

The ENSO Teleconnection to the Tropical Atlantic Ocean: Contributions of the Remote and Local SSTs to Rainfall Variability in the Tropical Americas*

ALESSANDRA GIANNINI, JOHN C. H. CHIANG, MARK A. CANE, YOCHANAN KUSHNIR, AND RICHARD SEAGER

Lamont-Doherty Earth Observatory, Columbia University, Palisades, New York

(Manuscript received 8 March 2001, in final form 5 July 2001)

ABSTRACT

Recent developments in Tropical Atlantic Variability (TAV) identify the El Niño–Southern Oscillation (ENSO) as one of the leading factors in the interannual climate variability of the basin. An ENSO event results in Tropic-wide anomalies in the atmospheric circulation that have a direct effect on precipitation variability, as well as an indirect effect, that is, one mediated by sea surface temperature (SST) anomalies generated in the remote ocean basins. In order to separate the relative contributions of the atmospheric and oceanic components of the ENSO teleconnection to the tropical Atlantic Ocean, results from two ensembles of atmospheric general circulation model (AGCM) experiments, differing in oceanic boundary conditions, are compared. AGCM integrations performed with the Community Climate Model version 3 (CCM3), forced by global, observed SST during 1950–94 reproduce the observed ENSO-related rainfall anomalies over the tropical Americas and adjacent Atlantic. A parallel ensemble of integrations, forced with observed SST in the tropical Atlantic only, and climatology elsewhere, is used to separate the effect of the direct atmospheric teleconnection from the atmosphere's response to the ENSO-forced SST anomalies in the Atlantic basin.

It is found that ENSO-related atmospheric and oceanic anomalies force rainfall anomalies of the same sign in northeast Brazil, of opposite sign in the Caribbean basin. The direct atmospheric influence of a warm ENSO event reduces model rainfall as a whole over the tropical Atlantic basin. This observation is consistent with the hypothesis that an ENSO-related Tropic-wide warming of the free troposphere forces the vertical stabilization of the tropical atmosphere. ENSO-related atmospheric anomalies are also known to force a delayed (relative to the mature phase of ENSO) warming of tropical North Atlantic SST through the weakening of the northeasterly trade winds and consequent reduction of surface fluxes. It is found that this delayed oceanic component forces a northward displacement of the Atlantic intertropical convergence zone, resulting in increased precipitation over the Caribbean and reduced precipitation over northeast Brazil during the boreal spring following the mature phase of ENSO.

1. Introduction

ENSO's impact on tropical Atlantic climate has received increasing attention in recent years. Studies have investigated the association of anomalously warm SST in the central and eastern equatorial Pacific with warmer than average tropical North Atlantic SSTs (Curtis and Hastenrath 1995; Enfield and Mayer 1997), and with rainfall anomalies in the tropical Atlantic, the Caribbean, and Central America (Enfield 1996; Waylen et al. 1996; Chen et al. 1997; Enfield and Alfaro 1999; Giannini et al. 2000, hereafter referred to as GKC), and the Nordeste region of Brazil (Nobre and Shukla 1996; Uvo et al. 1998; Chiang et al. 2000a, 2000b). The link between ENSO and West African rainfall has also

gained more credence (Janicot et al. 1998; Ward 1998). With the exception of the Nordeste (Hastenrath and Heller 1977; Hastenrath et al. 1987; Aceituno 1988; Ward and Folland 1991; Hameed et al. 1993) and northern South America (Ropelewski and Halpert 1987, 1989, 1996; Kiladis and Diaz 1989), ENSO's impact on these regions had not been previously considered reliable. Now the ENSO teleconnection to the Atlantic is coming to be recognized as a fundamental mode of Atlantic climate variability (Saravanan and Chang 2000; Sutton et al. 2000).

Previous studies of the ENSO teleconnection focused on the impact of the tropical Pacific warming on the atmospheric circulation of the Northern Hemisphere extratropics during winter (see reviews in Glantz et al. 1991; Trenberth et al. 1998; Wallace et al. 1998, and references therein). Because the extratropical atmosphere exhibits a high degree of internal variability, especially during winter, connecting its dynamics to a steady tropical heat source like the ENSO system promised increased predictability.

* Lamont-Doherty Earth Observatory Contribution Number 6241.

Corresponding author address: Dr. Alessandra Giannini, Advanced Study Program, NCAR, P. O. Box 3000, 1850 Table Mesa Dr., Boulder, CO 80307-3000.
E-mail: alesall@ucar.edu

In fact, the tropical atmosphere responds almost simultaneously to the establishment of anomalous SSTs in the central and eastern equatorial Pacific (Fig. 1). This adjustment manifests itself as a zonal seesaw in sea level pressure (SLP), active from the summer of year (0),¹ shortly after the onset of an ENSO event, through the mature phase during the winter of year (+1). In a warm ENSO event SLP is lower than average over the warm waters of the central and eastern equatorial Pacific, and higher than average over the equatorial Atlantic (Fig. 1; GKC). Related to this anomalous seesaw in SLP is a weakening of the meridional SLP gradient in both hemispheres of the Atlantic basin, consistent with reduced trade winds at subtropical latitudes (Fig. 1, right column). During boreal winter, the weakening of the meridional SLP gradient is reinforced at subtropical latitudes in the Northern Hemisphere by a low pressure center over the southeastern United States and southwestern North Atlantic, related to the Pacific–North American (PNA) wave train (Horel and Wallace 1981; Wallace and Gutzler 1981). The end result is the oceanic component of the teleconnection: a warming of the tropical North Atlantic Ocean surface that peaks in boreal spring, that is, with a delay of about one season with respect to the attainment in winter of maximum SST anomalies in the tropical Pacific (Enfield and Mayer 1997; Fig. 1, bottom-left panel).

Caribbean/Central American rainfall is reduced during summer and fall of year (0) of a warm ENSO event (GKC). The circulation in the basin is divergent at the surface, along a southwest to northeast axis, with convergence to the southwest, onto the eastern Pacific ITCZ, and weakened northeasterly trade winds in the North Atlantic (Fig. 2). During winter, dry conditions persist over northern South America, while PNA-related wetter than average conditions extend from the southeastern United States to Cuba and the Yucatan peninsula in Mexico. In spring of year (+1), at the start of a new rainy season, rainfall anomalies in the Caribbean reverse in sign. The warmer than average SSTs in the tropical North Atlantic are associated with an anomalously northern location of the Atlantic ITCZ; hence, wetter than average conditions in the Caribbean, and drier than average conditions in the Nordeste region of Brazil during the local rainy season. Dry conditions in the Nordeste can be exacerbated by the atmospheric circulation associated with persistent anomalous SSTs in the central and eastern equatorial Pacific, as in the prolonged ENSO

events of the most recent two decades (Chiang et al. 2000b). Hence ENSO affects rainfall in the tropical Americas in two ways—directly via an “atmospheric bridge,” and indirectly by first changing tropical Atlantic SSTs that subsequently impact tropical Atlantic rainfall.

In this study we compare two ensembles of AGCM integrations differing in boundary conditions to show that the ENSO-related atmospheric bridge, in its tropical (the SLP seesaw, GKC) and extratropical (the PNA, Lau and Nath 1994, 1996; Lau 1997) components, and remotely forced Atlantic SSTs play complementary roles in forcing rainfall anomalies over the tropical Americas. The two mechanisms can act constructively (e.g., northeast Brazil), or destructively (e.g., the Caribbean). This study presents many similarities, but also interesting interpretational differences with the study of Saravanan and Chang (2000, hereafter SC). The ensembles of integrations analyzed are the same. We follow the evolution of the mechanisms by which ENSO affects rainfall variability in the tropical Americas during the course of an entire ENSO life cycle, thereby extending SC to seasons different from the spring immediately following the ENSO event. The summer of year (0) and the winter of year (+1) are best suited to characterize the tropical and extratropical components of the atmospheric bridge, respectively. In the real world, this atmospheric bridge is also responsible for the development of anomalous SSTs in the tropical North Atlantic that peak during spring of year (+1; Enfield and Mayer 1997). We caution that the choice made in SC of the northern tropical Atlantic (NTA) SST index as representing tropical Atlantic variability might not be the most appropriate, since NTA is not independent of ENSO, especially in boreal spring (Giannini et al. 2001, hereafter GKC). We include the SST anomalies in the tropical North Atlantic that follow an ENSO event in our picture of ENSO’s impact on tropical Atlantic climate.

The model used and methodology followed are outlined in section 2. Features of the model’s simulation of the annual cycle of rainfall are briefly compared to observations in section 3. The main results from a comparison of ensembles of integrations forced with different oceanic boundary conditions are described in section 4. A summary and conclusions are drawn in section 5.

2. Data and methods

a. Model

We use the Community Climate Model, version 3 (CCM3), a state-of-the-art atmospheric general circulation model developed at the National Center for Atmospheric Research (NCAR), and described in detail in Kiehl et al. (1998). The integrations used triangular truncation at spectral wavenumber 42 in the horizontal, and 18 vertical levels. For each configuration of SST boundary conditions, five integrations were performed (SC),

¹ Throughout this study we use the phrases “warm ENSO event,” and “El Niño” interchangeably, to denote the warm phase of the ENSO phenomenon, characterized by positive SST anomalies in the central and eastern equatorial Pacific, the Niño-3 region, between 5°S and 5°N, 150° and 90°W. “Cold ENSO event” and “La Niña” refer to the opposite sign SST anomaly. We follow Rasmusson and Carpenter (1982) in denoting by “year (0)” the year of the onset of the warm (cold) event, which typically starts in the spring, and by “year (+1),” the year that begins with the mature phase and soon after witnesses the decay of the event.

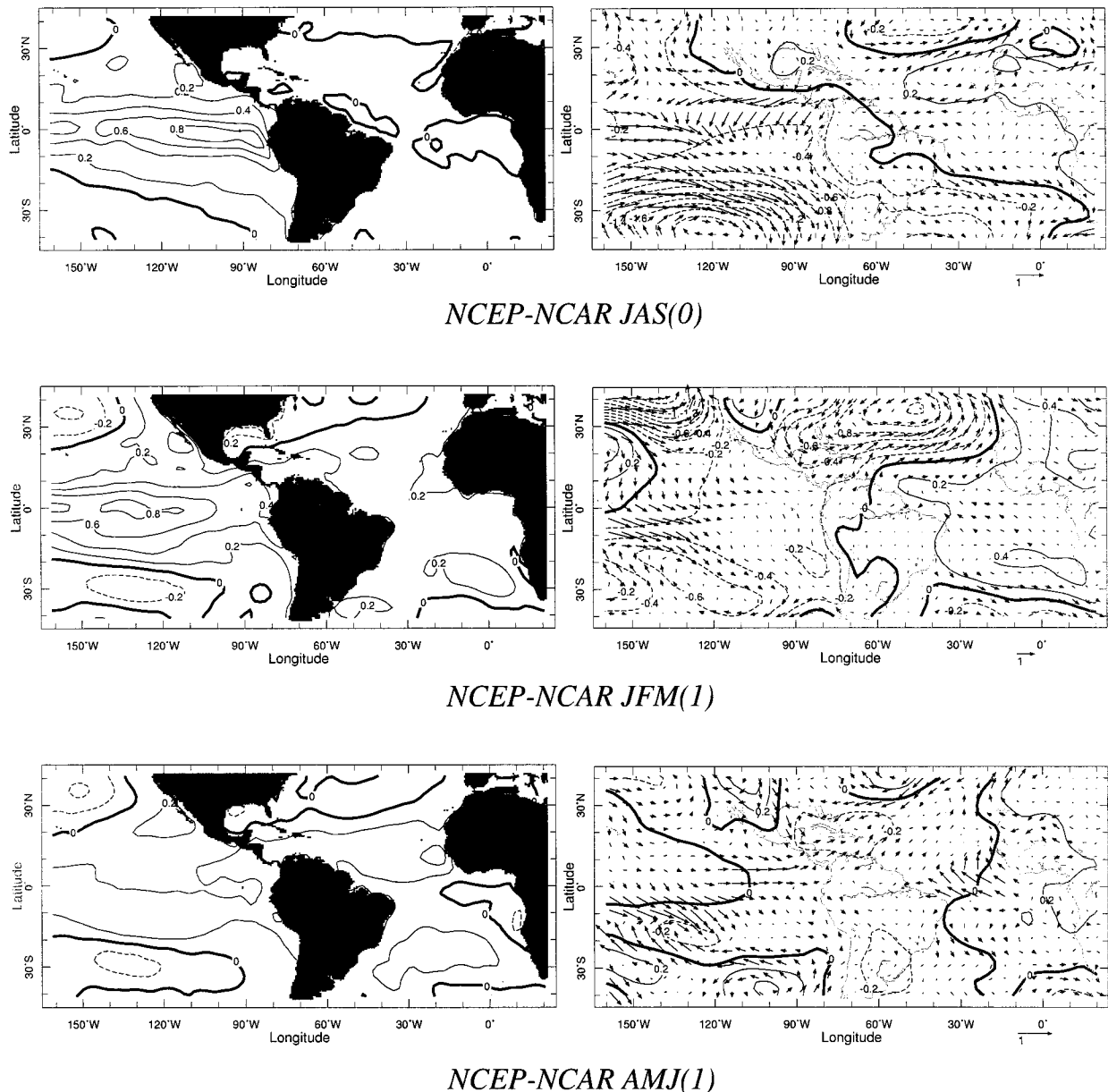


FIG. 1. Regressions of the NCEP-NCAR reanalysis (Kalnay et al. 1996) surface temperature (left column, contours), and surface winds (right column, vectors) onto the Dec (0)–Jan (+1) Niño-3 index (1949–99) (top panels) Jul–Sep (0), (middle panels) Jan–Mar (+1), (bottom panels) Apr–Jun (+1). Contours are 0.2°C in the left column, 0.2 mb in the right column.

each 45 yr long (1950–94). The two configurations were the Global Ocean Global Atmosphere (GOGA), where the atmosphere is coupled to observed SSTs (Smith et al. 1996) prescribed everywhere, and the Tropical Atlantic Global Atmosphere (TAGA), where the atmosphere is coupled to observed SST in the tropical Atlantic only, and to monthly climatology elsewhere. In TAGA the full, interannually varying SST were prescribed between 20°S and 20°N , and smoothed to climatology in a 10° latitudinal band. The two ensembles have already been analyzed by Saravanan and Chang

(SC; Chang et al. 2000), who kindly provided access to the output.

Interensemble discrepancies in the boundary conditions in the tropical Atlantic, where the anomalies should be exactly the same by experimental design, were noted only after completion of the integrations. They were a posteriori explained as an undesired outcome of the interpolation scheme used to smooth tropical Atlantic observations to the climatological fields imposed elsewhere. In briefly reporting on these discrepancies, which will be made the subject of some speculation in

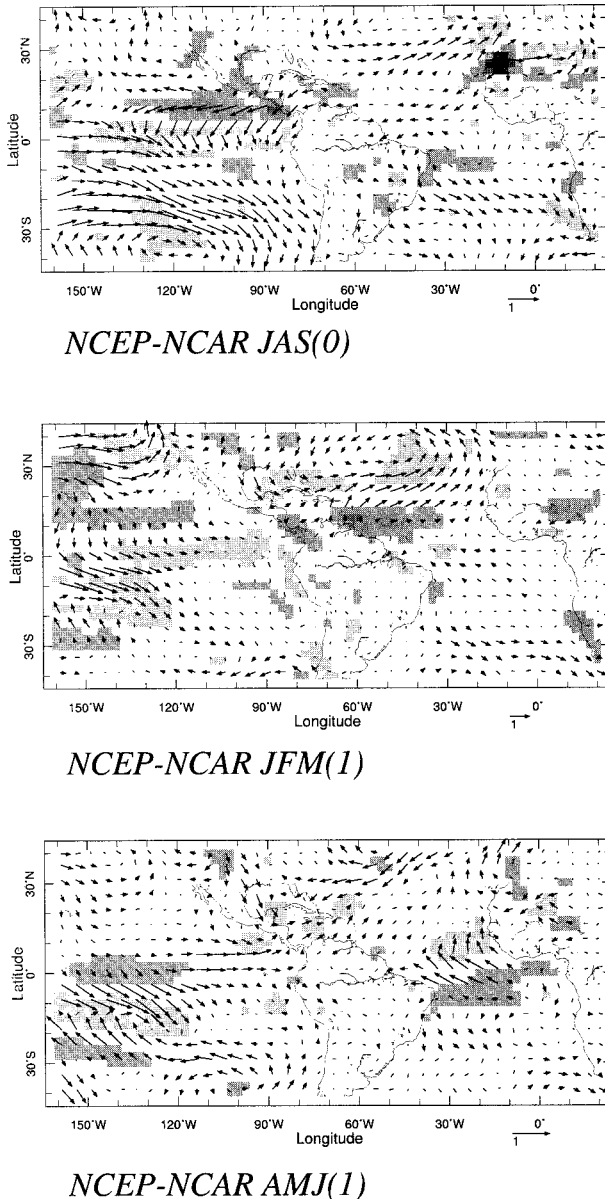


FIG. 2. Regressions of the NCEP-NCAR reanalysis (Kalnay et al. 1996) surface divergence (shading) and surface winds (vectors) onto the Dec (0)–Jan (+1) Niño-3 index (1949–99). (top panels) Jul–Sep (0), (middle panels) Jan–Mar (+1), (bottom panels) Apr–Jun (+1). Light shading indicates negative values, or convergence; dark shading indicates positive values, or divergence.

the continuation of this study, we note that these were corrected in integrations carried out subsequently by Saravanan and Chang, and that in a preliminary analysis these corrections do not appear to affect our conclusions (R. Saravanan 2001, personal communication). Composites of the anomalies of warm and cold ENSO events during the spring following mature ENSO conditions, the season when local SSTs are hypothesized to play a relevant role in rainfall variability in the tropical Americas, are shown in Fig. 3. Tropical North Atlantic anom-

alies are of the same sign as the equatorial Pacific anomalies, but are much smaller and peak with a delay of about a season (Enfield and Mayer 1997) in spring (+1). GOGA–TAGA differences are largest in the western equatorial Atlantic and along the northern coast of South America. Although these differences are typically of the order of 0.1°C or less, they occur in a very sensitive region, as will be discussed in section 4c.

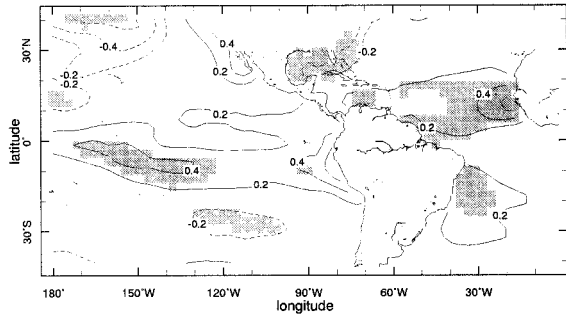
We compare model output to the monthly precipitation dataset of Xie and Arkin (1996, 1997), which covers the period 1979 to the present, and was obtained by merging satellite and rain gauge observations [provided by the National Oceanic and Atmospheric Administration's (NOAA) Climate Prediction Center, and available online at <http://ingrid.ldgo.columbia.edu/>].

b. Methods

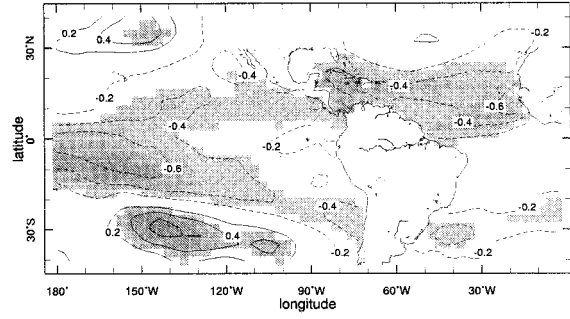
Since the principal rainy season in the Caribbean/Central American region lasts from May to October, and rainfall in the Nordeste region of Brazil is limited to boreal spring, we focus our analysis on the following:

- the summer of year (0), average of July–August–September [JAS(0)], when we expect only the tropical component of the ENSO teleconnection to be at work, manifest in the zonal seesaw in SLP between the Pacific and Atlantic, and in the associated weakening of the trade winds in the tropical North Atlantic;
- the winter of year (+1), average of January–February–March [JFM(+1)], when the PNA-related low SLP center in the western subtropical North Atlantic further weakens the meridional SLP gradient, and strengthens the SST response in the tropical North Atlantic;
- the spring of year (+1), average of April–May–June [AMJ(+1)], when we observe the delayed SST response in the Atlantic, and the local response of the atmosphere to it.

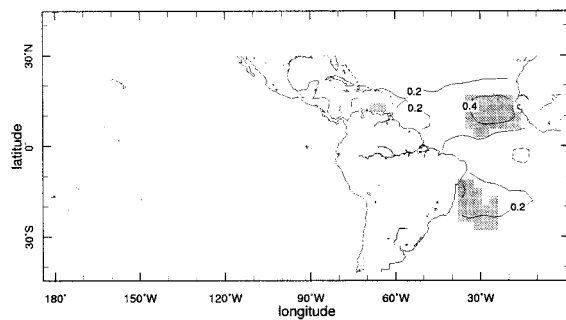
For each variable, we create a time-dependent (GOGA or TAGA) ensemble mean field by averaging together the five ensemble members, for each one of the 540 months in the integrations. We divide seasonal averages for summer, winter, and spring into three categories: warm ENSO events, when the December(0)/January(+1) Niño-3 index is greater than 1°C , cold ENSO events, when the same index is less than -0.75°C , and the remaining neutral years (see Table 1). The threshold for cold events is smaller to reflect the skewness in the distribution of Niño-3 SSTs. Separate consideration of warm and cold events has the advantage over correlation or regression maps in that the response to SST is not assumed to be linear (Hoerling et al. 1997). Also, identification of features common to events of either sign can help separate out features that reflect local climate variability rather than ENSO. To compute the statistical significance of the differences between warm, or cold



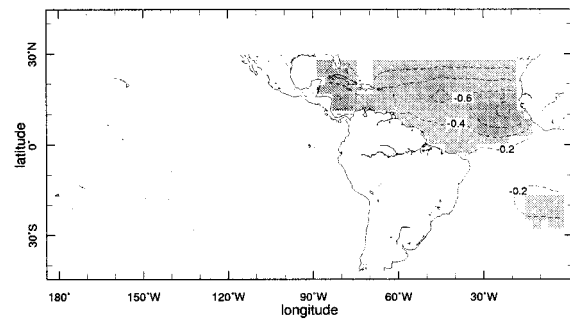
GOGA warm ENSO AMJ(+1) SST



GOGA cold ENSO AMJ(+1) SST



TAGA warm ENSO AMJ(+1) SST



TAGA cold ENSO AMJ(+1) SST

FIG. 3. SST anomalies in GOGA (top row) and TAGA (bottom row) during the spring following warm (left) and cold (right) ENSO events. Contour is 0.2°C. Shading represents statistical significance greater than 95%.

TABLE 1. Characterization of the ENSO events during 1950–94. The columns (column number in parentheses), list from left to right: 1) warm ENSO events; 2) cold ENSO events; 3) the phase of the NAO during the winter of year (0), in parentheses, and during the winter of year (+1); 4) the amplitude of the Niño-3 index during the spring of year (+1) [AMJ (+1)]; 5) the sign of the tropical South Atlantic (tSA) SST index during the same spring of year (+1).

Warm ENSO	Cold ENSO	NAO phase	AMJ(+1) Niño-3	AMJ(+1) tSA
	1949–50		-0.77	-
	1955–56		-0.32	-
1957–58		(+) -	0.42	-
	1964–65		0.70	+
1965–66		(-) -	-0.06	+
	1967–68		-0.37	-
1969–70		(-) -	-0.24	-
	1970–71		-0.57	+
1972–73		(+) +	-0.58	+
	1973–74		0.03	+
	1975–76		0.29	-
1976–77		(+) -	0.13	+
1982–83		(+) +	2.12	-
	1984–85		-0.53	+
1986–87		(+) -	1.40	+
1987–88		(-) +	-0.93	+
	1988–89		-0.24	+
1991–92		(+) +	1.32	-
1994–95		(+)		

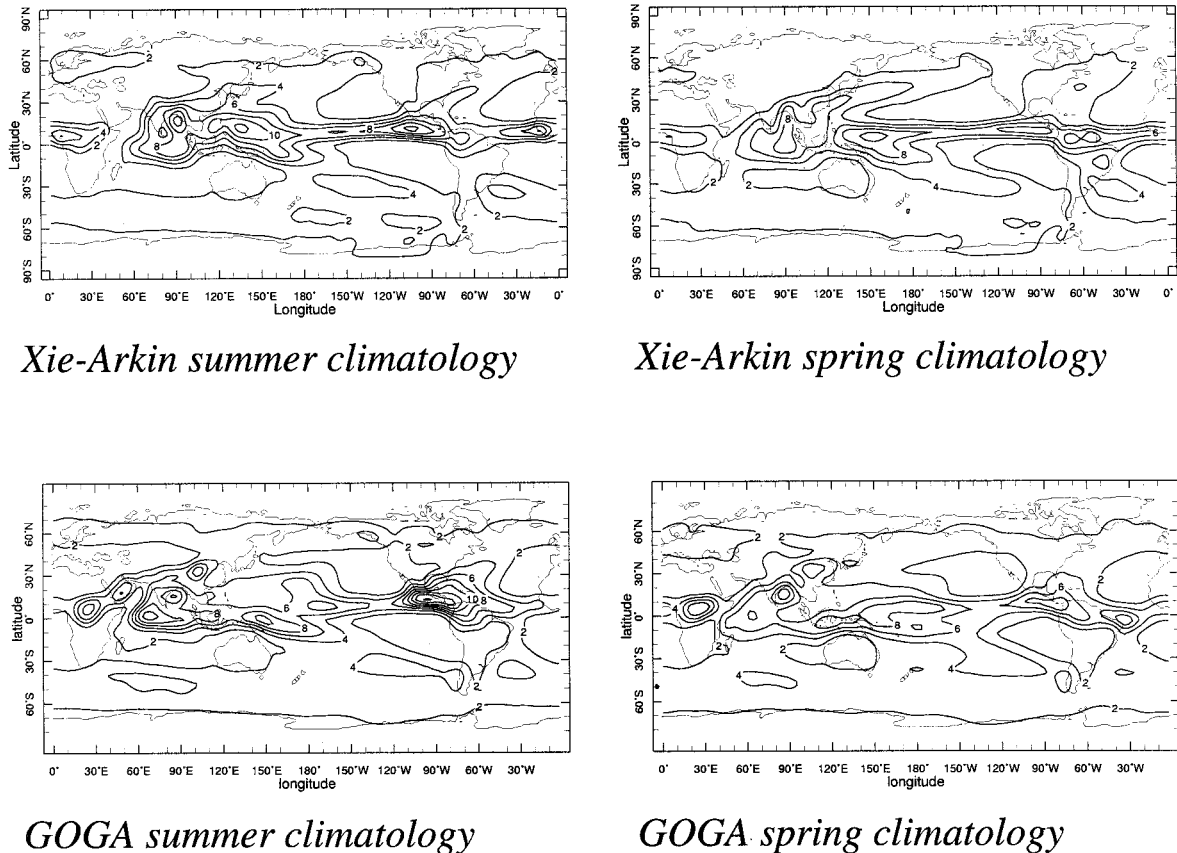


FIG. 4. The seasonal averages of rainfall in observations (Xie–Arkin 1979–94) (top left) summer, (top right) spring; and in the model’s GOGA ensemble mean, averaged over the same period (1979–94) (bottom left) summer, (bottom right) spring. Contour is 2 mm day^{-1} .

ENSO and neutral conditions, we apply Student’s *t*-test (Fisher 1970) to these differences.

3. A comparison of CCM3’s rainfall climatology with observations 1979–94

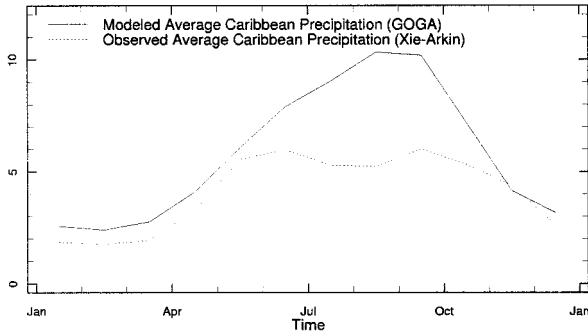
The ability of CCM3 to realistically simulate the annual cycle of precipitation in the Tropics varies greatly by region (Camargo et al. 2001; Fig. 4). The most striking failure is not in our region of interest, the tropical Americas, but over sub-Saharan Africa. The observed, spatially coherent pattern of Northern Hemisphere summer rainfall that extends across the African continent from its Atlantic to its Indian coasts is longitudinally confined in simulations between 10° and 30°E . The region of maximum rainfall remains confined to this longitudinal interval all year long, moving latitudinally following the sun to central Africa during fall, and to southeastern Africa during winter. In contrast, over the tropical Americas both the spatial extent and meridional migration of the rains between the Amazon and Caribbean basins are well captured.

The model has a wet bias over the Caribbean, evident during the May–October rainy season (Fig. 5, left panel). The double-peaked shape of the observed rainy

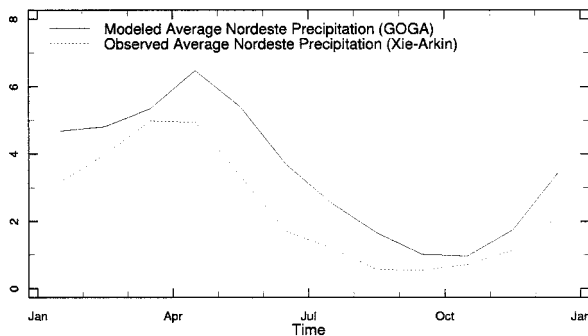
season, with the characteristic summer dry spell between spring and fall peaks (GKC), is not captured. Instead, rainfall in the model peaks in July–August–September. It is difficult to offer a satisfying explanation for this behavior, given that the observed features of the annual cycle of rainfall in the Caribbean/Central American region have not been fully explained to date. The annual cycle of rainfall over the Nordeste region of Brazil (Fig. 5, right panel), is better captured, in pattern and magnitude, though the model shows a slight wet bias in this region too. The rainy season is limited to the three months of March–April–May, when the Atlantic ITCZ reaches its southernmost position in conjunction with the reversal of the climatological meridional gradient of SST between the southern and northern tropical Atlantic.

4. The ENSO teleconnection to the tropical Americas in CCM3

To separate the effects on rainfall of the atmosphere’s direct response to Pacific SSTs from the local response to remotely generated anomalous tropical Atlantic SSTs, we compare the GOGA and TAGA integrations. We expect the GOGA ensemble to reproduce the ENSO



Caribbean annual cycle



Nordeste annual cycle

FIG. 5. Comparison of the observed and modeled (GOGA) annual cycles of rainfall over (left) the Caribbean region (5° – 25° N, 90° – 60° W), and (right) the Nordeste region (0° – 10° S, 45° – 15° W), in the model (solid line) and in observations (dashed line). The average annual cycle was computed over the same period (1979–94). (Units: mm day^{-1} .)

teleconnection in its entirety, and the TAGA ensemble to depict only the local atmospheric response to the tropical Atlantic SSTs that are remotely forced by ENSO. We expect significant differences between the two from the summer of year (0) through the winter of year (+1), that is, for as long as the SST anomalies in the central and eastern equatorial Pacific drive an anomalous atmospheric circulation in GOGA, but not in TAGA. With the decay of the Pacific SST anomalies and the consequent demise of the atmospheric bridge in spring of year (+1), we expect the local SST anomalies to take charge and effect changes in the atmospheric circulation common to the two ensembles.

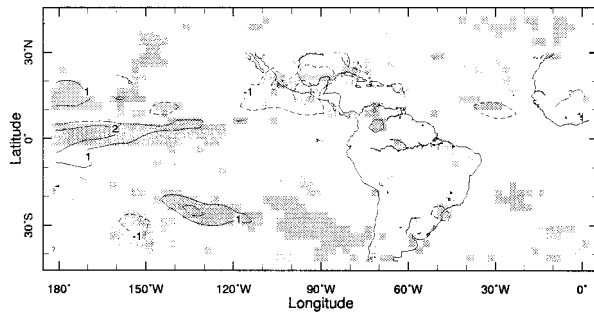
a. The tropical seesaw in sea level pressure

The tropical component of the ENSO teleconnection to the Atlantic basin, the “other lobe” of the Southern Oscillation, is part of the global rearrangement of air mass that follows the anomalous eastward extension of the Pacific warm pool during an El Niño. It is present from onset to decay of an ENSO event, but the summer

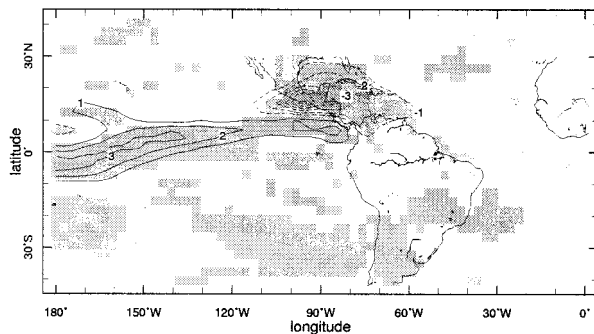
of year (0) is the best season to isolate its effect on rainfall in the tropical Americas, because this is the only active component of the ENSO teleconnection at this time. In the summer hemisphere the propagation of Rossby wave trains to the extratropics is hindered by the poleward extension of easterly winds (Tribbia 1991), hence a PNA-type pattern is absent. Also, the anomalous SSTs in the tropical Atlantic have not had time to build up yet. Because this component of the ENSO teleconnection is purely atmospheric, the comparison between rainfall composites of ENSO events in GOGA and TAGA should show large interensemble differences, and rainfall anomalies should be present only over those regions reached by the atmospheric bridge, only in the GOGA ensemble.

In Fig. 6 the difference in rainfall between warm ENSO and neutral years during JAS(0) is compared in observations (Xie–Arkin during 1979–94, top panel), and in the ensemble mean of GOGA (middle panel) and TAGA (bottom panel). The positive rainfall anomalies present in the tropical Pacific in observations, associated with the eastward shifts of equatorial deep convection and of the South Pacific convergence zone, are reproduced in GOGA. So are the negative rainfall anomalies over the Pacific warm pool (not shown) and over the Caribbean and Central America. The greater statistical significance of the model anomalies is partly due to the larger number of events that make up this composite compared to the observed composite. [The GOGA composite over 1979–94 also exhibits diminished statistical significance (not shown).] It may also be due in part to the fact that we are comparing one single realization in the case of observations with the average of five realizations in the case of the model integrations, where computation of the ensemble mean results in reduction of internal variability (F. Zwiers 2001, personal communication). Anomalies reverse in sign during cold ENSO events (not shown), and are weaker. Cold ENSO SST anomalies in the Pacific are weaker than warm anomalies, and the heating response to anomalous SSTs is nonlinear (Hoerling et al. 1997).

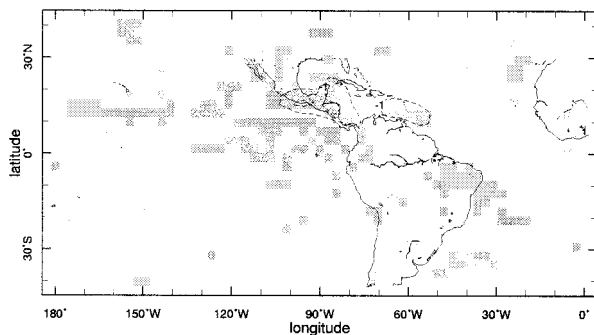
Contrary to expectations, an area of dry anomalies covering Central America is still present in the TAGA ensemble (Fig. 6, bottom panel), though not nearly as large or extensive as its GOGA counterpart. To account for this anomaly we call on the other known player in Caribbean rainfall variability, the North Atlantic oscillation (NAO). The NAO is hypothesized to affect springtime Caribbean rainfall indirectly, through the persistence of SST anomalies generated in winter (GKC; GCK). An anomalously positive wintertime NAO translates into anomalously strong trade winds, hence anomalous ocean–atmosphere heat fluxes and cooling of SSTs in the tropical North Atlantic (Cayan 1992; Seager et al. 2000). Extension of the NAO’s influence on North Atlantic SST anomalies and on Caribbean climate into the following summer has emerged from several recent studies. It is hinted by the recent analysis of Czaja and



OBS warm ENSO JAS(0) PRCP



GOGA warm ENSO JAS(0) PRCP



TAGA warm ENSO JAS(0) PRCP

FIG. 6. The rainfall response to warm ENSO conditions in JAS(0), in observations (top) (1979–94), and in the GOGA (middle) and TAGA (bottom) integrations (1950–94). Contour is 1 mm day^{-1} . Shading represents statistical significance greater than 95%.

Frankignoul (1999) and is briefly discussed in Robinson (2000, diagram 1 attributed to Kushnir on the persistence of SST anomalies generated by the winter NAO). It has been suggested that in the tropical Atlantic during spring and summer a positive feedback in the air–sea interaction occurs over the SST anomalies generated in winter in which the amount of low-level cloud-

iness changes to counter their damping via turbulent surface flux, and thus enhances their persistence. The SSTs used to force the model integrations bear the imprint of the observed NAO. When the 10 warm ENSO events in TAGA are classified according to the sign of the observed NAO (Hurrell 1995; available online at http://www.cgd.ucar.edu/cas/climind/nao_winter.html), 7 out of the 10 warm ENSO JAS(0) turn out to have followed winters characterized by a positive NAO index (Table 1). Comparison between positive and negative NAO years shows that the TAGA pattern of anomalously low rainfall during warm ENSO events may be understood in terms of the negative SST anomalies in the tropical North Atlantic forced during the previous winter and spring by an NAO of predominantly positive sign (not shown).

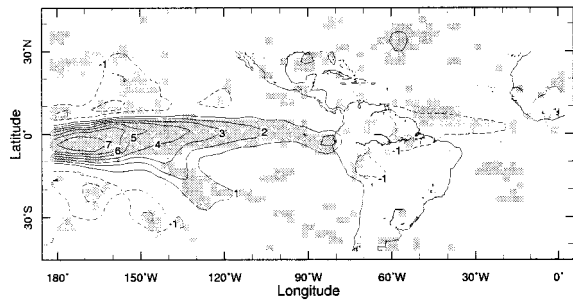
b. The extratropical component of the atmospheric bridge

The full impact of the ENSO-related anomalous global atmospheric circulation on Caribbean/Central American rainfall, untainted by the impact of the delayed Atlantic SST anomalies, is best observed in warm events in winter of year (+1) (Fig. 7). Since the delayed warming of the subtropical North Atlantic is only starting to build up during this season, rainfall anomalies can still be attributed primarily to the workings of the atmospheric teleconnection. Once again, we expect significant differences between the ENSO composites in GOGA and TAGA.

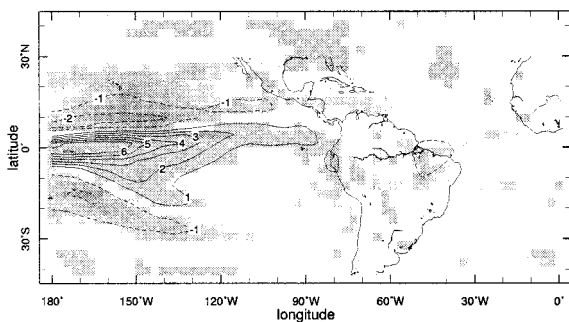
During the mature phase of warm ENSO events, rainfall anomalies in observations and in GOGA (Fig. 7) are positive all across the equatorial Pacific, from Papua New Guinea to the South American coast, and negative to the immediate north and south. The anomaly north of the equator extends eastward into the Caribbean basin, where surface divergence continues to prevail. North of the Caribbean, over the Gulf of Mexico and southeastern United States, is an anomaly of opposite sign obviously associated with the PNA-related low SLP center in the western subtropical Atlantic. In a warm ENSO event the subtropical jet is displaced southward of its climatological location, funneling disturbances into the Gulf of Mexico. No trace of these rainfall anomalies is left in the TAGA ensemble (Fig. 7, bottom panel), indicating that they are purely atmospheric driven. Remotely forced positive SST anomalies make their appearance during this season in the tropical North Atlantic, but they are too weak to influence rainfall even in the absence of the atmospheric bridge, in TAGA.

c. The remotely forced Atlantic SST response

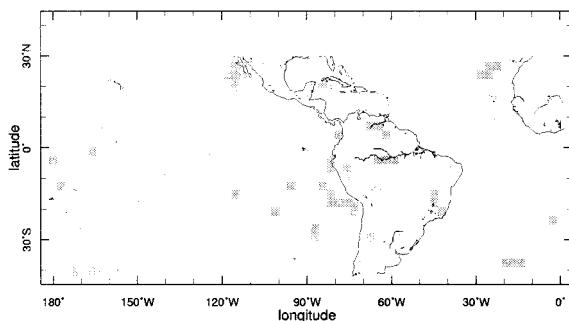
In spring of year (+1), there is potential for the ENSO-related atmospheric bridge and remote SST response to interact. This season usually marks the transition to normal atmospheric conditions. At the same



OBS warm ENSO JFM(+1) PRCP



GOGA warm ENSO JFM(+1) PRCP



TAGA warm ENSO JFM(+1) PRCP

FIG. 7. The rainfall response to warm ENSO conditions in JFM(+1), in observations (top) (1979–94), and in the GOGA (middle) and TAGA (bottom) integrations (1950–94). Contour is 1 mm day⁻¹. Shading represents statistical significance greater than 95%.

time the delayed Atlantic SST response becomes significant. If rainfall were affected by SST variability local to the Atlantic only, there would be no significant differences between GOGA and TAGA, and the role of Atlantic climate variability independent of ENSO would become more visible.

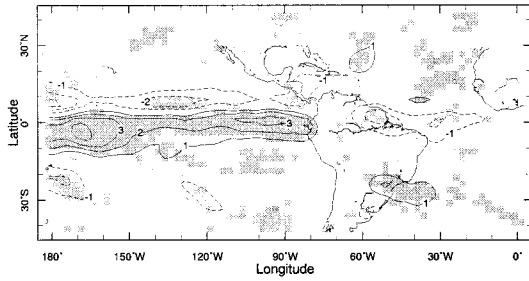
Warm and cold ENSO events in observations (Fig. 8, top panels) display similarities in the tropical Pacific, with rainfall anomalies of one sign at the equator, and

of opposite sign poleward in both hemispheres. They display differences in the tropical Atlantic: rainfall anomalies are symmetric about the equator in warm events, antisymmetric in cold events. We speculate that these differences can be explained by the duration of ENSO events. In three of the four events that make up the observed warm ENSO composite (1982–83, 1986–87, and 1991–92; Fig. 8, top left) Niño-3 anomalies are still above 1°C in spring (+1) (see Table 1, column 4). ENSO has not decayed yet, and is capable of driving Tropics-wide atmospheric anomalies. We will come back to the “unusual” character of the events of the last 20 years, which dominate the period spanned by the Xie–Arkin dataset analyzed here. In the two events that make up the cold ENSO composite (1984–85 and 1988–89; Fig. 8, top right); on the other hand, Niño-3 anomalies are already well into the decaying phase in spring (+1). The rainfall response in cold ENSO events can be explained in terms of local SST anomalies. The cooling of the tropical North Atlantic is accompanied in both cold events of the last 20 yr by a warming of the tropical South Atlantic (see Table 1, column 5), and the dipole in rainfall reflects the SST dipole.

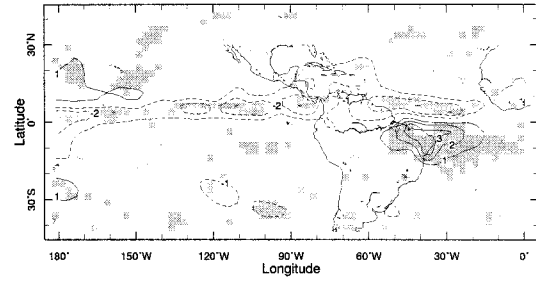
Anomalies in the GOGA ensemble mean (Fig. 8, middle panels) reproduce the spatial distribution of observed anomalies, but are overall weaker, especially in the central and eastern tropical Pacific. As hypothesized earlier, this is likely due to the differences in duration of the events. In all ENSO events prior to 1979 except the cold ENSO event of 1949–50 tropical Pacific SST anomalies were well into the decay phase by spring (+1).

Meaningful differences appear between GOGA and TAGA composites in the Caribbean and over northeast Brazil (Fig. 8, cf. middle and bottom panels). GOGA–TAGA differences stand out especially in cold ENSO events. North of the equator, rainfall anomalies in the tropical North Atlantic extend westward into the Caribbean basin only in TAGA (Fig. 8, bottom right). They exemplify the local thermodynamic nature of the rainfall response to the underlying SSTs. Negative rainfall anomalies are consistent with the underlying cold SST anomalies. The absence of Caribbean rainfall anomalies in GOGA, in warm and cold ENSO events alike (Fig. 8, middle panels), is an indication of the simultaneous presence of competing features of the ENSO teleconnection to the Atlantic. The atmospheric anomalies (surface divergence or convergence in the Caribbean basin in warm or cold events, respectively) cannot be identified because the oceanic anomalies (warm or cold tropical North Atlantic SSTs in warm or cold events, respectively) cancel their impact on rainfall.

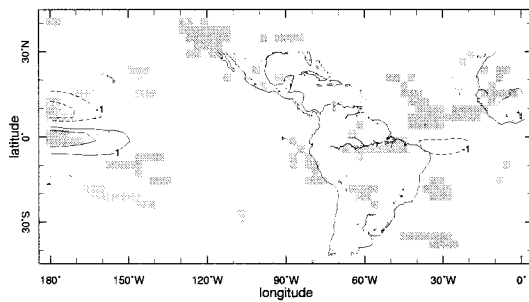
At the equator, still in cold ENSO events only, a conspicuous positive rainfall anomaly covers the Nordeste and adjacent western Atlantic in GOGA, but not in TAGA. We interpret it primarily as the effect of an anomalous atmospheric circulation remotely driven from the Pacific. When ENSO-related atmospheric



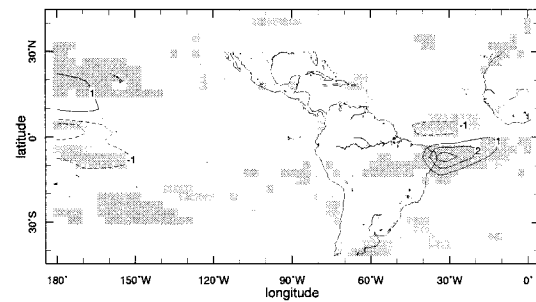
OBS warm ENSO AMJ(+1) PRCP



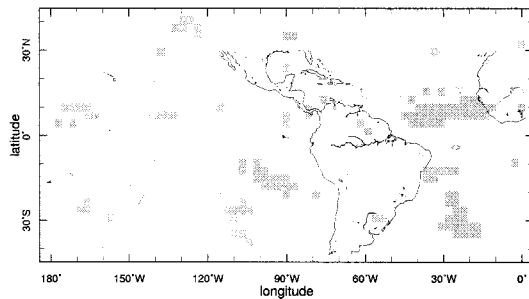
OBS cold ENSO AMJ(+1) PRCP



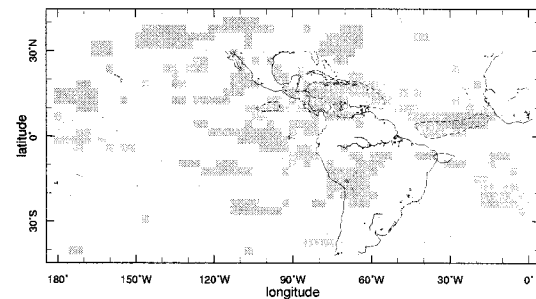
GOGA warm ENSO AMJ(+1) PRCP



GOGA cold ENSO AMJ(+1) PRCP



TAGA warm ENSO AMJ(+1) PRCP



TAGA cold ENSO AMJ(+1) PRCP

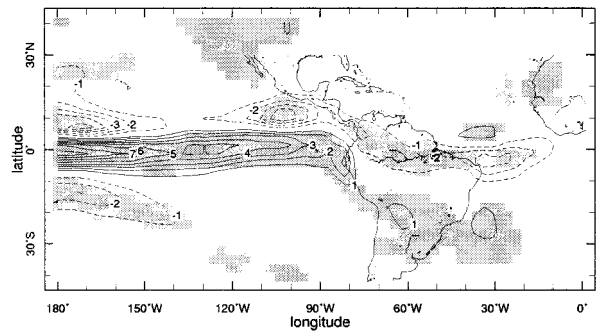
FIG. 8. The rainfall response to ENSO in AMJ(+1), in observations (1979–94), and in the GOGA and TAGA integrations (1950–94). (left column) warm ENSO events. (top left) observations, (middle left) GOGA, and (bottom left) TAGA. (right column) cold ENSO events (top right) observations, (middle right) GOGA, and (bottom right) TAGA. Contour is 1 mm day⁻¹. Shading represents statistical significance greater than 95%.

anomalies are still present in spring (+1) of a cold ENSO event, they translate into anomalous convergence at the surface and anomalous divergence aloft over the northeastern corner of South America (not shown). The ENSO-related cooling of the tropical North Atlantic plays a secondary role, but by favoring a southward

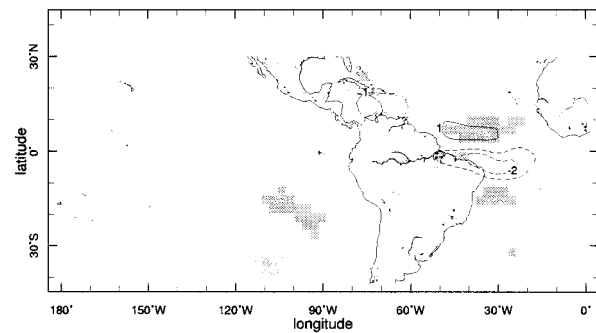
displacement of the Atlantic ITCZ, it adds on to the atmosphere's role in forcing Nordeste rainfall anomalies. The superposition of atmospheric and oceanic conditions favorable to Nordeste rainfall explains the significantly stronger rainfall anomalies just south of the equator found in the GOGA cold ENSO composite. As

mentioned earlier, there are discrepancies in the boundary conditions between the GOGA and TAGA ensembles, of the order of 0.1°C and most prominent at the equator (Fig. 3). In cold events these anomalies are consistent in sign with the GOGA–TAGA difference in the precipitation response. Slightly cooler SSTs just north of the equator in TAGA explain the modest decrease in rainfall with respect to GOGA local to these SSTs. The stronger positive rainfall response to the south, over the Nordeste, that is seen in GOGA compared with TAGA could be partially due to the stronger southward meridional gradient in SST, but that is unlikely totally so. If one considers an SST difference not greater than 0.1°C between GOGA and TAGA, at best marginally significant in dynamic terms, one should be convinced that this SST discrepancy cannot be anointed as the only explanation for a rainfall difference of 2 mm day^{-1} , which is a significant fraction of the seasonal climatology.

To confirm the dominant role of the atmospheric bridge when still active during spring (+1), as in the cold ENSO events described above, we analyze warm ENSO events more closely. Following Chiang et al. (2000b) we note that the majority (4 out of 6) of warm ENSO events during the most recent 20 yr (1979–99) have been characterized by the protraction of the warm Pacific anomaly well into this season. The rainfall response to persistent Pacific SST anomalies is evident in the composite of warm ENSO events in observations (Fig. 8, top left). We define “long” as those events for which the value of Niño-3 is still above 1°C in AMJ(+1) (see Table 1). Three out of these four events occurred during 1950–94 (the fourth is the 1997–98 event, not included in the simulation period). The remaining six events during 1950–94 are referred to as “canonical” (Rasmusson and Carpenter 1982). During a warm ENSO event the entire tropical free troposphere warms (Yulaeva and Wallace 1994) and vertical stability is increased everywhere outside of the region of deep convection in the tropical Pacific. The atmospheric teleconnection interferes constructively with the northward displacement of the ITCZ associated with the warming of the tropical North Atlantic (Lindzen and Nigam 1987), so that the negative Nordeste rainfall response is enhanced when both effects are present. The impact of the prolonged presence of the Pacific SST anomaly on the global atmospheric circulation and on rainfall in the tropical Americas is clearly seen in the comparison between the average of long events only and all events in GOGA (cf. Fig. 9, left panel, with Fig. 8, middle-left panel). Warm Pacific SST anomalies in the long case are consistent with the persistence of the SLP seesaw between the Atlantic and Pacific basins, of wet anomalies all along the equator in the Pacific, and of dry anomalies in the subtropical Pacific, the Caribbean, northern South America, and the equatorial Atlantic. However, they are also associated with cooler SSTs in the equatorial Atlantic (not shown), hence an increased



GOGA long warm ENSO AMJ(+1)



TAGA long warm ENSO AMJ(+1)

Fig. 9. The rainfall response in AMJ(+1) during long warm ENSO events: (left) GOGA and (right) TAGA. Contour is 1 mm day^{-1} . Shading represents statistical significance greater than 95%.

northward SST gradient. TAGA rainfall anomalies over the Nordeste in the long-event case (Fig. 9, right panel) are weaker than in GOGA, but stronger than in the composite of all warm events (cf. the latter with Fig. 8, bottom-left panel), also consistent with the appearance of this cold SST anomaly in the equatorial Atlantic.

Rainfall anomalies in the equatorial Atlantic during boreal spring can be partially explained by at least three mechanisms: anomalies in vertical stability associated with the Tropics-wide warming of the free troposphere, the meridional displacement of the ITCZ largely driven by the ENSO-related anomalous SST in the tropical North Atlantic, and SST anomalies generated by air–sea interaction local to the equatorial Atlantic (Chang et al. 1997). In the latter, it is hypothesized that the cross-equatorial response of the winds to SST anomalies of opposite sign on opposite sides of the equator reinforces the SST anomalies themselves, until wind-driven ocean heat transports cancel out the initial SST anomalies. It is not possible to separate the roles of the atmospheric bridge and of the local SST–surface wind feedback, since both features are present in spring (+1) only when the ENSO event is still ongoing during this season. The observed result is that they add up to the effect of the tropical North Atlantic warming, a more

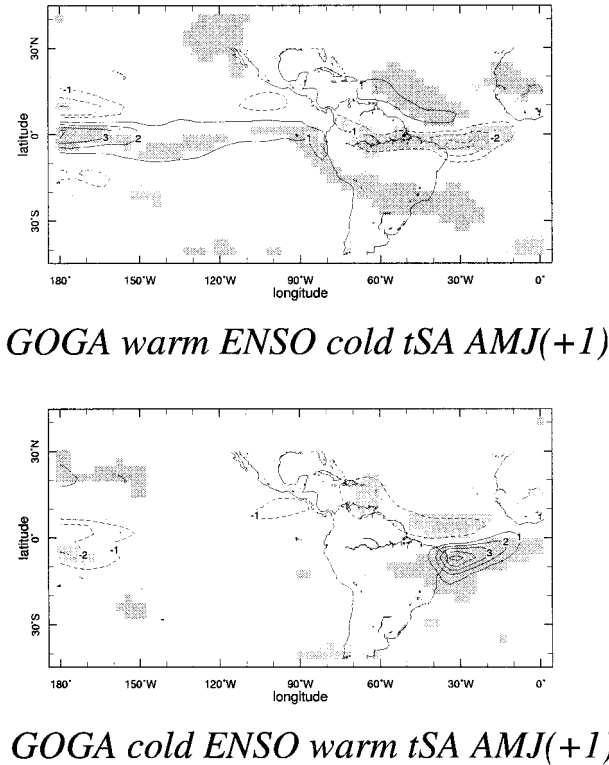


FIG. 10. The GOGA AMJ(+1) rainfall response to ENSO and to tropical South Atlantic SST: (left) warm ENSO and negative tSA and (right) cold ENSO and positive tSA. Contour is 1 mm day⁻¹. Shading represents statistical significance greater than 95%.

northern location of the ITCZ, with nefarious consequences for the Nordeste rainy season. Ideally, we would like to be able to compare the GOGA and TAGA ensembles with a “POGA” ensemble, that is, an ensemble forced with observed SST in the Pacific Ocean only, monthly climatology elsewhere. This third ensemble would produce only the remotely driven variability in the tropical Atlantic atmosphere without the two mechanisms caused by SST changes in the Atlantic. Alternatively, an ensemble forced with observed SSTs in the Pacific only, coupled to a simple mixed layer ocean in the tropical Atlantic would allow to evaluate the role of ocean–atmosphere feedbacks.

To balance our ENSO-centric findings we explicitly compare the role of the Pacific-driven mechanisms, oceanic or atmospheric, with that due to the SST variability

local to the Atlantic (Fig. 10). We separate ENSO events in GOGA according to the sign of “tSA,” a tropical South Atlantic SST index defined as the SST averaged between 5° and 25°S, 40°W and 20°E (see Table 1). The rainfall response in the equatorial Atlantic is significant only when the influences of ENSO-related and local forcings are concordant in sign. For cold ENSO events, in the five cases out of nine when tSA is positive (Fig. 10, right panel), the cooler than average SSTs and negative rainfall anomalies go together in the north, and the warmer than average SSTs and positive rainfall anomalies go together to the south of the equator. For warm ENSO events, in the four cases out of nine when tSA is negative (Fig. 10, left panel) (two are the long events of 1982–83 and 1991–92), rainfall anomalies are positive in the Caribbean, negative over northeast Brazil. These findings are consistent with the vast literature on the role of the meridional gradient in SST on Nordeste rainfall (e.g., Hastenrath and Greischar 1993).

5. Discussion and conclusions

In this study we detailed the seasonally dependent evolution of the ENSO-related interannual variability of rainfall in the Caribbean/Central American region, where the rainy season coincides with the boreal warm season, and over the northeastern corner of Brazil, where the rainy season is limited to boreal spring. We followed the development of an ENSO life cycle from shortly after its onset, in summer of year (0), to its decay almost a year later. To assess ENSO’s impact on the variability of rainfall in the tropical Americas (see Table 2), we compared two ensembles of integrations using CCM3, one (GOGA) forced with globally observed SST, the other one (TAGA) with climatological, monthly varying SST everywhere but in the tropical Atlantic basin, where observed boundary conditions were applied. The GOGA ensemble reproduced the full ENSO teleconnection, in its atmospheric and oceanic components, as expected, whereas the influence of ENSO in the TAGA ensemble was limited to the ENSO-related response in Atlantic SST, also as expected.

In the Caribbean/Central American region, the GOGA ensemble reproduces the observed features of the impact of the atmospheric bridge. The direct impact of the anomalous atmospheric circulation associated with anomalies in the center of deep convection in the equatorial Pacific dominates the rainfall response from soon

TABLE 2. Summary of the seasonal dependence of the rainfall response to a warm ENSO event in the Caribbean/Central American region and in the Nordeste region of Brazil.

	Mechanism	Caribbean/Central America	Northeast Brazil
Summer (0)	Atmospheric bridge	Dry	n/a
Winter (+1)	Atmospheric bridge	Dry	n/a
Spring (+1)	Atmospheric bridge	Dry	Dry
	Delayed North Atlantic SST response	Wet	Dry

after the onset of an ENSO event, during the summer of year (0), through its mature phase, during the winter of year (+1). During a warm ENSO event, the anomalous atmospheric circulation drives surface divergence away from the Caribbean basin, with the flow directed along a southwest–northeast axis. Rainfall is below average during the period leading to mature ENSO conditions. Anomalies of opposite sign characterize cold ENSO events. This effect is lost in the TAGA ensemble, due to the absence of the interannual variability of the Pacific source. TAGA, on the other hand, does capture the impact of the wintertime NAO on Caribbean rainfall via tropical North Atlantic SST anomalies, which last through spring and into summer. A rainfall pattern very similar to the ENSO-related one found in GOGA is associated in TAGA with the more frequent occurrence of a positive NAO in observations during the winter preceding the onset of warm ENSO events.

The atmospheric component of the teleconnection is also responsible for the development of SST anomalies in the tropical North Atlantic, of the same sign as in the eastern equatorial Pacific, but delayed by about a season. As the ENSO event winds down, rainfall anomalies in the Caribbean reverse in sign in response to these anomalous Atlantic SSTs. They have greater spatial extent in the TAGA ensemble, where they cover the SST anomaly from West Africa to Central America. They are not well defined in the Caribbean basin proper in the GOGA ensemble, presumably because of the competition between the atmospheric and oceanic components of the ENSO teleconnection in this region during this season. In the case of a warm ENSO event this means that the surface divergence associated with the atmospheric bridge tends to be canceled by the positive effect of warm SST on rainfall.

In the equatorial Atlantic/Brazilian Nordeste region, ENSO's impact is limited to the local rainy season between the end of (boreal) winter and the beginning of spring of year (+1). In this season the climatological SST gradient is southward, allowing for the brief displacement of the Atlantic ITCZ to the south of the equator. ENSO's impact varies considerably depending on the length of the event. Since in a "canonical" event the Pacific SST anomalies and related atmospheric bridge are usually well into the decay phase by the spring of year (+1), SSTs local to the Atlantic, be they remotely forced or independent of ENSO, determine the outcome of the rainy season in the Brazilian Nordeste. In the absence of remote atmospheric anomalies, negative rainfall anomalies are associated with the ENSO-related warming of the tropical North Atlantic. The northward SST gradient favors an anomalously northern location of the ITCZ. Conversely, a positive SST anomaly in the South Atlantic in conjunction with the cooling of tropical North Atlantic waters associated with a cold ENSO event, can redirect the SST gradient toward the south, favoring above-average rainfall over the Nordeste.

Nordeste rainfall anomalies stand out when the ENSO event lasts through the local rainy season, as has happened with the recent events of 1982–83, 1986–87, and 1991–92 (Chiang et al. 2000b). In this case, though this study cannot separate between the roles of the remotely forced anomalous circulation over the equatorial Atlantic and of the feedback of equatorial Atlantic ocean–atmosphere interaction onto the tropical North Atlantic SSTs, it is clear that both mechanisms contribute, adding up to the impact of the remotely forced anomalous SSTs in the tropical North Atlantic. In a warm ENSO event, the tendency for the ITCZ to stay north of the equator, related to the delayed warming of the tropical North Atlantic, is complemented by anomalous subsidence over the Nordeste, and by a northward cross-equatorial flow at the surface. Rainfall is scarce. In a cold ENSO event, the tendency for the ITCZ to migrate south, away from the cold tropical North Atlantic waters, is complemented by anomalously low SLP, and a southward cross-equatorial surface flow. Rainfall is copious, occasionally resulting in widespread flooding.

In this paper we focused on the features of tropical Atlantic variability associated with ENSO. We identified two mechanisms by which climate variability of tropical Pacific source can have a remote impact on rainfall in the tropical Americas: the atmospheric bridge and delayed SST response in the tropical North Atlantic. It is reassuring to note that a parallel study by Chiang et al. (2000b), by taking the complementary stance of deconstructing Atlantic ITCZ variability, identifies two mechanisms that bear a strong resemblance to those found here. The close relationship between the workings of the atmospheric bridge here presented and the Walker mechanism of Chiang et al. is self-evident. The fact that the ENSO-related atmospheric bridge and delayed SST response in the Atlantic can add up in their impact on the Atlantic ITCZ and on the Nordeste rainy season is borne out in the language of Chiang et al. in the observation that features of the gradient mechanism appear in the Walker composites (see their Fig. 6a). In this sense, even though ENSO is not the only source of tropical North Atlantic SST variability, its impact as discussed here is consistent with the meridional SST gradient hypothesis.

ENSO's impact on rainfall over the tropical Americas is ultimately related to the tendency for a globally warmer than average tropical climate during warm ENSO events, manifest in warmer than average free troposphere (Yulaeva and Wallace 1994) and remote ocean basins (Klein et al. 1999). The warming of the free troposphere translates into a more stable vertical profile, hence predominantly dry conditions away from the Pacific center of deep convection. The boundary layer response to the atmospheric bridge can locally contrast or enhance the Tropics-wide tropospheric anomalies. As detailed above, rainfall anomalies in the Nordeste region of Brazil are enhanced when both responses are present. Conversely, when the outcome depends on the com-

petition between forcings of opposite sign, from above in the free troposphere, and from below at the boundary layer, as is the case over equatorial East Africa (Goddard and Graham 1999) and in the Caribbean basin during the spring of year (+1), rainfall anomalies are less coherent interannually.

In concluding this study, we briefly consider implications for the predictability of rainfall in the tropical Americas.

- In the Caribbean/Central American region, knowledge of the state of the North Atlantic oscillation during the winter preceding the onset of ENSO, coupled with the prediction of ENSO onset can contribute to an increased predictability of rainfall during the warm season of year (0). Knowledge of mature phase ENSO conditions and of the prevailing state of the NAO during the same winter can be exploited to predict spring rainfall in year (+1).
- In the Brazilian Nordeste, knowledge that an ENSO event is ongoing during the months leading to the local rainy season is not always sufficient. Prediction of the duration of the event beyond the mature phase is critical, together with an understanding of tropical Atlantic SST variability, in particular what controls the cross-equatorial SST gradient.

The study also points to features of the climate system that need to be better understood in order to improve prediction of rainfall in the tropical Americas.

- What are the dynamics of the propagation of the ENSO signal to the tropical Atlantic, on timescales faster than a month?
- How does ENSO's impact on equatorial Atlantic SSTs relate to the dynamical feedback mechanism hypothesized by Chang et al. (1997)? Can a "long" ENSO event impart a strong enough perturbation to the system to initiate the feedback response?
- What causes the internal variability of ENSO, which manifests itself in the varying length of events?
- What would be the response to perpetual warm ENSO conditions in the tropical Pacific, which are predicted by some coupled models when integrated in doubled CO₂ experiments? Would the atmospheric response overwhelm the delayed SST response in the Atlantic, causing a negative rainfall trend both in the Caribbean and over the Nordeste?

Acknowledgments. This work was supported by NOAA (Grant NA86GP0515), by the National Science Foundation (Grant ATM-99-86072), partially supported by NASA InterAgency Agreement W-19, 750, and by NOAA Grant NA06GP0480. Special thanks to Ping Chang and R. Saravanan, who shared the output of their CCM3 integrations. Many thanks to the editor, Francis Zwiers, and referees, whose comments greatly helped in improving the original manuscript.

REFERENCES

- Aceituno, P., 1988: On the functioning of the Southern Oscillation in the South American sector. Part I: Surface climate. *Mon. Wea. Rev.*, **116**, 505–524.
- Camargo, S. J., S. E. Zebiak, D. G. DeWitt, and L. Goddard, 2001: Seasonal comparison of the response of CCM3.6, ECHAM4.5 and COLA2.0 atmospheric models to observed SSTs. IRI Tech. Rep. 01-01, 68 pp. [Available from International Research Institute for Climate Prediction, 61 Rt. 9W, P.O. Box 1000, Palisades, NY 10964-8000.]
- Cayan, D. R., 1992: Latent and sensible heat flux anomalies over the northern oceans: The connection to monthly atmospheric circulation. *J. Climate*, **5**, 354–369.
- Chang, P., J. Link, and H. Li, 1997: A decadal climate variation in the tropical Atlantic Ocean from thermodynamic air-sea interactions. *Nature*, **385**, 516–518.
- , R. Saravanan, L. Ji, and G. C. Hegerl, 2000: The effect of local sea surface temperatures on atmospheric circulation over the tropical Atlantic sector. *J. Climate*, **13**, 2195–2216.
- Chen, A., A. Roy, J. McTavish, M. Taylor, and L. Marx, 1997: Using SST anomalies to predict flood and drought conditions for the Caribbean. COLA Rep. 49, 24 pp. [Available from Center for Ocean–Land–Atmosphere Studies, 4041 Powder Mill Road, Suite 302, Calverton, MD 20705-3106.]
- Chiang, J. C. H., Y. Kushnir, and A. Giannini, 2000a: Deconstructing Atlantic ITCZ variability: Influence of the local cross-equatorial SST gradient, and remote forcing from the eastern equatorial Pacific. *J. Geophys. Res.*, in press.
- , —, and S. E. Zebiak, 2000b: Interdecadal changes in eastern Pacific ITCZ variability and its influence on the Atlantic ITCZ. *Geophys. Res. Lett.*, **27**, 3687–3690.
- Curtis, S., and S. Hastenrath, 1995: Forcing of anomalous sea surface temperature evolution in the tropical Atlantic during Pacific warm events. *J. Geophys. Res.*, **100**, 15 835–15 847.
- Czaja, A., and C. Frankignoul, 1999: Influence of the North Atlantic SST on the atmospheric circulation. *Geophys. Res. Lett.*, **26**, 2969–2972.
- Enfield, D. B., 1996: Relationship of inter-American rainfall to tropical Atlantic and Pacific SST variability. *Geophys. Res. Lett.*, **23**, 3305–3308.
- , and D. A. Mayer, 1997: Tropical Atlantic sea surface temperature variability and its relation to El Niño–Southern Oscillation. *J. Geophys. Res.*, **102**, 929–945.
- , and E. J. Alfaro, 1999: The dependence of Caribbean rainfall on the interaction of the tropical Atlantic and Pacific Oceans. *J. Climate*, **12**, 2093–2103.
- Fisher, S. R. A., 1970: *Statistical Methods for Research Workers*. 14th ed. Hafner Publishing, 362 pp.
- Giannini, A., Y. Kushnir, and M. A. Cane, 2000: Interannual variability of Caribbean rainfall, ENSO, and the Atlantic Ocean. *J. Climate*, **13**, 297–311.
- , M. A. Cane, and Y. Kushnir, 2001: Interdecadal changes in the ENSO teleconnection to the Caribbean region and the North Atlantic oscillation. *J. Climate*, **14**, 2867–2879.
- Glantz, M. H., R. W. Katz, and N. Nicholls, 1991: *Teleconnections Linking Worldwide Climate Anomalies*. Cambridge University Press, 535 pp.
- Goddard, L., and N. E. Graham, 1999: Importance of the Indian Ocean for simulating rainfall anomalies over eastern and southern Africa. *J. Geophys. Res.*, **104**, 19 099–19 116.
- Hameed, S., K. R. Sperber, and A. Meinstner, 1993: Teleconnections of the Southern Oscillation in the tropical Atlantic sector in the OSU coupled upper ocean–atmosphere GCM. *J. Climate*, **6**, 487–498.
- Hastenrath, S., and L. Heller, 1977: Dynamics of climatic hazards in northeast Brazil. *Quart. J. Roy. Meteor. Soc.*, **103**, 77–92.
- , and L. Greischar, 1993: Circulation mechanisms related to northeast Brazil rainfall anomalies. *J. Geophys. Res.*, **98**, 5093–5102.

- , L. C. de Castro, and P. Aceituno, 1987: The Southern Oscillation in the tropical Atlantic sector. *Beitr. Phys. Atmos.*, **60**, 447–463.
- Hoerling, M. P., A. Kumar, and M. Zhong, 1997: El Niño, La Niña, and the nonlinearity of their teleconnections. *J. Climate*, **10**, 1769–1786.
- Horel, J. D., and J. M. Wallace, 1981: Planetary scale atmospheric phenomena associated with the Southern Oscillation. *Mon. Wea. Rev.*, **109**, 813–829.
- Hurrell, J. W., 1995: Decadal trends in the North Atlantic Oscillation regional temperatures and precipitation. *Science*, **269**, 676–679.
- Janicot, S., A. Harzallah, B. Fontaine, and V. Moron, 1998: West African monsoon dynamics and eastern equatorial Atlantic and Pacific SST anomalies (1970–88). *J. Climate*, **11**, 1874–1882.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471.
- Kiehl, J. T., J. J. Hack, G. B. Bonan, B. P. Boville, D. L. Williamson, and P. J. Rasch, 1998: The National Center for Atmospheric Research Community Climate Model: CCM3. *J. Climate*, **11**, 1131–1149.
- Kiladis, G. N., and H. F. Diaz, 1989: Global climatic anomalies associated with extremes in the Southern Oscillation. *J. Climate*, **2**, 1069–1090.
- Klein, S. A., B. Soden, and N.-C. Lau, 1999: Remote sea surface temperature variations during ENSO: Evidence for a tropical atmospheric bridge. *J. Climate*, **12**, 917–932.
- Lau, N.-C., 1997: Interactions between global SST anomalies and the midlatitude atmospheric circulation. *Bull. Amer. Meteor. Soc.*, **78**, 21–33.
- , and M. J. Nath, 1994: A modeling study of the relative roles of tropical and extratropical SST anomalies in the variability of the global atmosphere–ocean system. *J. Climate*, **7**, 1184–1207.
- , and —, 1996: The role of the atmospheric bridge in linking tropical Pacific ENSO events to extratropical SST anomalies. *J. Climate*, **9**, 2036–2057.
- Lindzen, R. S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the Tropics. *J. Atmos. Sci.*, **44**, 2418–2436.
- Nobre, P., and J. Shukla, 1996: Variations of sea surface temperature, wind stress, and rainfall over the tropical Atlantic and South America. *J. Climate*, **9**, 2464–2479.
- Rasmusson, E. M., and T. H. Carpenter, 1982: Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño. *Mon. Wea. Rev.*, **110**, 354–384.
- Robinson, W. A., 2000: Review of WETS—The workshop of extratropical SST anomalies. *Bull. Amer. Meteor. Soc.*, **81**, 567–577.
- Ropelewski, C. F., and M. S. Halpert, 1987: Global and regional precipitation patterns associated with the El Niño/Southern Oscillation. *Mon. Wea. Rev.*, **115**, 1606–1626.
- , and —, 1989: Precipitation patterns associated with the high index phase of the Southern Oscillation. *J. Climate*, **2**, 268–284.
- , and —, 1996: Quantifying Southern Oscillation–precipitation relationships. *J. Climate*, **9**, 1043–1059.
- Saravanan, R., and P. Chang, 2000: Interaction between tropical Atlantic variability and El Niño–Southern Oscillation. *J. Climate*, **13**, 2177–2194.
- Seager, R., Y. Kushnir, M. Visbeck, N. Naik, J. Miller, G. Krahnmann, and H. Cullen, 2000: Causes of Atlantic Ocean climate variability between 1958 and 1998. *J. Climate*, **13**, 2845–2862.
- Smith, T. M., R. W. Reynolds, R. E. Livezey, and D. C. Stokes, 1996: Reconstruction of historical sea surface temperatures using empirical orthogonal functions. *J. Climate*, **9**, 1403–1420.
- Sutton, R. T., S. P. Jewson, and D. P. Rowell, 2000: The elements of climate variability in the tropical Atlantic region. *J. Climate*, **13**, 3261–3284.
- Trenberth, K. E., G. W. Branstator, D. Karoly, A. Kumar, N.-C. Lau, and C. F. Ropelewski, 1998: Progress during TOGA in understanding and modeling global teleconnections associated with tropical sea surface temperatures. *J. Geophys. Res.*, **103**, 14 291–14 324.
- Tribbia, J. J., 1991: The rudimentary theory of atmospheric teleconnections associated with ENSO. *Teleconnections Linking Worldwide Climate Anomalies*, M. H. Glantz, R. W. Katz, and N. Nicholls, Eds., Cambridge University Press, 285–308.
- Uvo, C. B., C. A. Repelli, S. E. Zebiak, and Y. Kushnir, 1998: The relationships between tropical Pacific and Atlantic SST and northeast Brazil monthly precipitation. *J. Climate*, **11**, 551–561.
- Wallace, J. M., and D. S. Gutzler, 1981: Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon. Wea. Rev.*, **109**, 784–812.
- , E. M. Rasmusson, T. P. Mitchell, V. E. Kousky, E. S. Sarachik, and H. von Storch, 1998: On the structure and evolution of ENSO-related climate variability in the tropical Pacific: Lessons from TOGA. *J. Geophys. Res.*, **103**, 14 241–14 259.
- Ward, M. N., 1998: Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and multi-decadal timescales. *J. Climate*, **11**, 3167–3191.
- , and C. K. Folland, 1991: Prediction of seasonal rainfall in the north Nordeste of Brazil using eigenvectors of sea surface temperature. *Int. J. Climatol.*, **11**, 711–743.
- Waylen, P. R., C. N. Caviedes, and M. E. Quesada, 1996: Interannual variability of monthly precipitation in Costa Rica. *J. Climate*, **9**, 2606–2613.
- Xie, P., and P. A. Arkin, 1996: Analyses of global monthly precipitation using gauge observations, satellite estimates, and numerical model predictions. *J. Climate*, **9**, 840–858.
- , and —, 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539–2558.
- Yulaeva, E., and J. M. Wallace, 1994: The signature of ENSO in global temperature and precipitation fields derived from the microwave sounding unit. *J. Climate*, **7**, 1719–1736.