry years 1967 and 1968 in the Sahel also resemble the equatorial pattern. Such correspondence could be coincidental, of course, and requires much more careful study. The decadal trends in Sahel rainfall appear to be related to a much-larger-scale SST pattern (Folland et al. 1986; Palmer 1986), which is apparently of separate origin. Relationships to other regional climate anomalies, such as those of northeast Brazil, the Caribbean, Angola, and the Congo basin (Hastenrath 1984) are not established, but neither are they ruled out. They, too, are deserving of further study.

The tropical Atlantic differs from the tropical Pacific in that it has proportionally more variability not attributable to the equatorial coupled mode. One aspect of this is the lower frequency, tropical basin-scale patterns. Additional contributors may be land surface interactions and global-scale forcing related to ENSO.

In the second part of this study, we have utilized a coupled Atlantic model that closely parallels that previously used to study ENSO. The model structure, and many of the parameters, could be carried over to the Atlantic setting directly. However, the specified mean fields (obtained from observations), and the model
parameters relating to the mean ocean stratification, differ from their Pacific counterparts. The standard version of the model, initialized with a west Atlantic zonal wind anomaly, produces modest interannual oscillations of preferred period 4 years, which maintain finite amplitude only for about 20 years. Experiments with different initial conditions maintain finite amplitude variously between less than 10 and about 40 years. In all instances, even the initial variability is severalfold smaller than ENSO, featuring SST anomalies on the order of 1°C. This result is in accord with observations. Also similar to observations is the tendency for individual warm and cool episodes in the Atlantic to be shorter-lived than in the Pacific and to attain largest amplitude during northern summer as opposed to northern winter.

In spatial structure, there are two notable differences between model Atlantic and Pacific anomalies: Atlantic anomalies are focused proportionally farther west in the basin, and they are less coherent from the equator northward. Both features are seen in observations as well; the first by examining longitude-time sections of SST and zonal winds along the equator of each basin, the second by examining the structure of correlations with equatorial SST indices. We were able to interpret the results using the model framework, finding that the zonal shift in anomalies owes to differences in the zonal structure of mean winds, upwelling, and SST between the two basins. The most significant contributor to the less coherent meridional structure in the Atlantic is the greater influence of zonal advection in the north equatorial countercurrent region, acting to disrupt SST anomalies advecbed northward from the equatorial upwelling zone. The north-south asymmetry in the model SST anomaly pattern is stronger than in nature, suggesting that the zonal advection effect is overstated in the model.

Several sensitivity experiments were conducted with the Atlantic model. Generally, it was found that the oscillation characteristics were somewhat less robust than was found for the Pacific, but for modest perturbations in several variables, behavior qualitatively similar to the standard case was obtained. With respect to at least one parameter, the sensitivities in Atlantic and Pacific models was opposite. We relate this to a more general difference in the balance of positive and negative feedback terms in the two models' oscillations, imposed by the more westward shift in the strongest coupling region for the Atlantic and the different relative configurations of mean thermocline depth, upwelling, and SST. Despite these differences, the primary mechanism underlying oscillatory behavior in the Atlantic model is the same as in the Pacific: the so-called delayed oscillator mechanism, in which ocean dynamics introduces a phase-lagged component into an otherwise tightly coupled and synchronous set of variations along the equator. In addition, time scales and processes associated with SST change were found to have some impact on the oscillation characteristics, making the most accurate description of the model behavior that of a mixed SST–ocean dynamics mode (Jin and Neelin 1993). The fact that in the Atlantic setting the preferred period of oscillation is similar to the Pacific, despite the large differences in basin size and wave propagation speeds, attests to the distinct character of coupled modes relative to free modes of the ocean (or atmosphere) alone.

In interpreting the Atlantic results, the picture that emerges is one in which the resultant variability is made up of partly internal and partly external processes. The internal part, evident in near-equatorial observations, and simulated in pure form by the model, behaves as an equatorial coupled mode akin to ENSO. Such a mode has its own intrinsic spatial patterns and time scales, but our results suggest that for the Atlantic this mode is insufficiently strong to determine the total variability by itself. By implication, the total variability is achieved only through the synthesis of this coupled mode, other dynamically distinct modes, and external or remotely forced perturbations. The elucidation of other such internal modes and/or sources of external forcing remain as important and interesting questions for future study.

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