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Tropical Atlantic Variability: Patterns, Mechanisms, and Impacts

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This chapter reviews the progress made in the past decade in understanding tropical Atlantic climate variability. In addition to an equatorially anti-symmetric seasonal cycle forced directly by the seasonal march of the sun, Atlantic sea surface temperature (SST) displays a pronounced annual cycle on the equator that results from continental monsoon forcing and air-sea interaction. This cycle interacts with and regulates the meridional excursions of the Atlantic intertropical convergence zone (ITCZ). On interannual timescales, there is an equatorial mode of variability that is similar to El Nino/Southern Oscillation (ENSO) in the Pacific. This Atlantic Nino is most pronounced in boreal summer coinciding with the seasonal development of the equatorial cold tongue. In boreal winter, both ENSO and the North Atlantic Oscillation exert a strong influence on the northeast trades and SST over the northern tropical Atlantic. In boreal spring when the equatorial Atlantic is uniformly warm, anomalies of cross-equatorial SST gradient and the ITCZ are closely coupled, resulting in anomalous rainfall over northeastern Brazil. There is evidence for a positive air-sea feedback through wind-induced surface evaporation that organizes off-equatorial SST anomalies to maximize their cross-equatorial gradient. The resultant anomalous shift of the ITCZ may affect the North Atlantic Oscillation, helping to organize ocean-atmospheric anomalies into a pan-Atlantic pattern.

1. INTRODUCTION

The Atlantic Ocean is flanked by two large tropical continents, which host major centers of atmospheric convection. As early as 320 years ago, *Halley* [1686] recognized the important influence of these continents on climate in the Atlantic sector and suggested that the intense surface heating over North Africa drives the southerly winds in the Gulf of Guinea. It was not until 1970s, however, that the influence of the tropical Atlantic Ocean on continental climate variability began to come to light. The studies that followed showed that interannual variability in rainfall over the semi-arid regions of South America and Africa is associated with well-organized, repeating patterns of sea surface temperature (SST) and trade wind anomalies over the tropical Atlantic. Furthermore, these patterns of ocean and atmospheric anomalies are so arranged that their interaction gives rise to positive feedback acting to amplify each other. Rapid progress has been achieved in the past decade in understanding these air-sea interaction mechanisms and modeling the resulting variability in climate over the tropical Atlantic and beyond. This chapter reviews the progress in describing the patterns and understanding the mechanisms for tropical Atlantic variability (TAV).

We begin with a brief overview of the seasonal cycle, which dominates tropical Atlantic variability. This is followed by a survey in Section 3 of interannual variability in the equatorial Atlantic, which is akin to the El Nino and Southern Oscillation (ENSO) phenomenon in the Pacific.

Section 4 concerns off-equatorial SST variability regarding which opposing views exist. We review recent efforts to understand air-sea interaction from the oceanic, atmospheric, and coupled points of view. The tropical Atlantic is not isolated, but is influenced by, and may influence climate variability in other regions, in particular ENSO and the North Atlantic Oscillation (NAO). The NAO is of central importance for climate variability in the extratropical North Atlantic and Europe. Extensive literature on NAO research exists, recently summarized in an American Geophysical Union monograph by *Hurrell et al*. [2003]. Sections 5 and 6 discuss how the NAO and ENSO influence TAV, respectively. Section 7 is a summary and includes discussion of the challenges ahead.

2. SEASONAL CYCLE

SST in the eastern equatorial Atlantic is dominated by the annual cycle. Temperatures reach their maximum in boreal spring when the equatorial winds are weakest and the thermocline is deepest in the east. During this season the sun is directly overhead, providing maximum incident solar radiation. The band of high SSTs exceeding 27^{o}C occupies an equatorial region extending from 8^{o}S to 5^{o}N (Fig. 1a). As the year progresses the trade winds along the equator intensify. The resulting zonal pressure gradient in the ocean and associated uplifting thermocline leads to seasonal cooling of SSTs in the eastern equatorial Atlantic. The SSTs reach their minimum along the eastern coast of Africa in July as a result of intensified coastal upwelling (Fig. 2a), and then in the southeastern Gulf of Guinea a month later. In July and August, a distinct cold tongue forms across the basin, centered slightly south of the equator (Fig. 1b).

The northeast and southeast trade wind systems meet at the narrow, roughly zonally oriented intertropical convergence zone (ITCZ). The time-mean latitude of the ITCZ and the collocated rain band, often called the thermal equator or climatic axis of symmetry, is displaced 5-10 degrees north of the geographical equator over the Atlantic [*Hastenrath*, 1991; *Mitchell and Wallace*, 1992; references therein], despite the fact that solar radiation at the top of the atmosphere is nearly symmetric about the equator on annual mean (see *Xie* [2004] for the latest review of research on this climatic asymmetry over the Atlantic and eastern Pacific). The ITCZ is also associated with the latitude of minimum seasonal variance of SST and the latitude of maximum vertical displacement of the thermocline [*Houghton*, 1991].

The ITCZ and its associated band of continental convection display large seasonal excursions over the Atlantic sector. Over the continents, the rain band largely follows the seasonal march of the sun, reaching its northernmost (southernmost) position in July-September (December-February) [*Mitchell and Wallace*, 1992; *Biasutti et al.*, 2003]. Since dry soil has a negligible heat capacity, the apparent lag in the meridional excursion of the continental rain band behind the sun may result from other heat reservoirs such as soil moisture and oceanic influences. For example, northeastern Brazil is in its wet season at the spring equinox (March) but is kept dry at the fall equinox (September) as strong northward SST gradients prevent the oceanic ITCZ from moving south of the equator [*Fu et al*., 2001].

Over the ocean, patterns of SST and the position of the ITCZ are tightly coupled, with major rainfall confined to a band of high SSTs above 27^{o}C. In March-April, the rain band is located nearly on the equator onto which the trades converge from both hemispheres. SST is uniformly warm in the equatorial zone of 10^{o}S-5^{o}N, making March-April the time when the Atlantic ITCZ is very sensitive to even small changes in interhemispheric SST gradient (Section 4). As the equatorial cold tongue develops in June and persists through September, the ITCZ is kept north of the equator following the northward movement of the high-SST band while a second, smaller, convective zone develops south of the equator west of 30^{o}W [*Grodsky and Carton*, 2003]. The oceanic ITCZ reaches its northernmost position in September, lagging its northward movement over the continents because of the large heat capacity of the ocean mixed layer. In July-August, rainfall in the ITCZ is considerably stronger than in March-April, despite a 1^{o}C drop in SSTs beneath the ITCZ (Fig. 2). This strengthening of ITCZ convection may be due to the abundance of strong westward propagating easterly wave disturbances that help trigger convection over the ocean. These disturbances originate from the African rain band [e.g., *Thorncroft et al*., 2003] and grow in the tropical Atlantic, some into tropical storms and hurricanes that devastate the Caribbean and southern United States [*Gray and Landsea*, 1992]. In addition to the 3-9 day African easterly waves the tropics also support a nearly stationary pattern of summer winds and precipitation with periods of two weeks that appears to result from land-atmosphere interaction [*Janicot and Sultan*, 2001; *Grodsky and Carton*, 2001].

Along the equator SST varies with a strong annual cycle despite the primarily semiannual nature of solar heating (this is also true of the eastern Pacific). At 10^{o}W, equator, SST reaches 28^{o}C March-April and drops to below 23^{o}C in July-August. This seasonal warming and cooling is highly asymmetric, with the latter taking only three months and the former taking seven months. From the oceanic point of view the rapid cooling is attributed to the sudden onset of the West African monsoon and the rapid intensification of the southerly winds in May-June in the Gulf of Guinea. These southerly winds cause upwelling slightly south and downwelling slightly north of the equator, and this upwelling cools the equatorial ocean [*Philander and Pacanowski*, 1981]. They also have strong convergence/divergence, decelerating over the cold tongue and then accelerating again over the warmer water a few degrees north of the equator. Equatorial zonal wind variations also play an important role in the equatorial SST annual cycle by inducing upwelling (through zonal and meridional divergence) and tilting the thermocline depth on the equator. From April to August, the thermocline shoals more than 60 m in the equatorial Gulf of Guinea [*Houghton*, 1983; *Philander and Pacanowski*, 1986]. Changes in zonal wind strength also affect wind-induced evaporation.

From the coupled point of view, *Xie* [1994] shows that the northward displacement of the climatological ITCZ is the ultimate cause of the annual cycle in equatorial SST in both the Pacific and Atlantic by maintaining southerly cross-equatorial winds that intensify in boreal summer/fall and relax in boreal spring [see also *Giese and Carton*, 1994]. *Mitchell and Wallace*'s [1992] observational analysis suggests that the annual cycle in the equatorial Pacific is initiated in the east by the seasonal monsoonal winds and propagates westward as the result of air-sea interactions. In contrast to the equatorial Pacific where air-sea interaction is the leading mechanism for the annual cycle, the narrow width of the tropical Atlantic and the presence of strong continental convective zones mean that continental monsoons play a much more important role. Atmospheric general circulation model (GCM) experiments show that the seasonal variations in the cross-equatorial winds in the Gulf of Guinea are mostly due to the continental monsoon [*Li and Philander*, 1997]. The annual cycle in equatorial zonal wind is driven both by the continental monsoon and by the interaction with equatorial SST, mechanisms that are important in the eastern and western half of the basin, respectively [*Okumura and Xie*, 2004]. In an experiment that removes the seasonal development of the equatorial cold tongue, anomalous easterlies still appear in May and June in the eastern equatorial Atlantic as a result of the increased cross-equatorial advection of zonal momentum and a redistribution of monsoonal rainfall.

3. EQUATORIAL VARIABILITY

Superimposed on these primarily annual variations of SST are anomalies during the boreal summer months (JJA) that frequently exceed 1^{o}C during the peak month. The warm anomalies are generally maximum in the zone of the boreal summer cold tongue between 6^{o}S and 2^{o}N, and between 20^{o}W and 5^{o}E (corresponding cool anomalies are less geographically oriented and less limited to the boreal summer). During some years, but not all, the warm anomalies appear along the southwestern coast of Africa as well. The period of the warm events appears to be approximately 30 months with 13 such warm events having occurred in the 40-year period since 1961 ('63 '66 '68 '73 '74 '81 '84 '87 '88 '93 '96 '97 '99). This past summer of 2003 provides just the most recent example.

The first well-documented event in 1963 received attention partly because of its magnitude, and partly because it coincided with the EQUALANT observational program [*Katz et al*., 1977; *Merle*, 1980]. The coincidence of warming sea surface temperatures, a relaxation of the trade winds and shifts in convection during that summer caused *Merle* [1980] and *Hisard* [1980] to dub this phenomenon the 'Atlantic Nino'. Further observational results by *Servain et al*. [1982] made clear the connection between changes in the trade winds and changes in SST.

The 1984 event occurred during another observational program called SEQUAL/FOCAL (summarized in the 1984 SEQUAL/FOCAL issue of *Geophysical Research Letters*) and just after the massive 1982-3 Nino. The extensive array of subsurface observations showed that the warming of the mixed layer occurred in conjunction with an anomalous deepening of the oceanic thermocline in the eastern basin [*Philander*, 1986], which resulted from an eastward shift of anomalous heat within the equatorial waveguide [*Carton and Huang*, 1994].

Associated with the warming SSTs are changes in the overlying atmosphere. The equatorial trade winds relax west of 20^{o}W while further eastward the meridional winds associated with the North African summer monsoon also weaken [*Horel et al*., 1986; *Zebiak*, 1993]. Figure 3 shows the anomaly pattern during Atlantic Ninos based on a recent joint ocean-atmospheric analysis [*Ruiz-Barradas et al*., 2000]. Corresponding increases in diabatic heating in the mid-troposphere occur along with a southward shift of tropical convection [*Wagner and da Silva*, 1994; *Carton et al*., 1996]. In particular, rainfall tends to increase on the Guinea coast during an Atlantic Nino [*Hirst and Hastenrath*, 1983]. These anomalous shifts in tropical convection and equatorial winds are well captured by a number of atmospheric GCMs that are forced by Atlantic Nino SST anomalies [*Chang et al*., 2000; *Sutton et al*., 2000; *Okumura and Xie*, 2004], confirming that they result from air-sea interaction much like their El Nino counterparts in the Pacific [*Zebiak* 1993]. This success is not shared by all the GCMs, however, which led *Vizy and Cook* (2002) to explore the use of a regional atmospheric model to simulate the atmospheric response to Atlantic Nino SST anomalies.

The periodicity of the Atlantic Nino seems to vary considerably. The decade beginning in 1974 had few warm events relative to the surrounding decades of the 1960s and 1980s. The reasons for these changes are still poorly understood. Key parameters such as the heat content of the tropical thermocline have only recently come to be measured regularly, while theoretical attention seems to be focusing on changes in the rates of subduction within the tropical thermocline.

As in the case of the eastern Pacific, eastward surges of warm water have important consequences along the southeastern boundary. During 'normal' austral winters the intensification of the North African monsoon as well as the tilting of the equatorial thermocline induces upwelling of cool nutrient-rich water along the coast. The result is a highly productive commercial fishery [*Crawford et al*., 1990; *Boyd et al., 1992*]. Relaxation of the equatorial trade winds and the meridional winds of the North African monsoon causes a southward surge of warm, saline tropical water at least as far south as Namibia raising sea level at Walvis Bay (23°S) by an observable 5 cm [*Brundrit*, 1995]. During the years of the Benguela Nino, the length of the upwelling season may be reduced by a factor of two [*Hagen et al*., 2001].

4. OFF-EQUATORIAL VARIABILITY

4.1. Empirical Studies

Early interest in the relationship between tropical rainfall and SST anomalies was motivated by observational studies of rainfall fluctuations in Northeast Brazil [*Markham and McLain*, 1977; *Hastenrath and Heller*, 1977; *Moura and Shukla*, 1981]. The Nordeste, a part of Brazil dependent on agriculture, has a strongly seasonal cycle in which much of the annual rainfall occurs in the months of March through May when the ITCZ is at its southernmost position. The great drought of 1958 forced 10 million people to emigrate from the Nordeste [*Namias*, 1972]. By matching wet and dry years in the Nordeste with patterns of SST, these studies found that drought associated with an anomalous northward shift of the ITCZ occurred in conjunction with an anomalous northward gradient of SST, an association often referred to as the Atlantic dipole. A second intense example of this circumstance occurred in 1993 [*Rao et al*., 1995] (this dipole pattern maximizes cross-equatorial gradient but does not necessarily imply strong correlations between its centers of action). Somewhat weaker relationships have also been identified between the northward gradient of SST and rainfall anomalies in West Africa [*Folland et al*., 1986; *Lough*, 1986; *Hastenrath*, 1990; *Lamb and Pepper*, 1992]. Nordeste rainfall is also influenced by Pacific El Nino, as in 1958. Section 6 discusses such El Nino effects in more detail.

Many observational studies that followed can be roughly divided into those limited to examining oceanic variables and those looking for covariability between the atmosphere and ocean. Early principal component analyses of SST variability that followed [*Weare*, 1977; *Servain*, 1991] seemed to confirm the presence of a pattern of variability in SST that was geographically stationary, with decadal time-scales. However, it was pointed out by *Houghton and Tourre* [1992] and confirmed by *Mehta* [1998] that when the assumption of spatial orthogonality of the principal components is dropped the northern and southern hemispheres appear to act independently. However in observational studies in which both atmospheric and oceanic variables were included such as *Nobre and Shukla* [1996], *Chang et al*. [1997] and *Ruiz-Barradas et al*. [2000], the results again indicated the presence of a stable pattern of variability across the equator.

The pattern identified by *Ruiz-Barradas et al*. [2000] and presented in Fig. 4 appears in five variables, anomalous wind stress components, diabatic heating, SST and thermocline heat content. The pattern is most pronounced in spring when it is the primary principal component. The SST pattern is most pronounced in the Northern Hemisphere and is accompanied by meridional wind anomalies along the equator heading down the pressure gradient and thus into the warmer hemisphere. Away from the equator the pattern of anomalous wind stress corresponds to an increase in surface winds in the cool hemisphere and a decrease in the warm hemisphere. A dipole pattern of diabatic heating is in its positive phase, reflecting enhanced convection, in the warm hemisphere, also associated with anomalous deepening of the mixed layer. Heat content anomalies seem to follow the thermocline across the equator leading to hemispheric symmetry [see *Ruiz-Barradas et al*., 2000, Fig. 7]. The nodal line of SST anomalies is displaced north of the equator, roughly coinciding with the mean ITCZ. There are considerable anomalies of wind and thermocline depth on the equator associated with this meridional mode. *Servain et al*. (1999) found significant correlation between the meridional and equatorial modes in certain frequency bands.

4.2. Ocean Response

While observational studies disagree on how to characterize TAV in terms of empirical modes, they agree on the following points: i) the meridional position of the Atlantic ITCZ is sensitive to the anomalous cross-equatorial SST gradient (CESG), especially in February-April when the ITCZ is at its southernmost position and the climatological CESG is weak; ii) a meridional dipole configuration of SST anomalies, although it rarely occurs, maximizes the anomalous CESG; iii) off-equatorial SST anomalies are associated with changes in the strength of the easterly trades on either side of the equator/ITCZ.

The third point is addressed by *Carton et al*. [1996] who present several experiments to examine the relative importance of the mechanical effects of wind stress, surface heating, and internal dynamics in controlling the model CESG. They find that wind-induced changes in surface turbulence heat flux are the dominant mechanism for off-equatorial SST variability. When the effect of wind variability on surface latent heat flux is artificially suppressed, the model CESG variability is substantially reduced (Fig. 5). By contrast, when *Carton et al*. remove interannual variability in wind stress but not in wind effect on latent heat flux, the model reproduces CESG variability despite a marked reduction in variability in ocean dynamic fields such as the thermocline depth. These results show that surface heat flux is the leading order process there, in contrast to the equatorial region where ocean dynamics are important [e.g. *Carton and Huang* 1994]. Subsequent calculations using ocean mixed layer models [*Xie and Tanimoto*, 1998; *Czaja et al*., 2002; *Kushnir et al*., 2002a] and a different ocean GCM [*Seager et al*. 2001] confirm the major role of wind-induced evaporation in off-equatorial SST variability, a result consistent with the ocean mixed layer heat budget analysis based on observations [*Wagner*, 1996].

The situation in the South Atlantic is less clear. *Hakkinen and Mo* [2002] suggest that ocean circulation changes may be important for southern tropical Atlantic SST variability. But lack of observations may also be responsible for the apparent weakening of the trade wind-SST relation there. *Tanimoto and Xie* [2002] point out that south of 10^{o}S, anomalies of surface wind velocity and sea level pressure based on historical ship observations are often not even in geostrophic balance indicating that the data coverage may be insufficient to draw meaningful conclusions there. *Wu et al*. [2004] in this volume offer further evidence for the effect of anomalous vertical heat advection on southern tropical Atlantic SST variability.

All the above model studies are based on coarse-resolution simulations that do not resolve mesoscale ocean eddies. *Jochum et al*. [2004, this volume] suggest that these eddies due to hydrodynamic instabilities of equatorial currents be a significant source of interannual variability in SST, especially in boreal spring when the ITCZ is sensitive to SST anomalies.

4.3. Air-sea Feedback

Chang et al. [1997] combined the wind-induced evaporation mechanism (point iii) with the direct CESG-atmospheric pressure gradient mechanism of driving cross-equatorial winds (point i) to provide an air-sea interaction scenario for Atlantic CESG variability. They hypothesize that the trade wind anomalies such as those in Fig. 4 are forced by SST anomalies with a strong CESG, an assumption supported to various degrees by *Moura and Shukla* [1981] and subsequent atmospheric GCM studies (the atmospheric response to off-equatorial SST anomalies is a complex issue by its own and will be discussed in detail in Subsection 4.5). A positive anomalous CESG sets up an anomalous southward pressure gradient in the atmospheric boundary layer [e.g., *Lindzen and Nigam*, 1987], inducing southerly cross-equatorial winds that decelerate the easterly trades north of the equator because of the Coriolis effect. *Chang et al*. [1997] suggest that these weakened trades north of the equator reduce surface evaporation, thereby acting to strengthen the initial CESG. South of the equator, the southeasterly trades accelerate, increasing surface evaporative cooling and the northward CESG. This positive thermodynamic wind-evaporation-SST (WES) feedback was originally proposed to explain the northward displacement of the climatological ITCZ over the eastern Pacific and Atlantic [*Xie and Philander*, 1994].

Besides WES, there seem to be additional feedback mechanisms that act between SST and clouds. In a composite analysis of historical ship observations based on a CESG index, a quadrupole banded structure emerges from the cloudiness field [*Tanimoto and Xie*, 2002]. Near the equator, a dipole of cloudiness anomaly appears in association with the shift of convective clouds in the ITCZ, acting as a negative feedback onto SST as more clouds form over the warmer side of the SST dipole. In the subtropics, more low-level clouds form over negative SST anomalies, reducing net radiation into the oceanic mixed layer causing more cooling and more clouds, etc. South of the equator, the positive feedback resulting from the negative SST-low cloud correlation results in roughly a 10% increase in cloud cover or a 20 Wm^{-2} reduction in incoming solar radiation at the surface for each 1^{o}C increase in SST. This SST-low cloud feedback mechanism is significantly weaker in the northern tropical Atlantic.

While the mechanisms underlying air-sea interaction have emerged only recently, their potential for maintaining CESG anomalies was recognized much earlier. For example, *Hastenrath and Greischar* [1993] state "the SST pattern—itself affected by the surface wind field—…is conducive to a steeper meridional pressure gradient, which in turn favors a stronger southerly wind component." Our expectation is that improved understanding of the physical mechanisms underlying air-sea interaction will lead to improved physically based numerical models, which may then have benefit for prediction systems.

4.4. Free Mode Analysis

Linear stability analysis of the coupled equatorial ocean-atmosphere system has yielded useful insights into the dynamics of ENSO [*Neelin et al*., 1998]. By including the effect of wind-induced evaporation on SST, *Zhou and Carton* [1998] and *Xie et al*. [1999] extend such stability analyses with a more sophisticated surface heat flux formulation including latent heat loss. Two types of coupled modes emerge from the latter extended analysis: one arising from the *Bjerknes* [1969] feedback involving interaction of the thermocline depth, upwelling, SST and zonal winds along the equator; and one due to the thermodynamic WES feedback involving air-sea interaction in the meridional direction. The zonal and meridional modes differ not only in spatial structure—the former with maximum amplitude at the equator while the latter off the equator—but also in the growth rate dependence on zonal wavenumber. The zonal mode favors a zonal wavelength about the size of the Pacific, the meridional mode grows fastest at zonal wavenumber zero, a property consistent with the fact that off-equatorial anomalies of Atlantic SST and wind are nearly zonally uniform in phase. In *Xie et al*.'s [1999] calculations, the growth rate of the equatorial mode at the size of the Atlantic basin is comparable to that of the zonally uniform meridional mode, being 0.6 and 0.8 year^{-1}, respectively, in the absence of SST damping. (The equatorial mode's growth rate is 1.5 year^{-1} at the Pacific basin size.) The wind-induced evaporation anomalies are about 10-20 Wm^{-2} in amplitude in the deep tropics and thus the WES feedback is only weakly positive. In fact, it may even be negative when the dependence of SST on surface evaporation is considered.

Chang et al. [1997] were the first to demonstrate the role of WES in CESG variability in a coupled model. Their atmospheric model is empirically constructed based on a singular value decomposition analysis of surface momentum/heat fluxes and SST over the tropical Atlantic. The coupling of this atmospheric model with either an intermediate ocean model of *Zebiak and Cane* [1987] or an ocean GCM yields oscillations on decadal timescales, in which SST anomalies are organized to maximize CESG with opposite polarities on either side of the equator **(**Fig. 6**)**. Chang et al. show that the growth rate of this meridional mode is highly sensitive to the coupling of surface heat flux with SST but not so to the coupling with momentum flux, a result consistent with *Carton et al*.'s [1996] ocean GCM experiments. In the Chang et al. model the CESG feedback is limited by cross-equatorial advection by the northward flowing North Brazil Current, which helps switch phase of the coupled oscillation, thus setting the timescale of reversal.

The SST advection by surface Ekman flow acts to dampen the growth caused by the positive WES feedback, an effect evident in *Xie*'s [1999] energy equation analysis and in a recent coupled model study of *Kushnir et al.* [2002a]. *Seager et al*.'s [2001] ocean GCM calculations also show that the advection by the mean ocean currents acts as a damping on CESG variability.

One important feature of the observed SST "dipole" pattern is the nodal line's displacement north of the equator along the mean ITCZ. *Okajima et al*. [2003] explore the importance of this asymmetry with an atmospheric GCM coupled to a Zebiak-Cane-type ocean model modified to allow the thermocline depth to vary in space but not in time. This modification suppresses the thermocline feedback and virtually eliminates the ENSO mode in the system, allowing a close look into air-sea interaction in the meridional direction. In a run of this hybrid coupled GCM with a perfectly symmetric land-sea distribution, SST variability in the tropics organizes itself into a distinct dipole pattern with its nodal line on the equator (Fig. 7a). The corresponding SST-wind speed relationship is consistent with positive WES feedback. When Okajima et al. perturb the shape of continents to force the mean position of the ITCZ into the Northern Hemisphere, the line of minimum SST variance shifts northward as well (Fig. 7b). Furthermore, SST variability becomes much less coherent across the mean ITCZ/SST nodal line with the SST correlation decreasing from 0.7 when the mean ITCZ is symmetric about the equator to 0.2 when the mean ITCZ is displaced a realistic distance off the equator.

The WES feedback owes its positive sign to the sign change in the Coriolis parameter across the equator. *Okajima et al*. [2003] suggest that the departure of the climatic equator from the geographical equator weakens the WES feedback and reduces the coherence between variability north and south of the ITCZ. This impact of the shift of the mean ITCZ may explain the lack of significant interhemispheric coherence in TAV [*Houghton and Tourre, 1992*; *Enfield et al*., 1999; *Mehta*, 1998] and why CESG variability and associated atmospheric anomalies are strongest in March-April, the time when the climatological ITCZ is nearly symmetric about the equator.

4.5. Atmospheric Response

In the extratropics, a negative correlation between anomalies of SST and wind speed is now recognized as evidence for atmospheric forcing of the ocean mixed layer rather than the other way around [*Frankignoul*, 1985; *Kushnir et al*., 2002b]. In the tropical Atlantic, however, the shift in dynamics toward a direct atmospheric response to SST suggests that the observed negative SST-wind speed correlation may support the hypothesis of a positive WES feedback that organizes SST and wind anomalies into a dipole pattern that maximizes CESG, as discussed in the previous two subsections.

To prove this hypothesis of the existence of a coupled meridional mode, it still needs to be shown that the atmosphere responds to CESG anomalies (rather than causes them) and that the sign of the response is such that it leads to positive WES feedback. The basic physics of the atmospheric response was originally examined by *Moura and Shukla* [1981] who explored the response of an atmospheric GCM to an imposed CESG. They report a decrease in sea level pressure over the hemisphere with positive SST and an anomalous shift of the ITCZ in the direction of the imposed CESG. Recent studies conduct an ensemble of multi-decade hindcasts forced by observed SST and use a signal-to-noise maximizing EOF technique to extract SST-forced signals [*Venzke et al*., 1999; *Chang et al*., 2000; *Sutton et al*., 2000; *Terray and Cassou*, 2002]. Some other studies impose time-invariant SST anomalies and integrate models for a long period of time to increase the sample size and thereby reduce the noise due to atmospheric internal variability [*Dommenget and Latif*, 2000; *Okumura et al*., 2001; *Sutton et al*., 2001; *Terray and Cassou*, 2002]. The difficulty of the current atmospheric GCMs in reproducing the WES feedback is discussed in *Wang and Carton* [2003].

Nearly all the GCMs agree in their response in the deep tropics within 10o latitude [*Chang et al*., 2000; *Sutton et al*., 2000; *Okumura et al*., 2001; *Terray and Cassou*, 2002]. In response to an SST dipole and the associated changes in CESG, these models generate cross-equatorial winds directed from the colder to the warmer hemisphere. The resultant low-level convergence causes the Atlantic ITCZ to move toward the warmer side of the SST dipole. The change in precipitation is not limited to the ocean but extends considerably inland over South America, probably as a result of the westward propagation of baroclinic Rossby waves forced over the ocean. The presence of substantial diabatic heating over the Amazon basin indicates a possible role for interactions with land surface processes.

In the deep tropics, the atmospheric response is baroclinic with wind anomalies being out of phase in the lower and upper troposphere. The easterly trade winds tend to weaken on the equatorward side of positive SST anomalies and strengthen on the equatorward side of negative SST anomalies. This trade wind response supports the WES feedback as envisioned by *Chang et al*. [1997]. While *Chang et al*. [2000] find the significant response confined to 10^{o}S-10^{o}N in their model, other models show a broader response in latitude. Based on a single multi-decadal integration, results of *Robertson et al*. [2000] and *Watanabe and Kimoto* [1999] hint at a response of the NAO to tropical Atlantic SST anomalies. Model studies with large ensemble members/long integrations seem to support this extratropical response [*Okumura et al*., 2001; *Sutton et al*., 2001; *Terray and Cassou*, 2002]. In response to a positive CESG, a barotropic low develops in the mid-latitude North Atlantic centered around 45^{o}N, in addition to a baroclinic response in the deep tropics [*Okumura et al*., 2001]. This barotropic response is strongest in boreal winter and spring [*Venzke et al*., 1999; *Okumura et al*., 2001; *Sutton et al*., 2001; *Terray and Cassou*, 2002] and allows the relaxed trades and hence the positive WES feedback to cover the entire northern tropical Atlantic.

Intermediate baroclinic models of the atmosphere are very useful in studying ENSO over the tropical Pacific but they are much less successful in reproducing the surface wind response to tropical Atlantic SST variations, especially in the subtropics [*Chiang et al*., 2001; *Chung et al*., 2002]. One likely explanation is because the barotropic response in the subtropical/midlatitude Atlantic modulates the trade winds. The exact mechanism for this barotropic response needs further study. One example is barotropic Rossby wave excitation by upper-tropospheric convergence/divergence associated with the anomalous shift of the ITCZ. Much as in the extratropical response to ENSO, the North Atlantic storm track varies in such a way as to reinforce barotropic stationary eddies [*Watanabe and Kimoto*, 1999; *Okumura et al*., 2001]. Another important factor may be diabatic heating over the surrounding continents that alters the zonal pressure gradient and preferentially affects the zonal component of the near-equatorial winds.

A related line of research is the use of atmospheric GCMs to investigate the relationship between the NAO and a pattern of anomalous SST in the form of a tripole (the tropical extension of which represents the CESG), which are the dominant modes of the atmosphere and ocean over the North Atlantic, respectively. Several studies show that the observed NAO time series can be reproduced, albeit at reduced amplitudes, in atmospheric GCMs forced by observed SST [*Venzke et al*., 1999; *Rodwell et al*., 1999; *Mehta et al*., 2000; *Peng et al*., 2002; *Lin and Derome*, 2003], a result that *Bretherton and Battisti* [2000] suggest is consistent with the null hypothesis of atmospheric stochastic forcing of the ocean. It is also likely, as GCM experiments of *Sutton et al*. [2001] and *Terray and Cassou* [2002] show, that the tropical part of the SST tripole is what forces the NAO-like response in the extratropics, perhaps by shifting the ITCZ. Indeed, *Watanabe and Kimoto* [1999] and *Okumura et al*. [2001] show that the extratropical part of the SST tripole can be reproduced in an ocean mixed layer coupled with an atmospheric GCM that is forced by a tropical SST dipole. To the extent that this tropical forcing scenario holds and that the tropical Atlantic anomalies result from local air-sea interaction, a certain degree of predictability may be achieved for the NAO-tripole pair. In ensemble atmospheric GCM experiments, percentage of SLP variance due to imposed SST variability generally decreases poleward, being 30-60% in the subtropics and 10-30% in the mid-latitudes [*Kushnir et al*., 2002b; *Rodwell*, 2003].

Figure 8 shows the regressions of observed SST, surface wind velocity and net surface heat flux upon a northern tropical Atlantic SST index for March-May [see also *Kushnir et al*., 2002a]. Large anomalies of trade winds take place during January-March, preceding the large SST anomalies during March-May. *Czaja et al*. [2002] attribute these trade wind anomalies in January-March exclusively to external forcing but the atmospheric GCM studies mentioned above suggest that these wind anomalies may partly result from the mid-latitude barotropic response to a northward shift of the ITCZ. A quantitative estimate of the importance of tropical SST forcing is difficult to make from observations. *Czaja et al*. [2002] note that wind anomalies in the deep tropics are likely a response to the northern tropical Atlantic SST anomalies, which appear beginning in January and persist into boreal summer. Despite an SST regression pattern confined north of the equator, the wind regression extends well into the Southern Hemisphere and is consistent with the phase relationship expected from WES feedback. Such a cross-equatorial response is seen in atmospheric GCM results when SST anomalies are imposed only on one side of the equator [*Sutton et al*., 2001].

5. INTERACTION WITH THE EXTRATROPICS

In the tropics SST variations are important in shaping the spatial pattern of variations in convection and hence other atmospheric fields. In the extratropics, on the other hand, the atmospheric dynamics organize low-frequency variability into large-scale patterns, even without feedback from the ocean. The NAO is such a preferred pattern that dominates the month-to-month atmospheric variability over the North Atlantic [*Hurrell et al*., 2003]. Atmospheric GCM simulations with climatological SST as the surface boundary condition confirm that the NAO is a dominant mode of atmospheric internal variability but that in the absence of SST variability its spectrum is likely white in time. The observed NAO shows enhanced power at decadal timescales, which may result from air-sea interaction within the extratropical North Atlantic [*Marshall et al*., 2001] or from teleconnections excited by anomalous shifts in the Atlantic ITCZ [*Okumura et al*., 2001; *Sutton et al*., 2001; *Terry and Cassou*, 2002], or both.

5.1. A pan-Atlantic Pattern

The NAO is correlated with the SST tripole over the North Atlantic in boreal winter/spring, with the tropical/subtropical lobe of the latter centered at 10-20^{o}N. *Rajagopalan et al*. [1998] find that the NAO is also correlated with both southern tropical Atlantic SST and CESG variability at decadal timescales. In fact, *Xie and Tanimoto*'s [1998] composite analysis based on a CESG index reveals a pan-Atlantic pattern, with bands of SST anomalies of alternating signs that span from the South Atlantic to Greenland (Fig. 9). This so-called pan-Atlantic decadal oscillation pattern features anomalous wind/sea level pressure (SLP) patterns similar to the NAO and a SST tripole over the North Atlantic [*Tanimoto and Xie*, 1999], and is captured in a joint analysis of SST and SLP over the whole Atlantic basin using a frequency domain method [*Tourre et al*., 1999].

There is some evidence that the above pan-Atlantic pattern favors decadal timescales. *Enfield et al*. [1999] show that SST variability in the northern and southern tropical Atlantic displays marginally significant coherence with anti-symmetric phase in the 8-12 year band in the boreal winter-spring. *Chu* (1984) reports a spectral peak in the frequency band of 12.7-14.9 years in northeast Brazil rainfall. *Mehta* [1998] notes a similar decadal peak in this regional rainfall variability as well as in Atlantic CESG. Such a decadal (12-13 years) peak is found in an 825-year long sediment core in the Cariaco Basin north of the Venezuelan coast, which *Black et al*. [1999] interpret as resulting from variability in the northern tropical Atlantic trades. *Tourre et al*.'s [1999] joint SST-SLP pattern is associated with a spectral peak centered at a period of 11.4 years. When the empirical orthogonal function analysis is performed separately for SST and SLP and over the separate northern and southern tropical Atlantic domains, the four independently obtained leading principal components are highly correlated at decadal timescales [*Tanimoto and Xie*, 2002]. Instrumental records, however, are too short to test further this hypothesis for a pan-Atlantic decadal oscillation.

5.2. Subtropical High as the Bridge

This statistical relationship between the CESG and NAO suggests an interaction between them. In one direction, as atmospheric GCM studies suggest, CESG variability affects the NAO through its effect on the Atlantic ITCZ and upper-tropospheric divergence (see Subsection 4.5). *Xie and Tanimoto* [1998], *Chang et al*. [2000], *Czaja et al*. [2002], and *Kushnir et al*. [2002a] investigate this interaction from the other direction by asking how the NAO affects the CESG? The subtropical/mid-latitude center of action of the NAO in SLP is located at 40^{o}N, driving changes in the strength of the northeasterly trades to the south and hence affecting northern tropical Atlantic SST. Assuming that the direct NAO influence on SST is strongest in the subtropics, *Xie and Tanimoto* [1998] apply an external forcing that is random in time and confined poleward of 20^o latitude and show that the WES feedback organizes the tropical response into a coherent dipole structure that favors low-frequency (interannual) variability [see also *Kushnir et al*., 2002a]. In particular, when the observed winds are used as the subtropical forcing, the coupled model reproduces the observed CESG evolution quite well despite the fact that the tropics are free of external forcing (Fig. 10). Thus, air-sea feedback may act to transfer the impact of subtropical anomalies like those associated with the NAO into the deep tropics.

Chang et al. [2001] derive an atmospheric noise field empirically based on a 145-year atmospheric GCM run that is forced by the monthly SST climatology. The dominant noise pattern resembles the NAO and features wind anomalies with large amplitudes in the subtropics that decay rapidly toward the deep tropics (Fig. 11, left). They force a hybrid coupled ocean GCM with this noise field with a white spectrum in time. In the absence of air-sea feedback, SST anomalies are confined to the subtropics with little response near the equator. When moderate coupling is allowed in the model, regions of maximum SST variability shift toward the equator to 10-20o in latitude, accompanied by large CESG variations reddened at interannual and lower frequencies. Associated with the change in SST spatial structure, the trade winds on both sides of the equator show large variations in the deep tropics, with stronger cross-equatorial coherence than in the uncoupled run. This result supports the notion that NAO's influence on CESG is rather indirect and requires the bridging effect of air-sea interaction to reach the deep tropics.

Namias [1972] reports a correlation between the North Atlantic subtropical high and northeast Brazil rainfall in boreal winter and spring. He notes "the northeast trades are often regulated by the Atlantic subtropical anticyclone" and "these variations alter the intensity of convergence into the ITCZ and perhaps change its position". *Czaja et al*. [2002] construct a simple model for northern tropical Atlantic SST and show that subtropical SLP variations—the forcing in the model—explain most of the observed SST variability. This seems to suggest that air-sea interaction within the tropical Atlantic is secondary, but as *Xie and Tanimoto* [1998] and *Chang et al*. [2001] show, the influence radius of subtropical forcing may be a function of local air-sea feedback; without the feedback, its effect may well be confined and not extend into the deep tropics to affect CESG and ITCZ.

The NAO and the SST tripole emerge as the leading mode from joint ocean-atmospheric analyses of coupled GCM simulations [*Grötzner et al*., 1998; *Delworth and Mehta*, 1998]. However, different studies disagree on how far the NAO influence can penetrate toward the south. *Delworth and Mehta* [1998] report that it is limited to north of the equator but a pan-Atlantic pattern emerges from *Watanabe et al*.'s [1999] simulation, with an SST dipole in the tropics.

The interaction of the TAV with the extratropics is much less well studied in the Southern than in the Northern Hemisphere. Based on a singular value decomposition analysis, *Venegas et al*. [1997] show that there is a meridional SST dipole pattern in the South Atlantic and that it is associated with variations in subtropical SLP, a co-variation pattern similar to that over the North Atlantic. They note that the SST-SLP pattern is most pronounced in the southern summer, a result that they suggest is indicative of "possible links with major climatic oscillations observed in the Northern Hemisphere". CESG/ITCZ variability and air-sea interaction in the tropical Atlantic may well be the mechanism for such interhemispheric links. In fact, the South Atlantic SST dipole of *Venegas et al*. [1997] is part of the pan-Atlantic pattern (Fig. 9) as documented by *Tanimoto and Xie* [1999; 2002]. *Barreiro et al*. [2004, this volume] investigate further the influence of South Atlantic extratropical variability on TAV.

5.3. Oceanic Pathways

So far, we have examined the link between TAV and the extratropics via the atmosphere. There are also oceanic pathways that link the subtropical with the equatorial Atlantic, via so-called subtropical cells [*Schott et al*., 2004, this volume]. These pathways carry water subducted in the subtropics during winter into the equatorial upwelling zones. In the Atlantic, these subtropical cells are highly asymmetric about the equator because of the deep meridional overturning circulation [MOC; *Jochum and Malanotte-Rizzoli*, 2001] and are sensitive to changes in wind stress [*Inui et al*., 2002].

Changes in the deepwater formation in the high-latitude North Atlantic can induce changes in cross-equatorial ocean heat transport, which *Yang* [1999] suggests give rise to a dipole SST pattern in the tropical Atlantic. Using a coupled GCM, *Dong and Sutton* [2002] show that this MOC-induced SST dipole amplifies in the tropical Atlantic by interacting with the atmosphere. In particular, in response to a sudden weakening of the MOC, a SST dipole develops in year 4-6, with a strong cooling over the northern and a weak warming over the southern tropical Atlantic. SLP increases over the region of sea surface cooling and decreases over warming. The resulting anomalous CESG causes the Atlantic ITCZ to shift southward, triggering further changes over the tropical Pacific in their model. *Dong and Sutton* [2002] suggest that the effect of changes in the high-latitude North Atlantic and the MOC can be felt quickly through the globe via such atmospheric feedback in the tropical Atlantic, and that this process occurs in years instead of the hundreds of years one would expect if only ocean processes were involved.

Analysis of paleo-proxies shows a strong correlation between the position of the Atlantic ITCZ and Greenland climate conditions [*Peterson et al*., 2000]. *Chiang et al*. [2003] suggest that the pan-Atlantic pattern discussed earlier in this section is a useful model, with the interaction and feedback between CESG and the position and strength of the ITCZ as a possible mechanism for this link between the high-latitude and tropical North Atlantic. Using an atmospheric GCM coupled with a slab ocean mixed layer, Chiang et al. show that continental ice sheets present during the last glacier maximum could trigger tropical air-sea interaction, by altering atmospheric stationary wave patterns, giving rise to pan Atlantic scale changes with a large anomalous CESG that is coupled with the oceanic ITCZ. Chiang et al. conclude that their model prefers the meridional mode in the tropics in response to various surface forcing terms during the last glacier maximum. Since their ocean model is one-dimensional and does not include any dynamics, they suggest that the atmospheric response to high-latitude changes in sea and land ice is an alternative means of triggering changes in the tropical air-sea system, besides the MOC mechanism of *Yang* [1999] and *Dong and Sutton* [2002]. In both the studies of *Dong and Sutton* [2002] and *Chiang et al*. [2003] the WES feedback seems to be a key to communicating the high-latitude changes to the deep tropics, leading to changes in CESG and ITCZ.

6. ENSO INFLUENCE

It has been known for some time that a basin-wide warming takes place in the tropical Atlantic a few months after the El Nino in the Pacific peaks in December-January. During and immediately following an El Nino event, precipitation generally decreases over the equatorial Atlantic. This section reviews studies of ENSO influence in the tropical Atlantic.

The Atlantic response to ENSO shows strong seasonality because both ENSO and its influence on the Pacific North American (PNA) teleconnection are seasonally phase-locked. The Atlantic response to La Nina is similar in spatial pattern to that to El Nino, albeit with anomalies reversing signs. For this reason, the following discussion describes the response to El Nino.

6.1. SST

In December and January when El Nino peaks in the Pacific, SLP in the subtropical and mid-latitude North Atlantic drops while increasing in the equatorial Atlantic [*Covey and Hastenrath*, 1978; *Aceituno*, 1988; *Giannini et al*., 2000; *Mestas-Nunes and Enfield*, 2001; *Alexander and Scott*, 2002]. The resultant anomalous pressure gradient drives anomalous southwesterlies over the tropical North Atlantic north of 10^{o}N. These anomalous winds are particularly strong in the western half of the basin, acting to weaken the prevailing northeast trades on the background and hence surface latent and sensible heat flux [*Aceituno*, 1988; *Curtis and Hastenrath*, 1995; *Lanzante*, 1996; *Enfield and Mayer*, 1997; *Klein et al*., 1999]. This reduced heat release from the ocean gives rise to a delayed warming of the ocean mixed layer that peaks in April-June in a zonal band between 20^{o}N and the latitude of the climatological ITCZ (Fig. 12). The decrease in surface evaporation prior to this tropical North Atlantic warming is captured in *Klein et al*.'s [1999] calculations based on ship observations. In addition, *Klein et al*. [1999] report a modest reduction in cloud cover south of 20^{o}N that further contributes to the ocean warming.

Based on atmospheric GCM simulations, *Saravanan and Chang* [2000] suggest that in addition to wind-induced evaporation variations, changes in air-sea difference in surface temperature and humidity are also important for the sea surface warming in the tropical North Atlantic in boreal spring following an El Nino event. *Chikamoto* [2002] confirm this air-sea temperature/humidity difference effect by performing a heat flux analysis based on historical ship reports. They show that much of the decrease in turbulent heat flux over the tropical northwestern Atlantic is due to an increase in air-sea temperature difference. Normally, SST anomalies are slightly higher than surface air temperature anomalies, but in this region and in the boreal winter-spring following an El Nino, anomalies of air temperature are larger and lead those of SST (Fig. 13), thereby suppressing surface heat release from the ocean.

ENSO-induced tropical North Atlantic warming induces further air-sea interaction within the tropical Atlantic. In April-June when this warming is at its maximum, significant southeasterly wind anomalies form in a region between the latitude of the mean ITCZ and 10-15^{o}S, apparently in response to the decrease in SLP over the band of positive SST anomalies to the north (Fig. 12). These anomalous southeasterlies are in the general direction of the climatological background winds and induce negative SST anomalies south of the equator, through the dependence of evaporation on wind speed. This cooling increases the northward SST gradient and hence the anomalous southeasterly cross-equatorial winds, implying a positive WES feedback discussed earlier. *Enfield and Mayer* [1997] discuss this tendency for the tropical Atlantic to develop a cross-equatorial SST gradient in the boreal summer following an El Nino [see also *Chiang et al*., 2002]. The SST correlation with ENSO is generally weaker in the South than in the North Atlantic. The abovementioned cooling south of the equator shows a correlation just above 0.25 (vs. well above 0.5 in the tropical North Atlantic). A positive correlation of around 0.5 develops in the subtropical South Atlantic southwest of the above cooling region [*Enfield and Mayer*, 1997; *Klein et al*., 1999].

Thus, local air-sea interaction is apparently involved in the Atlantic response to Pacific variability, but the feedback is not strong enough to sustain the cross-equatorial SST gradient anomalies through the boreal summer. In contrast to this ENSO influence on cross-equatorial SST variability, the correlation between the Atlantic Nino and ENSO is generally insignificant [*Zebiak*, 1993] despite the presence of significant southeasterly wind anomalies near the equator (Fig. 12; see also *Latif and Barnett* [1995]) that by themselves tend to induce a cold event in the equatorial Atlantic by increasing equatorial upwelling and shoaling the thermocline in the east. (The 1984 Atlantic Nino is one exception taking place following the major El Nino in 1982-83.) Between the anomalous southwesterlies north of 10^{o}N and southeasterlies near the equator, there is a significant band of negative wind curl during January-March that excites downwelling Rossby waves. The opposing effects of these Rossby waves and anomalous equatorial easterlies may be responsible for the lack of correlation between equatorial Pacific and Atlantic SST.

6.2. Precipitation

During an El Nino event, atmospheric convection intensifies in the central and eastern equatorial Pacific. The increased convective heating warms the Pacific troposphere. These positive tropospheric temperature anomalies created in the Pacific are rapidly spread along the equatorial belt via equatorial wave adjustment and occupy the global tropics [*Yulaeva and Wallace*, 1994; *Chiang and Sobel*, 2002; *Su et al*., 2004, this volume]. Outside the tropical Pacific, this tropical tropospheric warming is associated by the anomalous descending motion as part of the anomalous Walker circulation associated with ENSO. This anomalous subsidence, along with the increased static stability associated with the tropospheric warming over the global tropical belt, suppresses atmospheric convection and reduces precipitation over the equatorial Atlantic. During March-May as the tropical North Atlantic warms up, the Atlantic ITCZ shows a tendency to shift anomalously northward, with a dipole in the precipitation anomaly field. This precipitation dipole is not limited to the oceanic sector but extends into the South American continent as well, with a large decrease in rainfall over the Brazil's Nordeste region and a modest increase over the continent north of the equator. By comparing two atmospheric hindcasts with SST forcing prescribed over the global tropics and the tropical Atlantic, respectively, *Saravanan and Chang* [2000] show that the rainfall reduction over the equatorial Atlantic is the direct response to ENSO (via anomalous downdraft and tropospheric warming) while the northward shift of the ITCZ is an indirect response forced by Atlantic SST anomalies (notably the tropical North Atlantic warming and the attendant cross-equatorial SST gradient). *Chiang et al*. [2002] confirm this sequence of rainfall response based on observational analysis.

March-May rainfall over the Caribbean Sea increases following El Nino, while rainfall response over the land surrounding the Caribbean is also affected by orography [*Giannini et al*., 2000; *Taylor et al*., 2002].

6.3. Teleconnection Mechanism

Besides the adjustment through an anomalous Walker circulation, the PNA teleconnection is an additional mechanism by which ENSO affects the North Atlantic in boreal winter and early spring [*Nobre and Shukla*, 1996; *Klein et al*., 1999; *Lau and Nath*, 2001]. In particular, the center of action over Florida associated with the barotropic PNA pattern contributes to the lowering of SLP there, and to the weakening of the northeasterly trades and warming of the ocean mixed layer in the tropical North Atlantic. Based on an ensemble hindcast with an atmospheric GCM that is forced by tropical Pacific SST variations and coupled with a slab ocean mixed layer model, *Lau and Nath* [2001] show that the influence of ENSO is not limited to the tropical Atlantic but is also significant in the extratropical North Atlantic, a result that further supports the idea that the PNA mechanism plays an important role in the Atlantic response to ENSO [*Alexander et al*., 2002].

Whereas to first order ENSO is symmetric about the equator in the Pacific, the response of the Atlantic is quite equatorially asymmetric, with the strongest anomalies of SST and rainfall observed in the northern tropics. The PNA mechanism accounts partly for this asymmetry in the Atlantic response. Recently, *Chiang and Sobel* [2002] suggest that the tropical tropospheric warming associated with ENSO—an effect of the anomalous Walker circulation mechanism—is effectively communicated to the ocean surface through vertical mixing by moist convection. Thus, the resultant ocean mixed layer warming is confined to the convective regions such as the northward-displaced oceanic ITCZ (see Fig. 13).

6.4. Coupled Model Studies

Using a coupled GCM, *Huang et al*. [2002] carry out an experiment in which only the tropical Atlantic between 30^{o}S-30^{o}N is coupled with the atmosphere and the observed SST history is prescribed elsewhere. Huang et al. report that the prescribed ENSO exerts a strong influence on the tropical North Atlantic and explains up to 50% of the variance in their model. Consistent with *Enfield and Mayer* [1997], the model tropical North Atlantic warming subsequently induces southerly cross-equatorial winds, which interact further with the ocean, leading to SST anomalies south of the equator. In *Huang et al*.'s [2002] model ENSO's influence is weak on and south of the equator, where most of SST variability is due to air-sea interaction local to the tropical Atlantic.

In an independent study with a different coupled GCM, *Wu and Liu* [2002] confirm the importance of local air-sea interaction and in particular the WES feedback in TAV. In a so-called partially coupled experiment in which the active feedback onto the atmosphere is removed over the northern tropical Atlantic, SST variability is reduced more than half compared with a control. This leads Wu and Liu to suggest that the tropical North Atlantic is not just passively responding to external forcing such as ENSO and NAO but positive feedback arising from air-sea interaction is necessary to produce the right level of variability there [see also *Wu et al*., 2004, this volume].

7. SUMMARY AND DISCUSSION

The seasonal cycle is by far the largest source of climate variability in the tropical Atlantic. The seasonal cycle consists of a north-south anti-symmetric annual component that is forced directly by the seasonal march of the sun, and a north-south symmetric component with a maximum on or slightly south of the equator. This air-sea interaction component is triggered by the onset of the West African monsoon causing rapid equatorial cooling in May and June, which is further amplified by air-sea interaction along the equator through a mechanism similar to that *Bjerknes* [1969] envisioned for ENSO.

On interannual and longer timescales, no single mode seems to dominate. Instead, several mechanisms are responsible for tropical Atlantic variability. On the equator, both observational and modeling studies indicate that there is a Bjerknes-type air-sea coupled mode arising from the interaction of the equatorial zonal SST gradient, ITCZ convection, zonal wind, and thermocline depth. The resulting positive feedback here in the Atlantic is weaker than the corresponding feedback in the Pacific probably because of the smaller zonal width of the Atlantic basin. Thus, the anomalous warming on the equator, which occurs every few years, is modest in amplitude and lasts only for a few months in boreal summer. This warming is generally associated with an increase in rainfall along the coasts of Guinea and Angola.

In addition to an equatorial mode, observational and modeling studies generally support the notion that interannual variability in the cross-equatorial SST gradient and the position and strength of the oceanic ITCZ are coupled and that this coupling results to some degree from their mutual interaction. The interaction involves positive WES feedback between anomalous trades, wind-induced changes in surface evaporation, and SST anomalies. The coupled ITCZ/CESG variability affects rainfall over the surrounding continents, in particular over northeastern Brazil and to a lesser extent over the Sahel.

In addition to local interactions the tropical Atlantic is also subject to strong external forcing. ENSO warming in the equatorial Pacific reduces the northeasterly trades and gives rise to a delayed warming in the northern tropical Atlantic through the PNA teleconnection and subsidence associated with an anomalous Walker circulation. The NAO also modulates the strength of the northeast trades and hence SST in the subtropical North Atlantic. Such external forcing of the northeast trades explains a large percentage of observed SST variability in the northern tropical Atlantic, which subsequently triggers the ITCZ/CESG interaction in the deep tropics and induces changes on and across the equator. All these TAV mechanisms are highly seasonal: ENSO and NAO forcing is strongest in boreal winter; the ITCZ/CESG interaction in March-May when the equator is uniformly warm; and the equatorial mode is most pronounced in the boreal summer coinciding with the season of the cold tongue and the shallow thermocline in the east.

While the ITCZ/CESG interaction almost certainly exists, many uncertainties remain. It is unclear, for example, how far the ITCZ/CESG interaction extends toward the poles. There is observational evidence for a positive SST-low cloud feedback in the subtropics, indicative of an atmospheric reaction in the planetary boundary layer to SST anomalies. Some modeling studies suggest that the CESG-induced shift of the ITCZ and the resultant shift of upper-tropospheric divergence force a barotropic response that modulates the strength of the North Atlantic subtropical high and the northeast trades. This suggests that ITCZ/CESG in the deep tropics might interact with the subtropical Atlantic, a mechanism that may give rise in turn to the observed pan-Atlantic pattern of anomalies of SST, SLP, and surface wind. Interestingly, this pan-Atlantic pattern has been used to explain a link between the tropical and high-latitude North Atlantic observed in paleoclimate records.

There is a paradox in TAV research. While theoretical studies indicate that the WES feedback favors a dipole mode anti-symmetric about the mean position of the ITCZ, observed SST variability is not significantly correlated across this latitude. This paradox may be reconciled by considering the departure of the position of the climatological ITCZ from the geographical equator, which acts to reduce interhemispheric coherence of WES-induced variability. Other mechanisms for reducing this correlation include interference with other modes of variability like the Atlantic Nino, and disruption by external forcing that is generally not projected optimally onto the meridional mode.

Ocean-atmospheric interaction and feedback, when they exist, offer hope for useful predictability, a subject of *Saravanan and Chang*'s [2004] chapter in this volume. Indeed, *Hastenrath and Greischar* [1993] and *Folland et al*. [2001] show that northeast Brazil rainfall in boreal spring is quite predictable in their empirical models using SST in the tropical Pacific and Atlantic as predictors. Recent predictability studies using dynamical models support this conclusion and show improved hindcast skills if they are initialized with SSTs in the eastern equatorial Pacific and tropical Atlantic [*Chang et al*., 1998; *Penland and Matrosova*, 1998; *Chang et al*., 2003]. Initial SST anomalies in these regions allow inclusion of ENSO teleconnection and the ITCZ/CESG interaction within the tropical Atlantic, respectively. This result is consistent with diagnostic/modeling studies showing the importance of both ENSO forcing and local air-sea feedback.

A further prerequisite for successful dynamic prediction is the use of an air-sea-land coupled model that is unbiased. Unfortunately, strong biases persist in nearly all current climate models in the tropical Atlantic sector. Chief among these biases are the failure to keep the mean ITCZ north of the equator and to maintain the equatorial cold tongue. In most models, the ITCZ moves back and forth across the equator following the sun, and stays far too long south of the equator. The modeled zonal SST gradient on the equator is opposite to observations, with higher SSTs in the Gulf of Guinea than east of South America. Peculiarly, this reversal of SST gradient occurs despite prevailing easterly winds on the equator in some models [*Davey et al.*, 2002]. The seasonal northward-displacement of the ITCZ and corresponding development of the equatorial cold tongue are features of the seasonal climate that are necessary to the development, structure, and timing of the interannual/decadal TAV, as has been discussed in this review. It is thus a high priority to reduce and remove these biases in climate models.

The tropical Atlantic is a small ocean basin flanked by major continents that host major convection centers of the global atmosphere. Continents exert a strong influence on the annual-mean state and seasonal cycle of the tropical Atlantic, as exemplified by the northward displacement of the climatological ITCZ and the annual cycle in equatorial SST and zonal wind. We also know that interannual variability of the tropical Atlantic exerts a significant effect on the rainfall over both South America and Africa. Unclear is what role the continents play in TAV. Questions that remain to be explored include whether variability on continents provides any feedback to the TAV and to what extent internal variability of the continental monsoon can affect TAV. An accurate representation of the interaction of ocean, atmosphere, and land is imperative for a realistic simulation of the mean state and variability of the tropical Atlantic Ocean.

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REFERENCES

Aceituno, P., On the functioning of the Southern Oscillation in the South American sector: Part I: Surface climate, *Mon. Weather Rev.*, *97*, 505-524, 1988.

Alexander, M. A., and J. D. Scott, The influence of ENSO on air-sea interaction in the Atlantic, *Geophys. Res. Lett., 29*(14), 1701, doi: 10.1029/2001GL014347, 2002.

Alexander, M. A., I. Bladé, M. Newman, J. R. Lanzante, N.-C. Lau, and J. D. Scott, The atmospheric bridge: The influence of ENSO teleconnections on air–sea interaction over the global oceans, *J. Clim.*, *15*, 2205–2231, 2002.

Barreiro, M., A. Giannini, P. Chang, and R. Saravanan, On the role of the Southern Hemisphere atmospheric circulation in tropical Atlantic variability, this volume.

Biasutti, M., D. S. Battisti, and E. S. Sarachik, The annual cycle over the tropical Atlantic South America, and Africa, *J. Clim.*, *16*, 2491-2508, 2003.

Bjerknes, J., Atmospheric teleconnections from the equatorial Pacific, *Mon. Weather Rev.*, *97*, 163-172, 1969.

Black, D. E., L. C. Peterson, J. T. Overpeck, A. Kaplan, M. N. Evans, and M. Kashgarian, Eight centuries of North Atlantic Ocean atmosphere variability, *Science*, *286*, 1709-1713, 1999.

Boyd, A. J., J. Taunton-Clark, and G. P. J. Oberholster, Spatial features of the near-surface and midwater circulation patterns off western and southern South Africa and their role in the life histories of various commercially fished species, *S. Afr. J. Mar. Sci.*, *12*, 189–206, 1992.

Bretherton, C. S., and D. S. Battisti, An interpretation of the results from atmospheric general circulation models forced by the time history of the observed sea surface temperature distribution, *Geophys. Res. Lett., 27*, 767-770, 2000.

Brundrit, G. B., Trends of southern African sea level: Statistical analysis and interpretation, *S. Afr. J. Mar. Sci.*, *16*, 9–17, 1995.

Carton, J. A., and B. Huang, Warm events in the tropical Atlantic, *J. Phys. Oceanogr.*, *24*, 888-903, 1994.

Carton, J. A., X. Cao, B. S. Giese, and A. M. da Silva, Decadal and interannual SST variability in the tropical Atlantic Ocean, *J. Phys. Oceanogr.*, *26*, 1165-1175, 1996.

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

Chang, P., L. Ji, H. Li, C. Penland, and L. Matrosova, Prediction of tropical Atlantic sea surface temperature, *Geophys. Res. Lett.*, *25*, 1193-1196, 1998.

Chang, P., R. Saravanan, L. Ji, and G. C. Hegerl, The effects of local sea surface temperatures on atmospheric circulation over the tropical Atlantic sector, *J. Clim.*, *13*, 2195-2216, 2000.

Chang, P., L. Ji, and R. Saravanan, A hybrid coupled model study of tropical Atlantic variability, *J. Clim.*, *14*, 361-390, 2001.

Chang, P., R. Saravanan, and L. Ji, Tropical Atlantic seasonal predictability: The roles of El Nino remote influence and thermodynamic air-sea feedback, *Geophys. Res. Lett., 30*(10), 1501, doi: 10.1029/2002GL016119, 2003.

Chiang, J. C. H., and A. H. Sobel, Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate, *J. Clim., 15*, 2616-2631, 2002.

Chiang, J. C. H., S. E. Zebiak, and M. A. Cane, Relative roles of elevated heating and sea surface temperature gradients in driving anomalous surface winds over tropical oceans, *J. Atmos. Sci.*, *58*, 1371-1394, 2001.

Chiang, J. C. H., Y. Kushnir, and A. Giannini, Deconstructing Atlantic ITCZ variability: influence of the local cross-equatorial SST gradient, and remote forcing from the eastern equatorial Pacific, *J. Geophys. Res.*, *107*(D1), 4004, doi: 10.1029/2000JD000307, 2002.

Chiang, J. C. H., M. Biasutti, and D. S. Battisti, Sensitivity of the Atlantic ITCZ to conditions during Last Glacial Maximum, *Paleoceanograophy*, *18*(4), 1094, doi: 10.1029/2003PA000916, 2003

Chikamoto, Y., *Tropical Atlantic Ocean-Atmospheric Response to Tropical Pacific SST Variations*, Master's Thesis, pp. 90, Hokkaido University, Sapporo, Japan, 2002.

Chu, P.-S., Time and space variability of rainfall and surface circulation in the northeast Brazil-tropical Atlantic sector, *J. Meteorol. Soc. Jpn.*, *26*, 363-369, 1984.

Chung, C., S. Nigam, and J. A. Carton, SST-forced surface wind variability in the tropical Atlantic: An empirical model, *J. Geophys. Res., 107*(D15), 4244, doi: 10.1029/2001JD000324, 2002.

Covey, D. L., and S. Hastenrath, The Pacific El Nino phenomenon and the Atlantic circulation, *Mon. Weather Rev.*, *106*, 1280-1287, 1978.

Crawford, R. J. M., W. R. Siegfried, L. V. Shannon, C. A. Villacastin-Herroro, and L. G. Underhill, Environmental influences on marine biota off southern Africa, *S. Afr. J. Sci.*, *86*, 330-339, 1990.

Curtis, S., and S. Hastenrath, Forcing of anomalous sea surface temperature evolution in the tropical Atlantic during Pacific warm events. *J. Geophys. Res., 100*, 15,835-15,847, 1995.

Czaja, A., P. van der Vaart, and J. Marshall, A diagnostic study of the role of remote forcing in tropical Atlantic variability, *J. Clim.*, *15*, 3280-3290, 2002.

Davey, M. K., et al., STOIC: a study of coupled model climatology and variability in tropical ocean regions, *Clim. Dyn*., *18*, 403-420, 2002.

Delworth, T. L., and V. M. Mehta, Simulated interannual to decadal variability in the tropical and sub-tropical North Atlantic, *Geophys. Res. Lett. 25*, 2825-2828, 1998.

Dommenget, D., and M. Latif, Interannual to decadal variability in the tropical Atlantic, *J. Clim., 13*, 777-792, 2000.

Dong, B.-W., and R. T. Sutton, Adjustment of the coupled ocean-atmosphere system to a sudden change in the thermohaline circulation, *Geophys. Res. Lett.*, *29*(15), 1728 doi: 10.1029/2002GL015229, 2002.

Enfield, D. B., and D. A. Mayer, Tropical Atlantic SST variability and its relation to El Nino-Southern Oscillation, *J. Geophys. Res.*, *102*, 929-945, 1997.

Enfield, D. B., and A. M. Mestas-Nuez, Multiscale variabilities in global sea surface temperatures and their relationships with tropospheric climate patterns, *J. Clim., 12*, 2719–2733, 1999.

Enfield, D. B., A. M. Mestas-Nunez, D. A. Mayer, and L. Cid-Serrano, How ubiquitous is the dipole relationship in tropical Atlantic sea surface temperatures?, *J. Geophys. Res. 104*, 7841-7848, 1999.

Folland, C. K., T. N. Palmer, and D. E. Parker, Sahel rainfall and world-wide sea temperatures, *Nature*, *320*, 602-607, 1986.

Folland, C. K., A. W. Colman, D. P. Powell, and M. K. Davey, Predictability of northeast Brazil rainfall and real-time forecast skill, 1987-98. *J. Clim.*, *14*, 1937-1958, 2001.

Frankignoul, C., Sea surface temperature anomalies, planetary waves, and air-sea feedback in the middle latitudes, *Rev. Geophys.*, *23*, 357-390, 1985.

Fu, R., R. E. Dickinson, M. Chen, and H. Wang, How do tropical sea surface temperatures influence the seasonal distribution of precipitation in the equatorial Amazon?, *J. Clim.*, *14*, 4003-4026, 2001.

Giannini, A., Y. Kushnir, and M. A. Cane, Interannual variability of Caribbean rainfall, ENSO, and the Atlantic Ocean, *J. Clim.*, *13*, 297-311, 2000.

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

Gray, W. M., and C. W. Landsea, African rainfall as a precursor of hurricane-related destruction on the U.S. east coast, *Bull. Am. Meteorol. Soc*., *75*, 1352-1364, 1992.

Grodsky, S. A., and J. A. Carton, Coupled land/atmosphere interactions in the West African monsoon, *Geophys. Res. Letts.*, *28*, 1503-1506, 2001.

Grodsky S. A., and J. A. Carton, Intertropical Convergence Zone in the South Atlantic and the equatorial cold tongue, *J. Clim.*, *16*, 723-733, 2003.

Grötzner, A., M. Latif, and T. P. Barnett, A decadal climate cycle in the North Atlantic Ocean as simulated by the ECHO coupled GCM, *J. Clim.*, *11*, 831-847, 1998.

Hagen, E., R. Feistel, J. J. Agenbag, and T. Ohde, Seasonal and interannual changes in intense Benguela upwelling (1982–1999), *Oceanol. Acta*, 24, 557-567, 2001.

Hakkinen, S., and K.C. Mo, The low-frequency variability of the tropical Atlantic Ocean, *J. Clim., 15*, 237-250, 2002.

Halley, E., A historical account of the trade winds, and monsoons, observable in the seas between and near the Tropicks, with an attempt to assign the phisical cause of the said winds, *Philos. Trans. R. Soc. London*, *16*, 153-168, 1686.

Hastenrath, S., Interannual variability and annual cycle: Mechanisms of circulation and climate in the tropical Atlantic sector, *Mon. Weather Rev.*, *112*, 1097-1107, 1984.

Hastenrath, S., Decadal-scale changes of the circulation in the tropical Atlantic sector associated with Sahel drought, *Int. J. Climatol.*, *10*, 459-472, 1990.

Hastenrath, S., *Climate dynamics of the tropics*, 488 pp., Kluwer Academic, Boston, USA, 1991.

Hastenrath, S., and L. Greischar, Further work on the prediction of northeast Brazil rainfall anomalies, *J. Clim., 6*, 743-758, 1993.

Hastenrath, S., and L. Heller, Dynamics of climate hazards in Northeast Brazil, *Q. J. R. Meteorol. Soc.*, *103*, 77-92, 1977.

Hirst, A., and S. Hastenrath, Atmosphere-ocean mechanisms of climate anomalies in the Angola-tropical Atlantic sector, *J. Phys. Oceanogr.*, *13*, 1146-1157, 1983.

Hisard, P., Observation de response du type "El Nino" dans l'Atlantique tropical oriental – Golfe de Guinee, *Oceanol. Acta*, *3*, 69-78, 1980.

Horel, J. D., V. E. Kousky, and M. T. Kagano, Atmospheric conditions in the Atlantic sector during 1983 and 1984, *Nature*, *310*, 248–251, 1986.

Houghton, R. W., Seasonal variations of the subsurface thermal structure in the Gulf of Guinea, *J. Phys. Oceanogr*., *13*, 2070-2081, 1983.

Houghton, R. W., The relationship of sea surface temperature to thermocline depth at annual and interannual time scales in the tropical Atlantic Ocean, *J. Geophys. Res.*, *96*, 15173-15185, 1991.

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

Huang, B., P. S. Schopf, and Z. Pan, The ENSO effect on the tropical Atlantic variability: A regionally coupled model study, *Geophys. Res. Lett.*, *29*(21), 2039, doi: 10.1029/2002GL 014872, 2002.

Hurrell, J. W., Y. Kushnir, G. Ottersen, and M. Visbeck (Eds.), *The North Atlantic Oscillation: Climatic Significance and Environmental Impact*, 279 pp., American Geophysical Union, Washington, D.C., 2003.

Inui, T., A. Lazar, P. Malanotte-Rizzoli, and A. Busalacchi, Wind Stress Effects on Subsurface Pathways from the Subtropical to Tropical Atlantic, *J. Phys. Oceanogr*., *32*, 2257–2276, 2002.

Janicot S., and B. Sultan, Intra-seasonal modulation of convection in the West African monsoon, *Geophys. Res. Letts*., *28*, 523-526, 2001.

Jochum, M., and P. Malanotte-Rizzoli, Influence of the meridional overturning circulation on tropical–subtropical pathways, *J. Phys. Oceanogr*., *31*, 1313–1323, 2001.

Jochum, M., R. Murtugudde, P. Malanotte-Rizzoli, and A. J. Busalacchi, Intrinsic variability of the tropical Atlantic Ocean, this volume.

Kalnay, E., et al., The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc*., *77*, 437-471, 1996.

Klein, S. A., B. J. Soden, and N.-C. Lau, Remote sea surface temperature variations during ENSO: Evidence for a tropical atmospheric bridge, *J. Clim., 12*, 917-932, 1999.

Kushnir, Y, R. Seager, J. Miller, and J. C. H. Chiang, A simple coupled model of tropical Atlantic decadal climate variability, *Geophys. Res. Lett. 29*(23), 2133, doi: 10.1029/2002GL 015874, 2002a.

Kushnir, Y, W. A. Robinson, I. Bladé, N. M. J. Hall, S. Peng, and R. Sutton, Atmospheric GCM response to extratropical SST anomalies: Synthesis and evaluation, *J. Clim*., *15*, 2233–2256, 2002b.

Lamb, P. J., and R. A. Peppler, Further case studies of tropical Atlantic surface atmospheric and oceanic patterns associated with sub-Saharan drought, *J. Clim., 5*, 476-488, 1992.

Lanzante, J. R., Lag relationships involving tropical sea surface temperatures, *J. Clim*., *9*, 2568-2578, 1996.

Latif, M., and T. P. Barnett, Interactions of the tropical oceans. *J. Clim*., *8*, 952-964, 1995.

Lau, N.-C., and M. J. Nath, Impact of ENSO on SST variability in the North Pacific and North Atlantic: Seasonal dependence and role of extratropical sea–air coupling, *J. Clim*., *14*, 2846–2866, 2001.

Li, T., and S. G. H. Philander, On the seasonal cycle of the equatorial Atlantic Ocean, *J. Clim*., *10*, 813-817, 1997.

Lin, H., and J. Derome, The atmospheric response to North Atlantic SST anomalies in seasonal prediction experiments, *Tellus*, *55A*, 193-207, 2003.

Lindzen, R. S., and S. Nigam, On the role of sea surface temperature gradients in forcing low-level winds and convergence in the tropics, *J. Atmos. Sci*., *44*, 2418-2436, 1987.

Lough, J. M., Tropical Atlantic sea surface temperature and rainfall variations in subsaharan Africa, *Mon. Weather Rev.*, *114*, 561-570, 1986.

Luksch, U., Simulation of North Atlantic low-frequency SST variability, *J. Clim*., *9*, 2083-2092, 1996.

Marengo, J. A., and S. Hastenrath, Case studies of extreme climatic events in the Amazon basin, *J. Clim*., *6*, 617-627, 1993.

Markham, C. G., and D. R. McLain, Sea surface temperature related to rain in Ceara, Northeastern Brazil, *Nature*, *265*, 320-323, 1977.

Marshall, J., et al., North Atlantic climate variability: Phenomena, impacts and mechanisms, *Int. J. Climatol*., *21*, 1863-1898, 2001.

Mehta, V. M. Variability of the tropical ocean surface temperatures at decadal-multidecadal time scales. Part I: The Atlantic Ocean, *J. Clim.,* **11**, 2351-2375 1998.

Mehta, V. M., M. J. Suarez, J. Manganello, and T. L. Delworth, Oceanic influence on the North Atlantic Oscillation and associated Northern Hemisphere climate variations: 1959-1993, *Geophys. Res. Lett.*, *27*, 121-124, 2000.

Merle, J., Annual and interannual variability of temperature in the eastern equatorial Atlantic Ocean – hypothesis of an Atlantic El Nino, *Oceanol. Acta*, *3*, 209-220, 1980.

Mestas-Nunez, A. M., and D. B. Enfield, Eastern equatorial Pacific SST variability: ENSO and non-ENSO components and their climatic associations, *J. Clim.*, *14*, 391-402, 2001.

Mitchell, T. P., and J. M. Wallace, The annual cycle in equatorial convection and sea surface temperature, *J. Clim.*, *5*, 1140-1156, 1992.

Moura, A., and J. Shukla, On the dynamics of droughts in northeast Brazil: Observations, theory and numerical experiments with a general circulation model, *J. Atmos. Sci., 38*, 2653-2675, 1981.

Namias, J., Influence of northern hemisphere general circulation on drought in Northeast Brazil, *Tellus*, *24*, 336-42, 1972.

Neelin, J. D., D. S. Battisti, A. C. Hirst, F.-F. Jin, Y. Wakata, T. Yamagata, and S. E. Zebiak, ENSO theory, *J. Geophys. Res.*, *103*, 14,261-14,290, 1998.

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

Okajima, H., S.-P. Xie, and A. Numaguti, Interhemispheric coherence of tropical climate variability: Effect of climatological ITCZ, *J. Meteorol. Soc. Jpn.*, *81*, 1371-1386, 2003.

Okumura, Y. and S.-P. Xie, Interaction of the Atlantic equatorial cold tongue and African monsoon, *J. Clim.*, in press, 2004.

Okumura, Y., S.-P. Xie, A. Numaguti, and Y. Tanimoto, Tropical Atlantic air-sea interaction and its influence on the NAO, *Geophys. Res. Lett.*, *28*, 1507-1510, 2001.

Parker, D. E., P. D. Jones, A. Bevan, and C. K. Folland, Interdecadal changes of surface temperature since the late 19th century, *J. Geophys. Res.*, *99*, 14373-14399, 1994.

Peng, S., W. A. Robinson, and S. Li, North Atlantic SST forcing of the NAO and relationships with intrinsic hemispheric variability. *Geophys. Res. Lett.*, *29*(8), 1276, doi: 10.1029/2001 GL014043, 2002.

Penland, C., and L. Matrosova, Prediction of tropical Atlantic sea surface temperatures using linear inverse modeling, *J. Clim.*, *11*, 483-496, 1998.

Peterson, L. C., G. H. Huag, K. A. Hughen, and U. Rohl, Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial, *Science*, *290*, 1947-1951, 2000.

Philander, S. G. H., Unusual conditions in the tropical Atlantic Ocean in 1984, *Nature*, *222*, 236-238, 1986.

Philander, S. G. H., and R. C. Pacanowski, The oceanic response to cross-equatorial winds (with application to coastal upwelling in low latitudes), *Tellus*, *33*, 201-210, 1981.

Philander, S. G. H., and R. C. Pacanowski, A model of the seasonal cycle in the tropical Atlantic Ocean, *J. Geophys. Res.*, *91*, 14192-14206, 1986.

Rajagopalan, B., Y. Kushnir, and Y. Tourre, Observed decadal mid-latitude and tropical Atlantic climate variability, *Geophys. Res. Lett.*, *25*, 3967-3970, 1998.

Rao, V. B., K. Hada, and D. L. Herdies, On the severe drought of 1993 in Northeast Brazil, *Int. J. Climatol.*, *15*, 697-704, 1995.

Robertson, A. W., C. R. Mechoso, and Y.-J. Kim, The influence of Atlantic sea surface temperature anomalies on the North Atlantic Oscillation, *J. Clim., 13*, 122-138, 2000.

Rodwell, M. J., On the predictability of North Atlantic climate, in *The North Atlantic Oscillation: Climatic Significance and Environmental Impact,* edited by J. W. Hurrell et al., pp. 173-192, American Geophysical Union, Washington, D.C., 2003.

Rodwell, M. J., D. P. Rowell, and C. K. Folland, Oceanic forcing of the wintertime North Atlantic Oscillation and European climate, *Nature*, *398*, 320-323, 1999.

Ruiz-Barradas, A., J. A. Carton, and S. Nigam, Structure of interannual-to-decadal climate variability in the tropical Atlantic sector, *J. Clim.*, *13*, 3285-3297, 2000.

Ruiz-Barradas, A., J. A. Carton, and S. Nigam, Role of the atmosphere in climate variability of the tropical Atlantic, *J. Clim.*, *16*, 2052-2065, 2003.

Saravanan, R., and P. Chang, Interaction between tropical Atlantic variability and El Nino-Southern Oscillation, *J. Clim.*, *13*, 2177-2194, 2000.

Saravanan, R., and P. Chang, Thermodynamic coupling and predictability of tropical sea surface temperature, this volume.

Schott, F. A., J. P. McCreary, and G. C. Johnson, Shallow overturning circulations of the tropical-subtropical oceans, this volume.

Seager, R., Y. Kushnir, P. Chang, N. Naik, J. Miller, and W. Hazeleger, Looking for the role of the ocean in tropical Atlantic decadal climate variability, *J. Clim.*, *14*, 638-655, 2001.

Servain, J., Simple climate indices for the tropical Atlantic Ocean and some applications, *J. Geophys. Res.*, *96*, 15,137-15,146, 1991.

Servain, J., J. Picaut, and J. Merle, Evidence of remote forcing in the equatorial Atlantic Ocean, *J. Phys. Oceanogr.*, *12*, 457–463, 1982.

Servain, J., I. Wainer, J. P. McCreary, and A. Dessier, Relationship between the equatorial and meridional modes of climatic variability in the tropical Atlantic, *Geophys. Res. Lett.*, *26*, 485-488, 1999.

Su, H., J. D. Neelin, and J. E. Meyerson, Tropical tropospheric temperature and precipitation response to sea surface temperature forcing, this volume.

Sutton, R. T., S. P. Jewson, and D. P. Rowell, The elements of climate variability in the tropical Atlantic region, *J. Clim.*, *13*, 3261-3284, 2000.

Sutton, R. T., W. A. Norton, and S. P. Jewson, The North Atlantic Oscillation—What role for the ocean?, *Atmospheric Science Letters*, *1*(2), doi: 10.1006/asle.2000.0018, 2001.

Tanimoto, Y., and S.-P. Xie, Ocean-atmosphere variability over the pan-Atlantic basin, *J. Meteorol. Soc. Jpn.*, *77*, 31-46, 1999.

Tanimoto, Y., and S.-P. Xie, Inter-hemispheric decadal variations in SST, surface wind, heat flux and cloud cover over the Atlantic Ocean, *J. Meteorol. Soc. Jpn.*, *80*, 1199-1219, 2002.

Taylor, M. A., D. B. Enfield, and A. A. Chen, Influence of the tropical Atlantic versus the tropical Pacific on Caribbean rainfall, *J. Geophys. Res.*, *107*(C9), 3127 doi: 10.1029/2001JC 001097, 2002.

Terray, L., and C. Cassou, Tropical Atlantic sea surface temperature forcing of quasi-decadal climate variability over the North Atlantic-European region, *J. Clim., 15*, 3170-3187, 2002.

Thorncroft, C. D., et al., The JET 2000 Project: Aircraft observations of the African easterly jet and African easterly waves, *Bull. Am. Meteorol. Soc.*, *86*, 337-351, 2003.

Tourre, Y. M., B. Rajagopalan, and Y. Kushnir, Dominant patterns of climate variability in the Atlantic Ocean during the last 136 years, *J. Clim.*, *12*, 2285-2299, 1999.

Venegas, S. A., L. A. Mysak, and D. N. Straub, Atmosphere-ocean coupled variability in the South Atlantic, *J. Clim.*, *10*, 2904-2920, 1997.

Venzke, S., M. R. Allen, R. T. Sutton, and D. P. Rowell, The atmospheric response over the North Atlantic to decadal changes in sea surface temperature, *J. Clim., 12*, 2562-2584, 1999.

Vizy, E. K., and K. H. Cook, Development and application of a mesoscale climate model for the tropics: Influence of sea surface temperature anomalies on the West African monsoon. *J. Geophys. Res*., *107*(D3), 4023, 10.1029/2001JD000686, 2002.

Wagner, R. G., Mechanisms controlling variability of the interhemispheric sea surface temperature gradient in the tropical Atlantic, *J. Clim.*, *9*, 2010-2019, 1996.

Wagner, R. G., and A. da Silva, Surface conditions associated with anomalous rainfall in the Guinea coastal region, *Int. J. Climatol.*, *14*, 179-200, 1994.

Wang, J., and J.A. Carton, Modeling climate variability in the tropical Atlantic atmosphere, *J. Clim.*, *16*, 3858–3876, 2003.

Watanabe, M., and M. Kimoto, Tropical-extratropical connection in the Atlantic atmosphere-ocean variability, *Geophys. Res. Lett.*, *26*, 2247-2250, 1999.

Watanabe, M., M. Kimoto, T. Nitta, and M. Kachi, A comparison of decadal climate oscillations in the North Atlantic detected in observations and a coupled GCM, *J. Clim.*, *12*, 2920–2940, 1999.

Weare, B. C., Empirical orthogonal analysis of Atlantic Ocean surface temperatures, *Q. J. R. Meteorol. Soc.*, *103*, 467-478, 1977.

Wu, L., and Z. Liu, Is tropical Atlantic variability driven by the North Atlantic Oscillation? *Geophys. Res. Lett. 29*(13), 1653, 10.1029/2002GL014939, 2002.

Wu, L., Q. Zhang, and Z. Liu, Toward understanding the tropical Atlantic variability using coupled modeling surgery, this volume.

Woodruff, S. D., R. J Slutz, R. L. Jenne, and P. M. Steurer, A comprehensive ocean-atmosphere dataset, *Bull. Am. Meteorol. Soc*., *68*, 521-527, 1987.

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

Xie, S.-P., On the genesis of the equatorial annual cycle. *J. Clim.*, *7*, 2008-2013, 1994.

Xie, S.-P., A dynamic ocean-atmosphere model of the tropical Atlantic decadal variability, *J. Clim.*, *12*, 64-70, 1999.

Xie, S.-P., The shape of continents, air-sea interaction, and the rising branch of the Hadley circulation, in *The Hadley Circulation: Past, Present and Future*, edited by H. F. Diaz and R. S. Bradley, Kluwer Academic Publishers, Dordrecht, in press, 2004.

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

Xie, S.-P., and Y. Tanimoto, A pan-Atlantic decadal climate oscillation, *Geophys. Res. Lett.*, *25*, 2185-2188, 1998.

Xie, S.-P., Y. Tanimoto, H. Noguchi, and T. Matsuno, How and why climate variability differs between the tropical Pacific and Atlantic, *Geophys. Res. Lett.*, *26*, 1609-1612, 1999.

Yang, J.Y., A linkage between decadal climate variations in the Labrador Sea and the tropical Atlantic Ocean, *Geophys. Res. Lett.*, *26*, 1023-1026, 1999.

Yulaeva, E., and J. M. Wallace, The signature of ENSO in global temperature and precipitation fields derived from the microwave sounding unit, *J. Clim.*, *7*, 1719-1736, 1994.

Zebiak, S. E., Air-sea interaction in the equatorial Atlantic region, *J. Clim.*, *6*, 1567-1586, 1993.

Zebiak, S. E., and M. A. Cane, A model El Nino-Southern Oscillation, *Mon. Weather Rev.*, *115*, 2262-2278, 1987.

Zhou, Z., and J. A. Carton, Latent heat flux and interannual variability of the coupled atmosphere-ocean system, *J. Atmos. Sci.*, *55*, 494-501, 1998.

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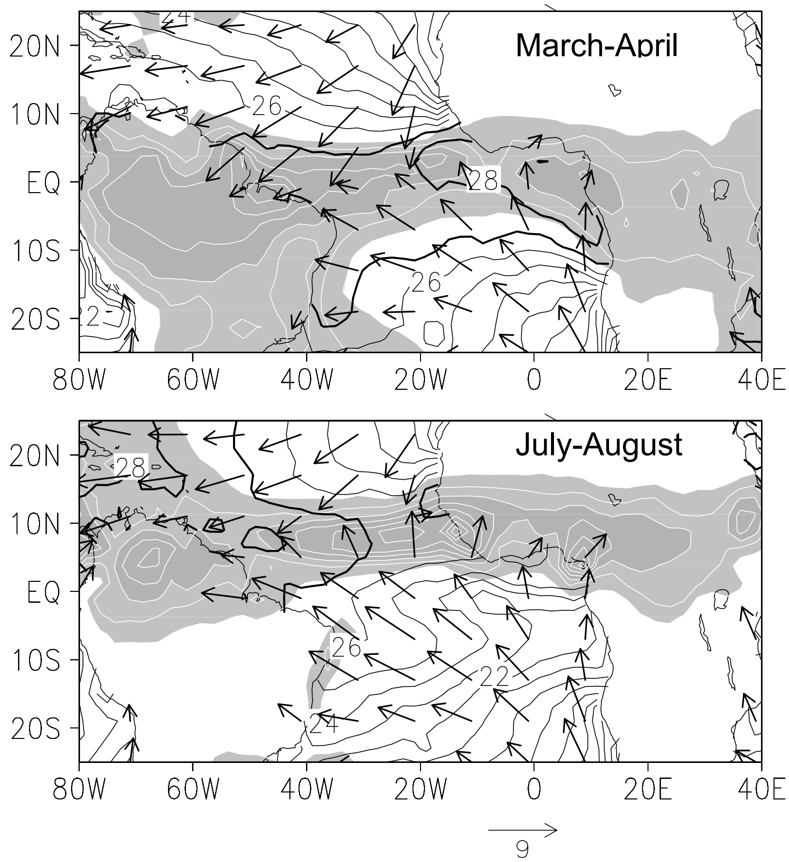

Figure 1. Climatological distributions of rainfall (light shade > 2mm/day; dark shade > 6 mm/day), SST (contours in ^{o}C) and surface wind velocity (vectors in m/s) for March-April (upper) and July-August (lower panel), based on the Climate Prediction Center Merged Analysis of Precipitation (CMAP; *Xie and Arkin* 1996) and Comprehensive Ocean-Atmospheric Data Set (COADS; *Woodruff et al*. 1987).

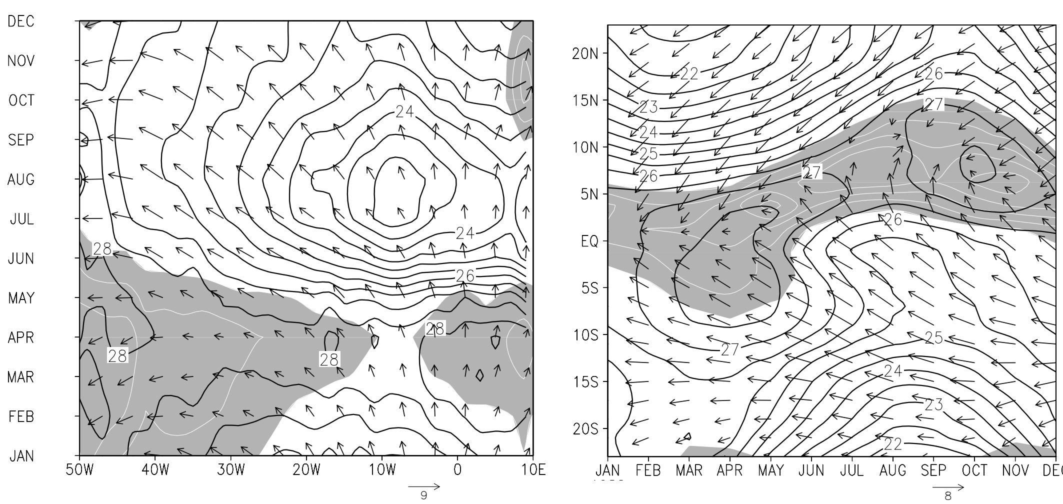

Figure 2. Left: longitude-time sections of COADS SST (black contours in ^{o}C) and surface wind velocity (vectors in m/s) at 1^{o}S, and CMAP rainfall in 1.25^{o}S-1.25^{o}N. Right: time-latitude section of SST, surface wind velocity, and rainfall, averaged in 30-25^{o}W. Rainfall are in white contours at 2.5 mm/day intervals with shade > 5 and 2.5 mm/day in the left and right panels, respectively.

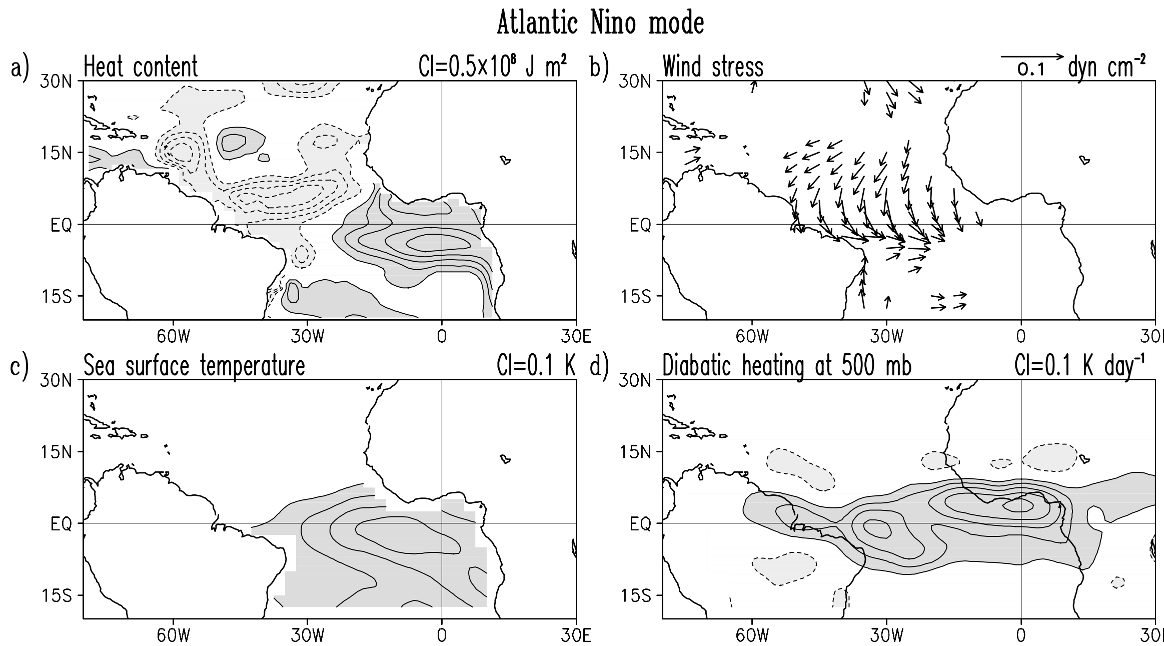

Figure 3. Anomalies associated with the Atlantic Nino principal component from a five-variable rotated principal component analysis: observed heat content (10^8 J m^{-2}, upper left), vector wind stress (dyn cm^{-2}, upper right), SST (^{o}C, lower left), and diabatic heating at 500 mb (^{o}C day^{-1}, lower right). Dark (light) shading denotes positive (negative) anomalies, with the zero contours omitted. Contour intervals (CI) are shown at the upper right corner. From *Ruiz-Barradas et al*. [2000].

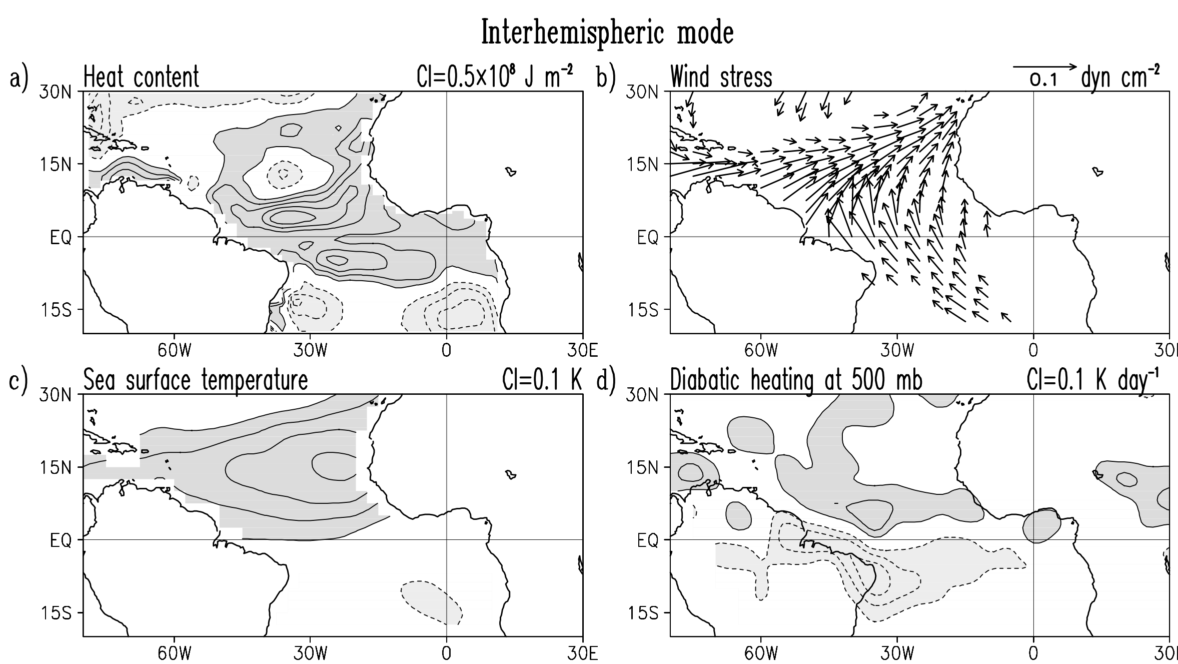

Figure 4. Same as Fig. 3 except for the meridional mode. From *Ruiz-Barradas et al*. [2000].

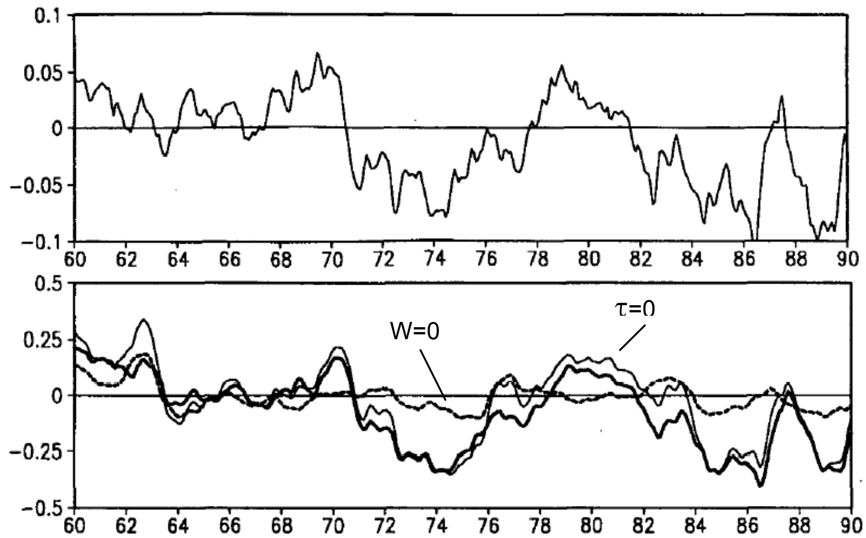

Figure 5. Time series of 12-month smoothed interhemispheric differences in surface wind stress (upper panel in 10^{-1} Nm^{-2}) and SST (lower in ^{o}C). In the lower panel, results from the full simulation is in thick solid line, from runs removing wind variability in latent heat and momentum fluxes are in thin solid and dashed lines, respectively. From *Carton et al*. [1996].

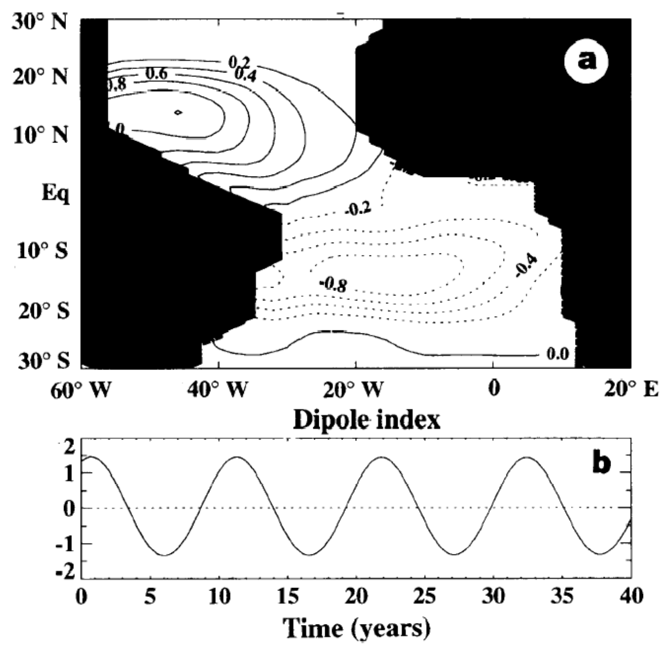

Figure 6. SST regression (upper panel) against a CESG index (lower) in *Chang et al*.'s [1997] intermediate coupled model.

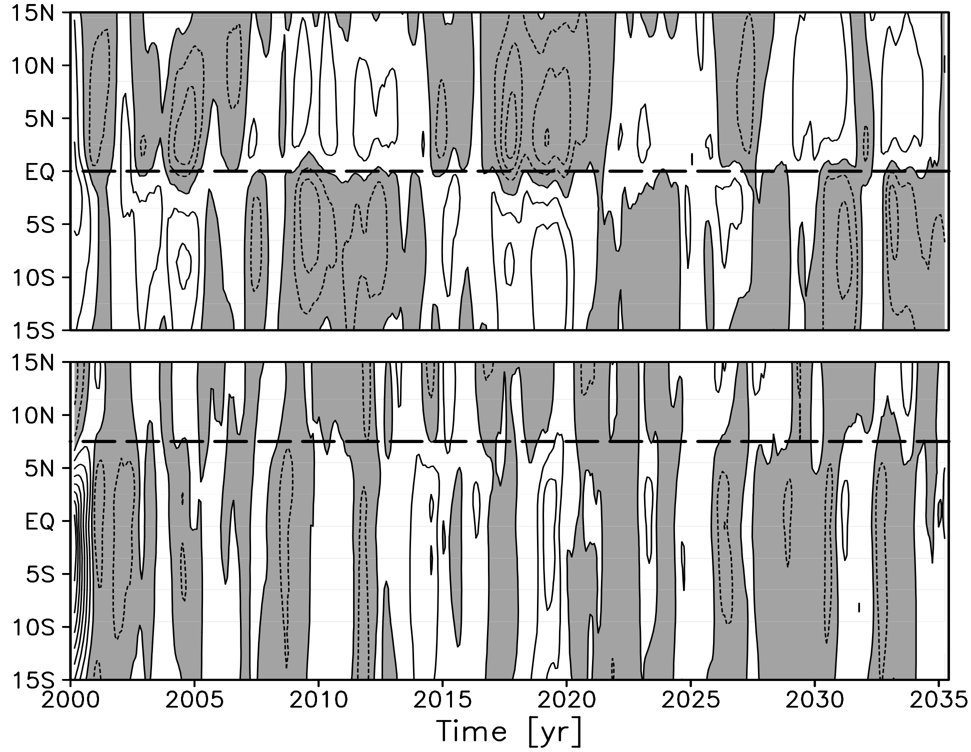

Figure 7. Time-latitude sections of zonal-mean SST anomaly (contours at 0.3°C interval, with negative values shaded) in a coupled model. The thick dashed line indicates the latitude of the climatological ITCZ, which is symmetric about and displaced north of the equator in the upper and lower panels, respectively. From [*Okajima et al*., 2003].

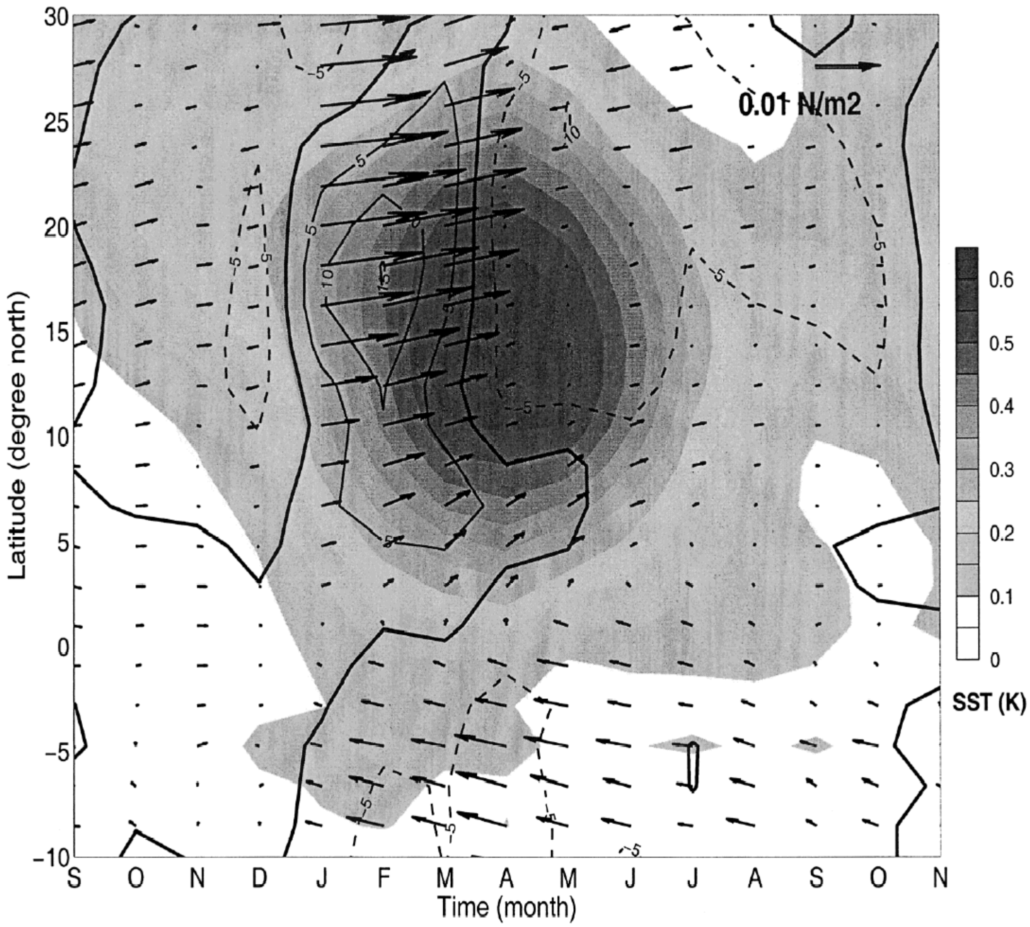

Figure 8. Regression map of surface wind stress (arrows, scale in the top-right corner); net surface heat flux (contoured every 5 W m^{-2}, positive into the ocean, dashed when negative, zero contour thickened); and SST (shaded, in K) onto the NTA SST index time series in MAM. From *Czaja et al*. [2002].

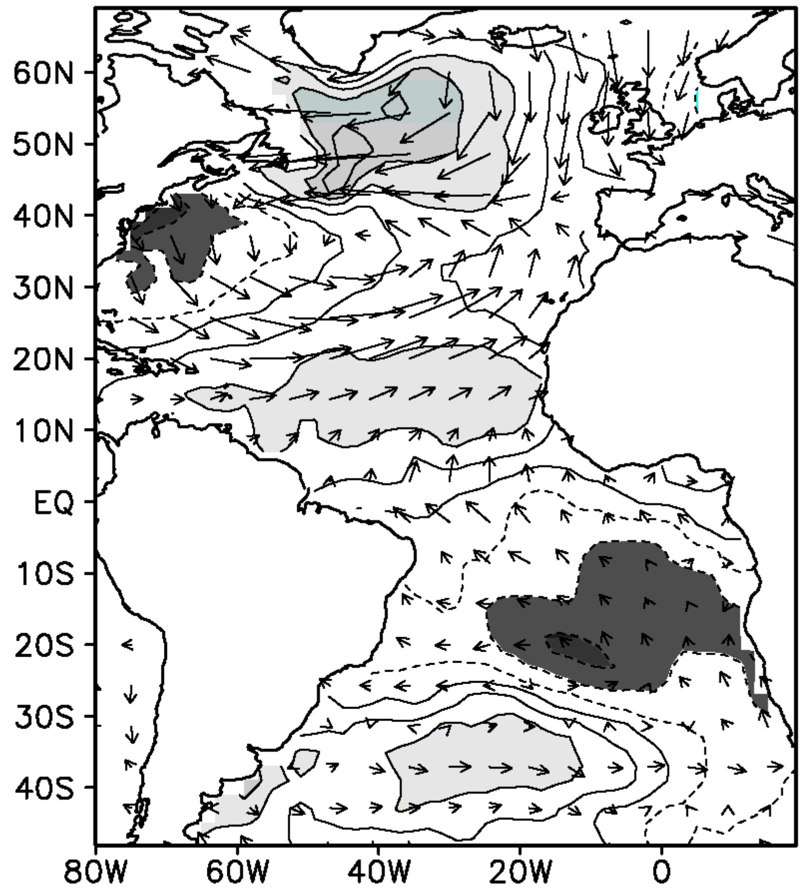

Figure 9. Composite anomalies of SST and surface wind velocity in boreal winter based on a pan-Atlantic decadal oscillation index of *Tanimoto and Xie* [2002]. Global Sea Ice and SST (GISST) dataset [*Parker et al*., 1994] and National Centers for Environmental Prediction (NCEP) Reanalysis [*Kalnay et al*., 1996] are used for SST and surface wind, respectively.

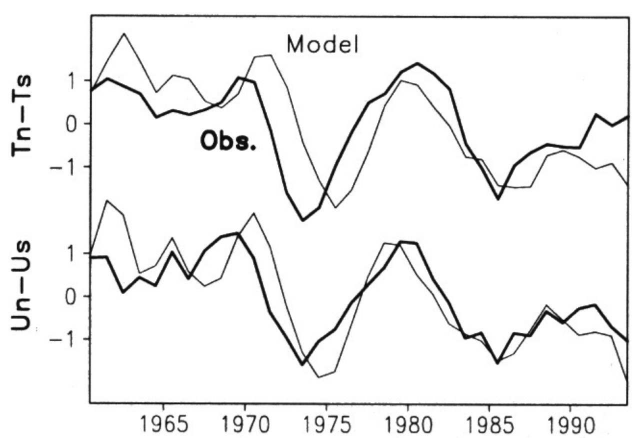

Figure 10. Interhemispheric differences in SST and surface zonal wind velocity in COADS observations (thick) and simulated by a coupled model (thin). All the time series are normalized by their respective standard deviations. From *Xie and Tanimoto* [1998].

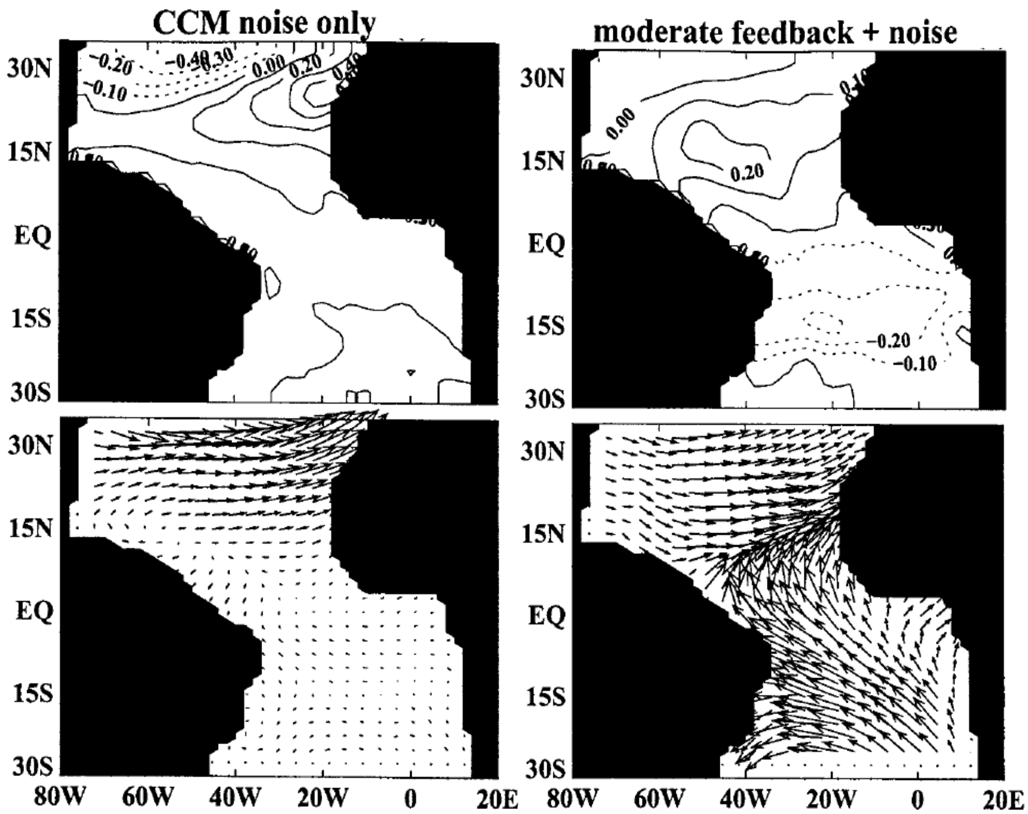

Figure 11. The leading joint SVD modes of SST (upper panels), surface wind stress (lower), and heat flux in coupled model runs forced by atmospheric noise, without (left panels) and with (right) feedbacks onto the atmosphere. From *Chang et al*. [2001], with panels for heat flux omitted.

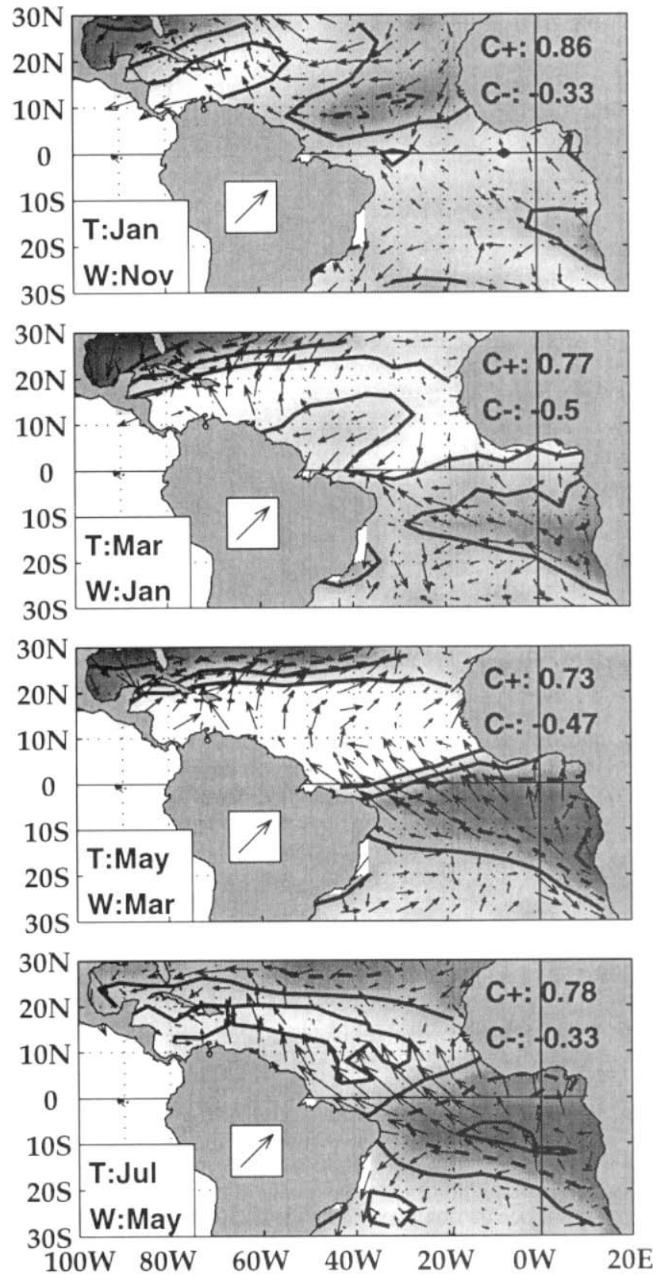

Figure 12. Sequence of development of SST (T) and surface wind (W) correlations in the tropical Atlantic in response to ENSO. The zero contours are dashed, and negative values are dark shaded. Contour intervals are 0.25 (95% significance = 0.25). From *Enfield and Mayer* [1997].

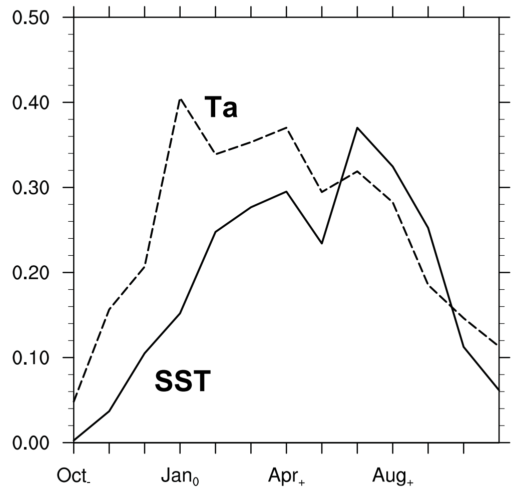

Figure 13. Composite anomalies of SST and surface air temperature averaged in the Caribbean Sea (80°W-60°W, 10°N-20°N), associated with an El Nino in the Pacific. GISST and NCEP reanalysis for 1948-99 are used. From *Chikamoto and Tanimoto* [2004].