HE ROLE OF THE SOUTH ATLANTIC IN THE VARIABILITY OF THE ITCZ

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ABSTRACT

In the tropical Atlantic region, the ITCZ and its associated atmosphere and ocean elements are the most outstanding climatic feature. The ITCZ controls rainfall and other climatic impacts (e.g., storms and dust transport) in regions with high population density. Its annual migration north and south and the significant interannual variability thereof have direct impact on society. The ITCZ in the Atlantic is extremely sensitive to small changes in regional surface temperature gradients and external atmospheric influences. Much smaller SST anomalies than those associated with ENSO have significant impact on rainfall in NE Brazil and West Africa. There is modeling evidence of strong coupling between SST and convection variability near the equator that allows for external influences to create significant impacts. This setting implies that anomalies in the South Atlantic atmosphere and ocean can set off an interaction that is as effective in modulating ITCZ variability as the well documented effects of ENSO and the North Atlantic Oscillation. This paper reviews the pattern and impact of ITCZ variability in the tropical Atlantic region to set the stage for understanding how the South Atlantic can impact this sensitive system. Evidence from recent research is presented that the South Atlantic does indeed play a role in ITCZ variability and issues for future research are discussed.

1. INTRODUCTION

The climate, mean and variability, of the South Atlantic (hereafter SA) and of other parts of the vastly water covered Southern Hemisphere, are less well undestood than their Northern Hemisphere counterparts. This is primarily because of the paucity of in-situ data, which have been collected mainly along narrow ship tracks (see *Figure 1*). Moreover, even along the relatively better sampled ship tracks there is a lack of important information such as rainfall and upper-air data. This situation has been somewhat alleviated by the introduction of data from remote sensing satellite instruments since the early 1980s (although such data need to be continually calibrated against in-situ observations to allow the removal of instrumental drifts and biases) and by the ongoing global dissemination of ARGO profiling floats that provide in-situ hydrographic data. Any attempt to better understand the global climate system and its response to natural and anthropogenic forcing should recognize as a goal the improvement of our understanding of the Southern Ocean Basins.

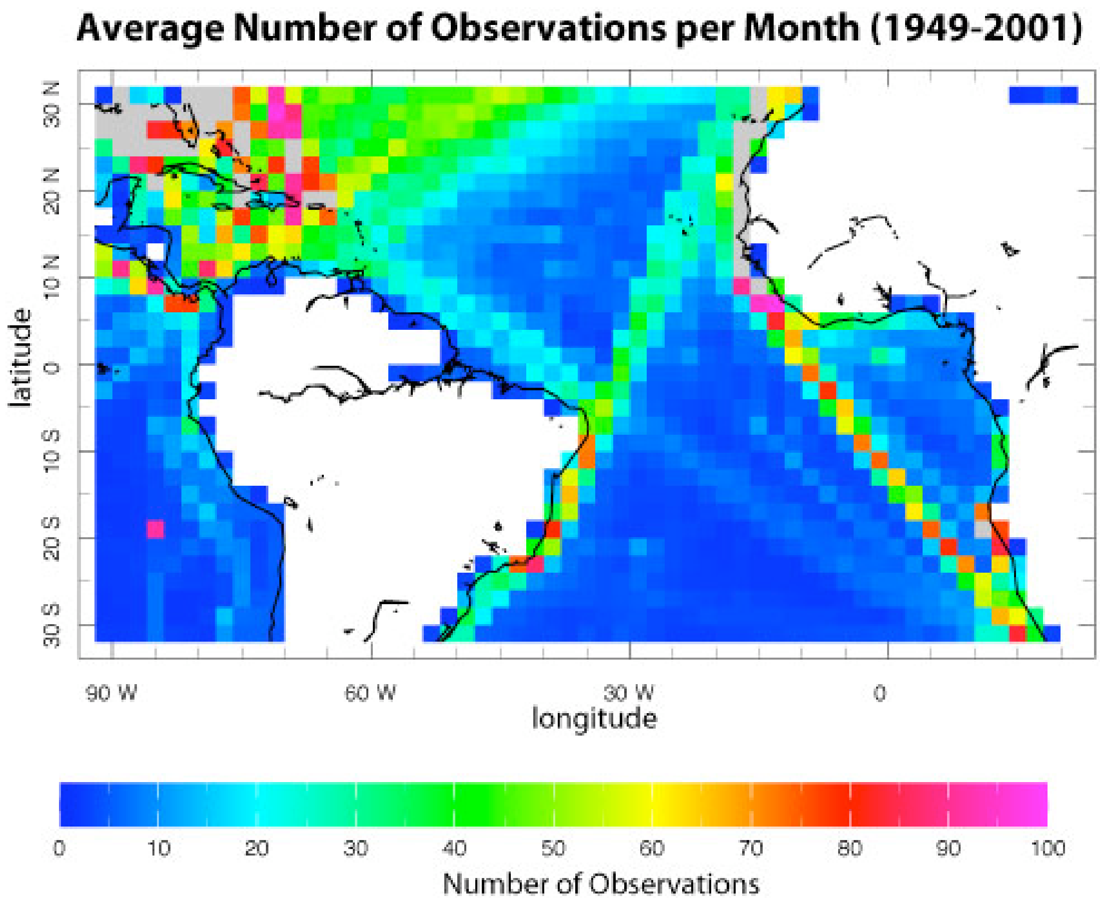

Figure 1: *The year-round average number of ship meteorological observations per months (COADS data) for the years 1949-2001. The regions of relatively large numbers of observations denote the ship tracks.*

In the Atlantic, the climatic variability of the marine ITCZ and the tropical ocean-atmosphere system associated with it is of major consequence to society (Hastenrath and

Heller, 1977). The annual migration of the ITCZ related rainfall and its interannual variability have direct impact on the health and livelihood of millions of people living in equatorial West Africa and northeastern South America (see *Figure 2*). Within this domain, the semi-arid region of Northeast Brazil and sub-Saharan Africa are particularly sensitive to interannual fluctuations in the intensity and position of the ITCZ (*Figure 3*), as it is only during the part of the year that it extends far enough from its annual mean position that these regions receive their rain (*Figures 2, 3* and *4*).

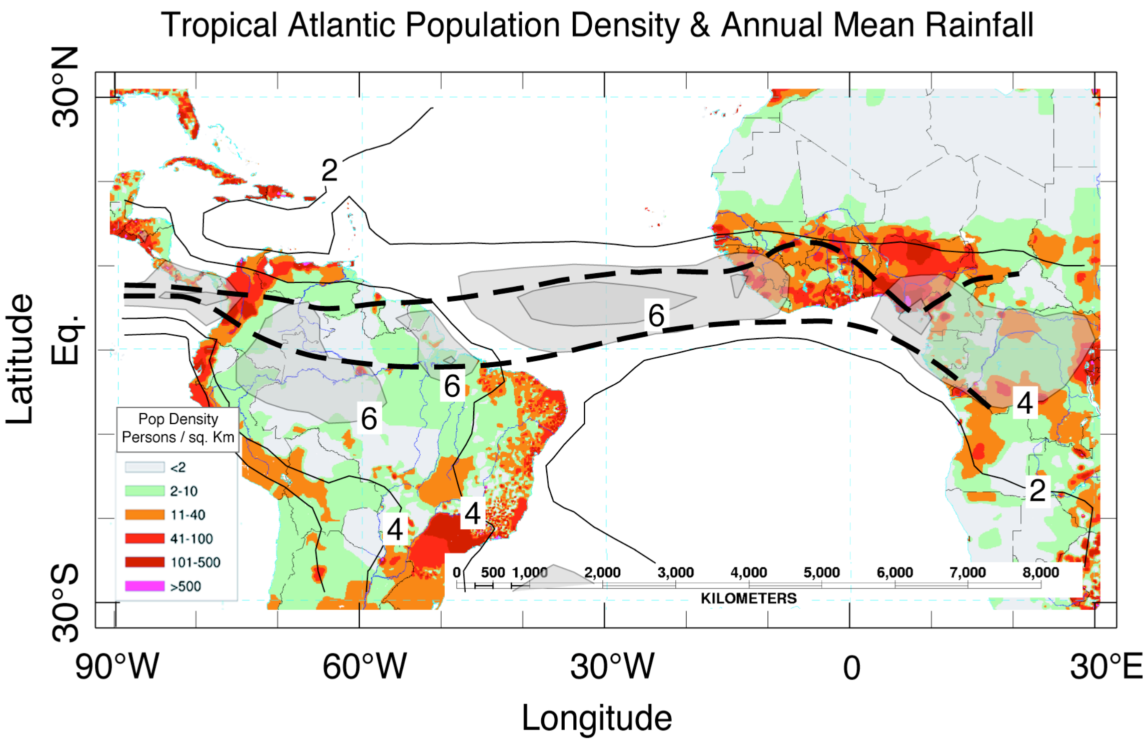

Figure 2: *A map of population density in persons/km^2 (colors, see inserted legend), overlaid by contours of annual mean rainfall (in mm/day) with values > 4 mm/day (~150 cm/yr) shaded in transparent gray to indicate the mean position of the ITCZ. The north (July-August) and south (March-April) limits of the ITCZ climatological annual migration are indicated by thick dashed lines.*
Population density data for 1994 are from Tobler et al., 1995 also available at: http://www.nrcs.usda.gov/technical/worldsoils/mapindx/popden.html. Precipitation data are from NASA/GPCP (Huffman et al., 1997) available from: http://precip.gsfc.nasa.gov/index.html.

The topic of SA and tropical Atlantic (hereafter TA) inter-relationship has been the subject of a handful of studies in the recent few decades. These studies include data and modeling investigations to look at atmosphere-ocean patterns and mechanisms in a domain including the tropics and the extratropical Southern Hemisphere, as well as the entire Atlantic Ocean. Of related importance are studies that investigated the effect of ENSO on both TA and SA regions. The purpose of this white paper is to review the available information on this topic, assess the degree of influence exerted by the SA on the TA, evaluate the importance of this influence to the state of the Atlantic ITCZ, and point at outstanding issues or questions regarding this relationship.

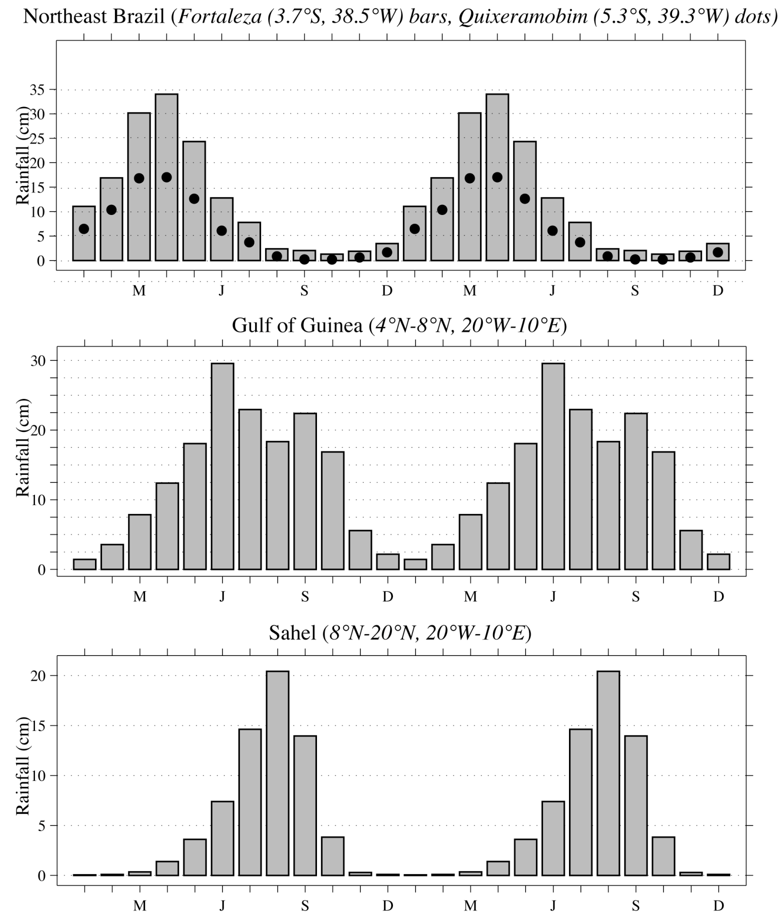

Figure 3*: The annual cycle of precipitation in regions affected by the interannual variability in Atlantic ITCZ position and intensity. Adopted from JISAO website graphics calculated and plotted by T. P. Mitchell (http://tao.atmos.washington.edu/data_sets/).*

The paper is organized as follows: Section 2 reviews our current understanding of mechanisms of interannual ITCZ variability. It compares the role of various interactions internal to the Atlantic (in contrast to ENSO influences) and points at the ways SA influences can be communicated to the TA region. Section 3 examines the known patterns of atmosphere-ocean variability in the SA regions and our current understanding regarding their dynamics, in order to evaluate their effect on the TA. Section 3 also discusses the role of ocean dynamics in the SA that can affect TA SST variability and by inference the ITCZ. Section 4 offers in conclusion an evaluation of outstanding issues and problems related to the SA effect on the TA region and the ITCZ.

2. TROPICAL ATLANTIC ITCZ VARIABILITY

2.1 Annual cycle

In a narrow sense, the Atlantic marine ITCZ (AMI) is the region within the TA Ocean Basin, close to the equator, where the mean rainfall associated with average effect of transient convective systems reaches a maximum. In a broader context the ITCZ represents a complex of phenomena, which include the equatorial low-pressure trough, the trade wind convergence zone, and the regional maximum in SST values. The AMI complex stretches across the TA basin in a nearly perfect, east-west orientation (*Figures 2 and 4*) and migrates north and south with the season, reaching furthest south in the boreal spring (March-April) and furthest north in the boreal summer (July-August) as illustrated in *Figures 1 and 4* (see also Mitchell and Wallace, 1992).

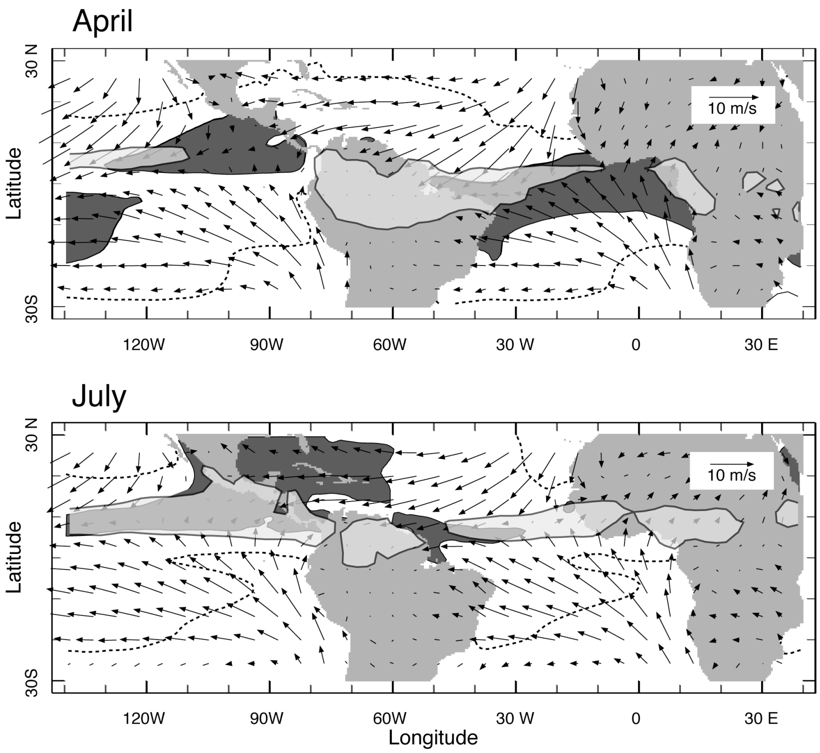

Figure 4*: April and July climatologies of the tropical Atlantic and eastern Pacific. Dark shaded areas are regions with SST ≥28°C. Light, semi-transparent areas are regions with rainfall ≥6 mm/day (the ITCZ). The arrows depict the surface (10 m) wind vectors with scale indicated in the figure. The dotted contour is the 24°C isotherm demarcating the regions of relatively cold water and the eastern ocean cold tongues. SST and wind data are from NCEP/NCAR CDAS-1 (Reanalysis) and rainfall from GPCP.*

2.2 Patterns of interannual variability

Interannual variability of AMI location and intensity is closely linked with interannual SST variability within the broader TA region (Moura and Shukla, 1981; Hastenrath and Greischar, 1993; Nobre and Shukla, 1996; Fontaine and Janicot, 1996; Servain et al., 1999; Chiang et al., 2002). Thus most studies of the AMI variability focused on identifying and studying the patterns anomalies in SST, surface winds, and convection/rainfall and their inter-relationships.

The association between ocean and atmosphere variability depends on the season. During its furthest excursion southward, in the boreal spring, the AMI is diverted from its climatological position towards the anomalously warmer hemisphere. This is manifested in a well-documented pattern (e.g., Nobre and Shukla, 1996; Ruiz-Barradas et al., 1999, Dommenget and Latif, 2000). Here the pattern is derived from a principal component analysis of the March-April averaged rainfall anomaly in the TA Basin (*Figure 5*). The pattern depicts an anomalous (stronger than normal) northward SST gradient in the TA region, i.e., warmer than normal SST in the north TA (NTA) and somewhat colder than normal SST in the south TA (STA). This SST anomaly pattern is associated with a northward cross equatorial surface wind anomaly, with weaker than normal trades in the NTA and stronger than normal trades in the STA. In rainfall this SST-wind pattern is associated with weaker than normal rainfall over the southern flank of the climatological ITCZ position (compare *Figures 5* and *4* during April) and somewhat stronger than normal rainfall to the north. This implies a weakening in the ITCZ strength together with a northward shift in its position towards the warmer hemisphere (see also Chiang et al., 2002). Because the SST variability in this pattern exhibits a north-south contrast, it is often dubbed as the TA "meridional mode" of variability (Servain et al., 1999).

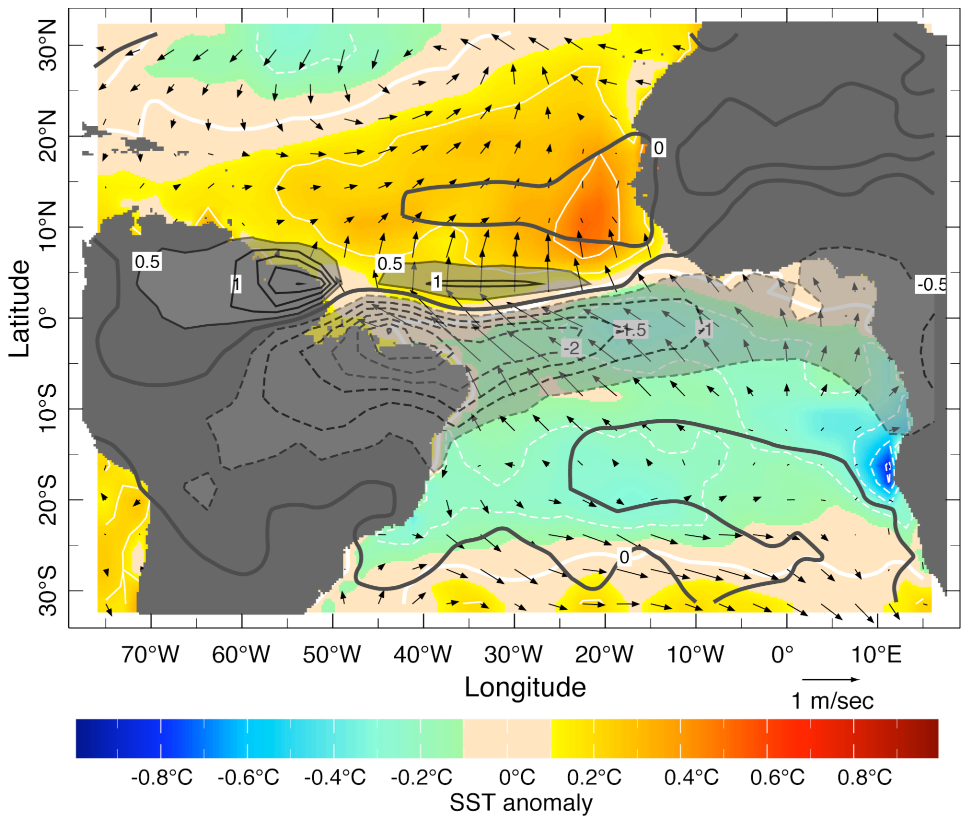

Figure 5*: The dominant pattern of surface ocean-atmosphere variability in the tropical Atlantic region during boreal spring. The black contours depict the first EOF of the regional March-April rainfall anomaly (from GPCP data, 1979-2001) in units of mm/day. This EOF explains 33% of the seasonal variance. The colored field is the March-April SST anomaly regressed on the principal component time series of the rainfall EOF (units are °C, see scale below; white contours every 0.2° are added for further clarity). Arrows depict the seasonal surface wind vector anomaly in m/sec, regressed on the same time series (see arrow scale below frame).*

In the boreal summer season, when the ITCZ moves to its furthest position in the NTA, SST variability peaks just north of the so-called Atlantic cold tongue region, in the equatorial east Atlantic. This is the time when SST reaches its coldest annual state there (see Mitchell and Wallace, 1992) but also the time when it is most prone to the appearance of warm anomalies thought to be similar to El Niño (see Zebiak, 1993 and Carton and Huang, 1994). Unlike their equatorial Pacific counterpart, the warm events in the Atlantic cold tongue persist no longer than a season or two (Carton and Huang, 1994). The pattern in Figure 6 is derived from a principal component analysis of TA rainfall during the summer (June-August, GPCP data1979-2001 is used) and is not entirely identical to an analysis based on the TA cold tongue index (e.g., as in Zebiak, 1993). However, it depicts the phenomenon in a consistent manner: an appearance of a warm anomaly along the equator, on the eastern side of the basin and a convergence of the surface winds towards the warmest region (*Figure 6*, see also Zebiak, 1993). The effect on the ITCZ entails a southward shift and intensification in convection and rainfall (*Figure 6*, see also Carton and Huang, 1994). The associated pattern of SST (and subsurface ocean) variability is also referred to as the TA "zonal mode" of variability (Servain et al., 1999).

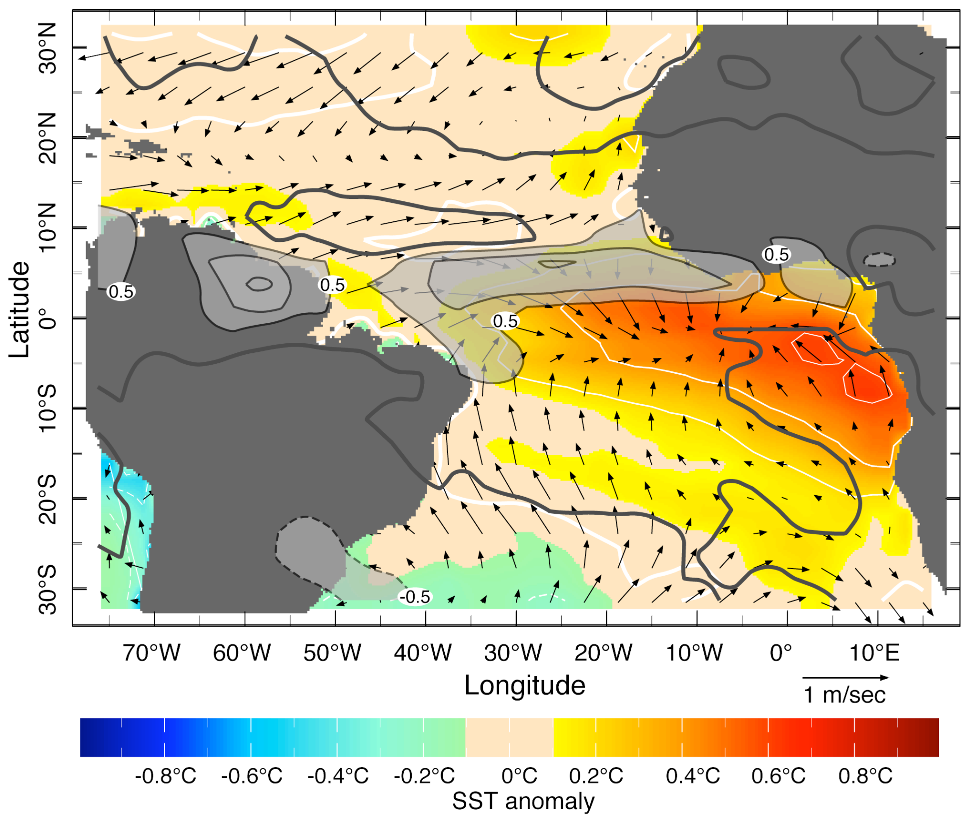

Figure 6*: As in Figure 5 but for the boreal summer season (June-August). The rainfall EOF of this season explains 23% of the variance.*

Mechanisms

The close diagnostic link between SST in the ITCZ location and intensity does not, of course, reveal the cause and effect or the dynamical processes that are at work. In fact, there are serious gaps in our understanding of processes in the TA region. Yet recent diagnostic and modeling studies reveal that there are several key climatic processes that lead to the formation of the variability.

As indicated above, there is some evidence from observations and models that the *boreal summer*, eastern equatorial Atlantic variability involves a Bjerknes mechanism akin to that acting in ENSO; that is, a coupled interaction between the strength in equatorial upwelling and related SST values on one hand, and the atmospheric circulation, most prominently overlying surface winds stress anomalies on the other hand (Houghton, 1991; Zebiak, 1993; Carton and Huang, 1994; Sutton et al., 2000; Chang et al., 2000). There are questions regarding the degree of the subsurface ocean involvement in this variability, particularly the response of the thermocline to the wind forcing. What also seems to be in question is the ability of such variability to self-generate and self-sustaine as ENSO. Influences from the Pacific ENSO (see more below) and the boreal spring mode of TA variability have been implicated in its forcing (Zebiak, 1993; Servain et al., 1999; Murtugudde et al., 2001). As far as the ITCZ is concerned, there are gaps in our understanding of the relationship between this 'zonal mode' and the meridional propagation of the ITCZ and its intensity.

The *boreal spring* variability associated with the meridional SST contrast has received considerable attention since it was first clearly identified (Hastenrath, 1978; 1984; Servain, 1991). While it has become relatively well established that the meridional ITCZ displacement involves a strong mutual interaction between convection, cross-equatorial surface winds, and the meridional SST gradient across the mean latitude of the ITCZ (Hastenrath and Greischar, 1993; Chang et al., 2000, Chiang et al., 2002), how the gradient is initially formed and maintained is still not fully understood. Early investigations suggested that it is associated with heat flux exchange between ocean and atmosphere (Curtis and Hastenrath, 1995; Carton et al., 1996); that is, the SST anomaly centers in the subtropics, particularly in the NTA (*Figure 5*), is generated primarily by surface heat flux forcing due to changes in windspeed (see also Seager et al., 2000; Seager et al., 2001; Kushnir et al., 2002). Such interaction however, can be due entirely to external forcing such as that resulting from ENSO or the North Atlantic oscillation (NAO) as has been indeed suggested based on compelling observational and modeling evidence (e.g., Enfield and Mayer, 1997; Hastenrath et al., 1987; Dommenget and Latif, 2000; Saravanan and Chang, 2000; Chiang et al., 2002; Sutton et al., 2000; Czaja et al., 2002). This is because both these phenomena affect trade wind intensity and hence ocean-atmosphere heat exchange in the NTA.

Evidence that "coupled" air-sea interaction within the near-equatorial region, affects the temporal characteristics of the 'meridional mode' of TA variability is increasing (Chang et al., 1997; Chang et al., 2000; Chang et al., 2001; Chiang et al., 2002; Kushnir et al., 2002; Tanimoto and Xie, 2002). This atmosphere-ocean coupling mechanism relies on the effect that a change in the inter-hemispheric, meridional SST gradient during boreal spring has on the position of the ITCZ (convection and surface wind pattern included). In that season, the climatological SST gradient is at its weakest state (Chiang et al., 2002) and the trade wind system is almost symmetric around the equator. In such state, even small SST anomalies north or south of the equator can tilt the slope of the large-scale inter-hemispheric pressure gradient and affect the ITCZ such that it is displaced towards the anomalously warmer hemisphere (see Hastenrath and Greischar, 1993; Hastenrath, 2000a; Chiang et al., 2002). This, in turn, reinforces the strength of the trades, weakening them further in the anomalously warm hemisphere and strengthening them in the cold one. This effect enhances the heat flux anomaly that lead to the SST formation, particularly in the near-equatorial region, yielding a positive wind-SST-evaporation (WES) feedback (Xie, 1999, see also *Figure 5*). The debate is still on however, whether or not such mechanism can give rise to self-sustained, long-term variability (Xie, 1999; Chang et al., 2000; Kushnir et al., 2002; Okajima et al., 2003).

The role of external influences in TA climate variability, including the effect on the AMI, cannot be overstated. Czaja et al. (2002) ascribe the majority of the strong NTA SST anomalies in the recent half-century to the forcing from ENSO and the NAO. The discussion of the mechanisms associated with the two patterns of variability introduced much of this issue, particularly its role in the meridional mode of the variability due to the effects of ENSO and the NAO on the TA SST. The impact of ENSO in particular is pervasive and thought to come through several "atmospheric bridge" pathways. The impact on the NTA is thought to come through the ENSO impact on the Pacific-North American Pattern (Horel and Wallace, 1981; Hastenrath, 2000b). The ENSO impact on the South Tropical Atlantic is less well established. Mo and Hakinen (2001) have argued for a teleconnection through the Pacific-South American pattern, but acting only the quasi-biennial component of ENSO. A recent coupled model study of the 97-98 ENSO event, also identifies an ENSO impact on south tropical Atlantic SST (Elliott et al., 2001)

ENSO also has a direct effect on the AMI through its influence on the circulation and vertical stability of the tropical atmosphere. It causes a shift in the Walker circulation (and thus enhanced subsidence over the TA) and a warming (and stabilization) of the entire tropical atmosphere surrounding the equatorial Pacific region. These two effects lead to a marked suppression of AMI intensity during the entire ENSO cycle, but particularly during the boreal winter and spring (Hastenrath et al., 1987; Chiang et al., 2002; Chiang and Sobel, 2002). Recently, Giannini et al. (2003) found that for rainfall anomalies in NE Brazil (and the ITCZ location) during boreal spring, the interplay between ENSO influence and the underlying state of the Atlantic interhemispheric SST gradient present at the time of ENSO influence is important. When the gradient is "pre-conditioned" such that the NTA region is warmer than normal before the teleconnection from the equatorial Pacific affect the region, the impact of ENSO is stronger than normal and *vice versa*. The importance of the overall sensitivity of TA climate variability is that it opens the door for influences from the South Atlantic region as well. This topic is discussed in the following section.

3. ROLE OF THE SOUTH ATLANTIC

From the nature of TA variability discussed above it is plausible that remote effects from the SA invoke interannual variations in AMI. It is noteworthy however, that most of the research on TA climate variability has focused on the interactions along the equator and to the north and work on the role of the SA is just beginning. One obvious way for the SA to affect the TA region and the ITCZ is through changes in the SA trade wind intensity, and through that SST in the STA region that influences the ITCZ (see section 2 above). This is then a similar mechanism to that acting in the northern trades, which are affected by such Northern Hemisphere phenomena as the NAO. Other ways by which the SA can be important is through a more active ocean role such as related to changes in the ocean transport of heat, whether at the surface or in the thermocline level. The goal of this section is to examine the processes by which the SA Ocean can affect the TA and the ITCZ. We do not discuss here the general issue of SA ocean-atmosphere variability, a subject deserving separate attention.

3.1 The influence of the SA atmosphere on the TA

A detailed analysis of the patterns of ocean-atmosphere variability in the SA was offered by Venegas and collaborators in a few consecutive studies (Venegas et al., 1996; 1997; 1998). Their work is based on analyses of in-situ data (COADS observations), which are prone to sampling errors (see Introduction and *Figure 1*). However, the analysis results, particularly in term of the leading pattern of joint sea level pressure (SLP) and SST variability is consistent with more recent analyses based on the NCAR/NCEP Reanalysis (e.g., Sterl and Hazeleger, 2002; Barreiro et al., in preparation, see below).

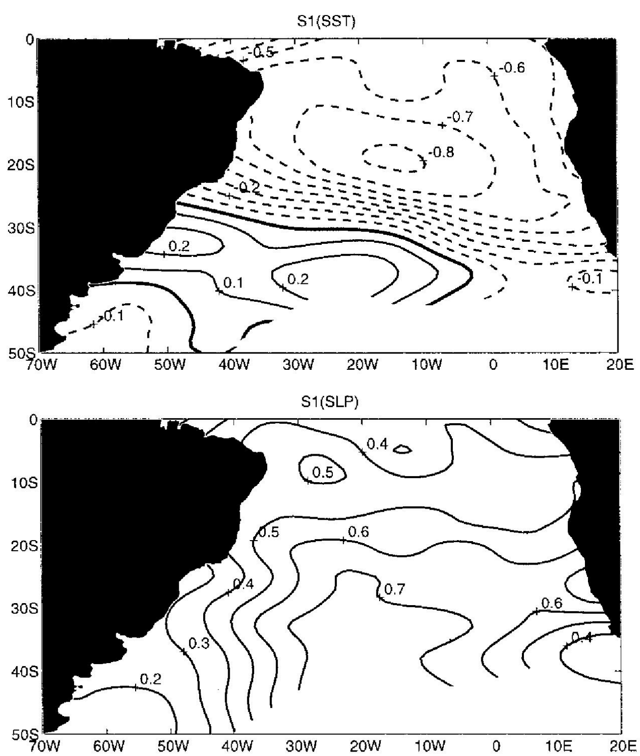

***Figure** 7: The first joint mode of sea level pressure and SST in the South Atlantic from Venegas et al. (1997) based on monthly anomalies from spatially interpolated COADS data, 1953-1992, that were processes with an SVD analysis.*

The leading SA pattern (*Figure 7*) depicts an association between the strengthening of the SA subtropical high (see Venegas et al., 1997 for a figure of the climatological mean state) and a trade wind region SST anomaly north of 30°S. According to Sterl and Hazeleger this pattern explains almost 40% of the combined SST and SLP variance. The relationship between the centers of action in the SLP and SST fields and a lag correlation analysis of the SST and SLP associated time series, where SLP is leading SST by 1-2 months, indicate that fluctuations in the surface atmospheric circulation force the SST variability. This is confirmed by diagnosing the terms in the SST tendency equation in the Sterl and Hazeleger study.

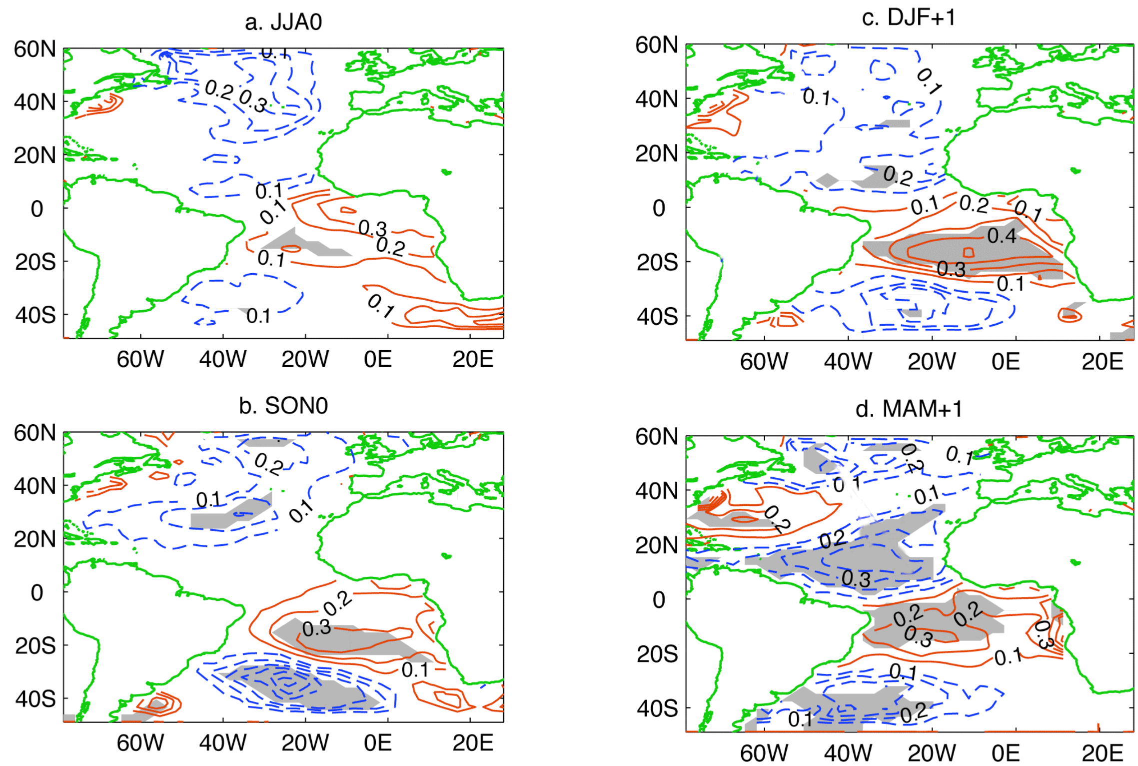

Figure 8*: Composite of observed SST constructed as the difference between negative and positive years of the boreal spring, tropical Atlantic cross-equatorial SST gradient for (top to bottom): the states in the preceding summer, preceding fall, preceding winter, and the concomitant spring. Contour interval is 0.1°C with zero contour omitted and negative (positive) contours in blue (red). Shading denotes statistical significance at the 95% confidence level. From Barreiro et al. (in preparation).*

Barreiro et al. (in preparation) examine the evolution of SA SST and atmospheric circulation associated with the formation of an inter-hemispheric SST gradient during the boreal spring. They begin with selecting all the years in the Reanalysis period where a strong gradient was observed in the absence of an El Niño. Using four years with a strong negative gradient (NTA colder than STA) and four with a strong positive gradient, they form a composite of their difference during boreal spring (March-May) in the entire Atlantic basin. They denote this state MAM+1. They then use the same years to create the composite SST pattern form the previous winter (December of the previous year to January of the same year), previous fall (September-November of the previous year), and previous summer (June-August of the previous year), denoting them DJF+1, SON0, and JJA0, respectively.

The results, shown in *Figure 8*, exhibit an evolution from a weak austral winter (JJA0) SST dipole in the SA, to a strong SST dipole flanking the equator in the following boreal spring. An analysis of the SLP field (not shown) indicates that the process begins by the presence of a strong SH SLP dipole akin to the one discerned by Venegas et al. (1997), forcing the ocean during the austral winter. The SA SST dipole development lags that of the SLP, reaching maximum in the Austral spring (SON0). The southern pole of the SA dipole disappears in the boreal winter. However, the northern pole of this pattern persists and in the boreal spring an SST anomaly of the opposite sign appears in the NTA. Barreiro et al. demonstrate the robustness of this evolution using a large ensemble of coupled GCM integrations (*Figure 9*) where a slab ocean is freely interacting with the atmosphere over the entire global ocean accept the equatorial Pacific, where observed SST from 1950 to 1994 were prescribed. In this model the SA-TA interaction is a coherent "free-mode" of the coupled climate system in the Atlantic, that is, variability that is independent of ENSO. The GCM results (*Figure 9*) show that SA SST dipole forms in response to internal atmospheric variability exerting an anomalous surface heat exchange with the ocean (note that in the GCM composite, unlike the observed one, years with strong NAO anomalies were also excluded). The STA pole of the SA SST anomaly invokes a southward shift of the ITCZ and a surface wind response as early as the boreal winter (DJF+1), consistent with the mechanism explained in Section 2.2 (see also *Figure 5*). This wind response forces an equatorial SST dipole, which persists throughout the boreal spring (MAM+1). Thus there appears to be clear evidence that atmospheric variability in the SA is affects the TA and the ITCZ through the mediation of the more slowly evolving and more persisting upper ocean heat content anomaly.

Looking at the GCM ensemble, Barreiro et al also find that the SA-TA interaction does not influence the Northern Hemisphere beyond the NTA region and vice versa, variability in the Northern Hemisphere does not invoke variability in the SA. This is in contrast to previous studies that found that SST in the STA region is associated with or can invoke Northern Hemisphere variability in the form of the NAO (Rajagopalan et al., 1998; Xie and Tanimoto, 1998; Tanimoto and Xie, 1999; Robertson et al., 2000). The latter studies however, emphasized longer timescale phenomena while Berreiro et al. Focused on interannual variability. In addition the present analysis excluded an examination of the interhemispheric impacts by eliminating *a-priori* years with strong NAO variability.

Robertson and collaborators (Robertson et al., 2003) offer a somewhat different, albeit consistent, view on the influence of SA anomalies in the TA regions. They force an atmospheric GCM with SA SST anomaly patterns derived from observations and examined the response during the austral winter and fall seasons (January through June). Of particular interest is their response to an SST pattern similar to that of Venegas et al. (1997), which displays an evolution from a lattitudinally broad pattern (similar to *Figure 7*) during January-March to a more equatorially confined pattern during May and June (as in *Figure 6*). The model TA atmosphere responds strongly to this SST pattern by creating an anomalous westerly low-level flow when the SSTs south of the equator are warmer than normal. The response is strongest in boreal spring with a strong enhancement of rainfall over the Atlantic cold tongue region and a southward shift of the ITCZ.

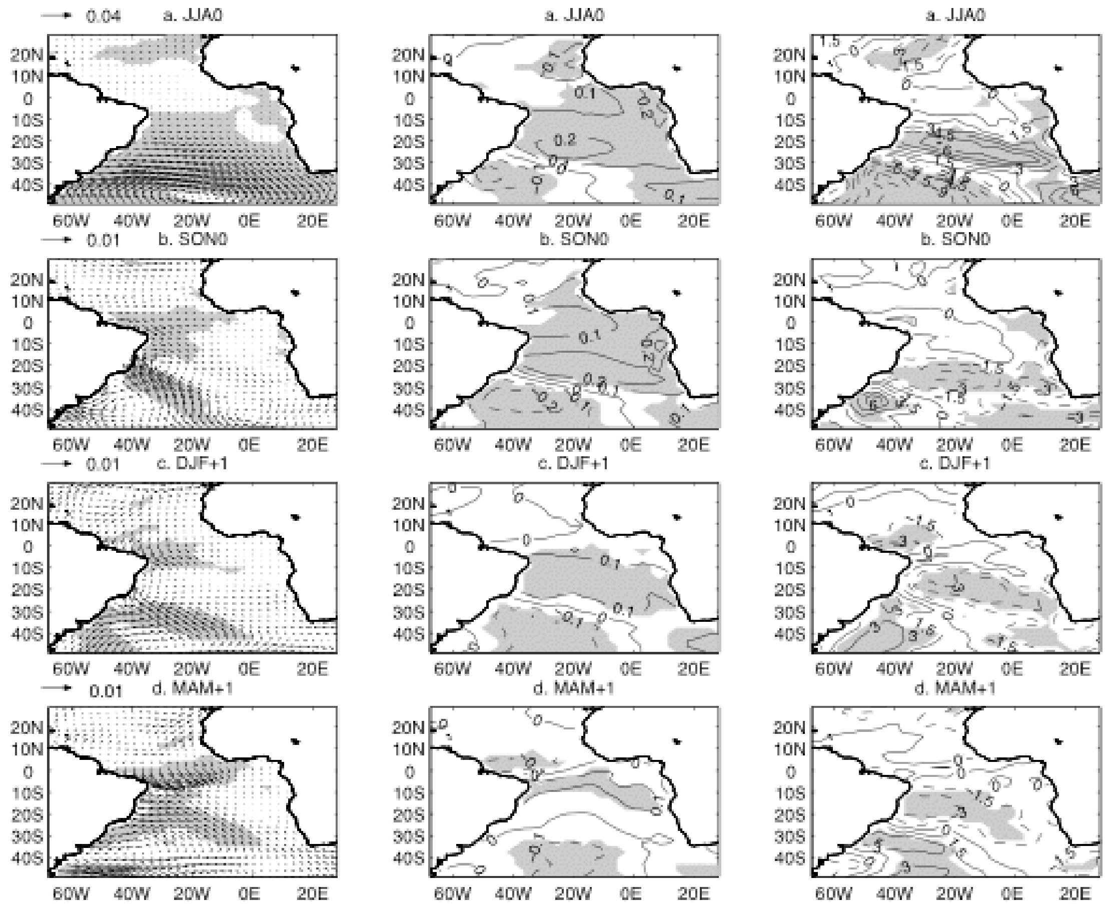

***Figure 9**: Composite evolution of wind stress (left panels, in Nm^{-2}), SST (middle panels, in °C) and net surface heat flux (right panels, in Wm^{-2}) anomalies constructed as the difference between positive and negative years of SA winters with large cyclonic anomaly at 50°S and 20°W in a large ensemble of a atmospheric GCM coupled to a slab ocean (see text above). ENSO years and years with strong NAO forcing were excluded from the analysis. Shading denotes statistically significant regions. From Barreiro et al. (in preparation).*

Overall, as Sterl and Hazeleger (2002) find in a careful budget analysis of the SST tendency, SST anomalies in the SA are mainly generated through heat flux and momentum exchange between the ocean and atmosphere. The former creates thermal anomalies in the mixed layer while the latter creates vertical stirring and horizontal Ekman transport which act on the mean temperature gradient (vertical and horizontal). It is only along the Angolan coast and in the equatorial cold tongue region that they find evidence that ocean dynamics (i.e., transports by ocean geostrophic currents) are important. These ocean mechanisms are discussed in the next sub-section.

3.2 The influence of the SA Ocean

The ocean circulation can impact the climate by affecting SST through lateral advection or/and propagation of anomalies within the mixed layer or the thermocline. Formally there are at least three types of mechanisms through which this can be achieved:

- The transport of temperature or/and salinity (active tracers) anomalies by the mean circulation or a $\overline{V}$ T' mechanism (e.g.; Gu and Philander, 1997; Schneider et al., 1999).
- The transport of the mean temperature gradient by circulation anomalies or a V'$\overline{T}$ mechanism (e.g., Kleeman et al., 1999).
- The coupling of circulation and active tracer anomalies, or a V'T' mechanism, in particular, the propagations of subsurface wave signals (Rossby or Kelvin) within the interior ocean or along coastlines (e.g.; Liu, 1999; Huang and Pedlosky, 1999; Schneider, 2000; Lazar et al., 2001).

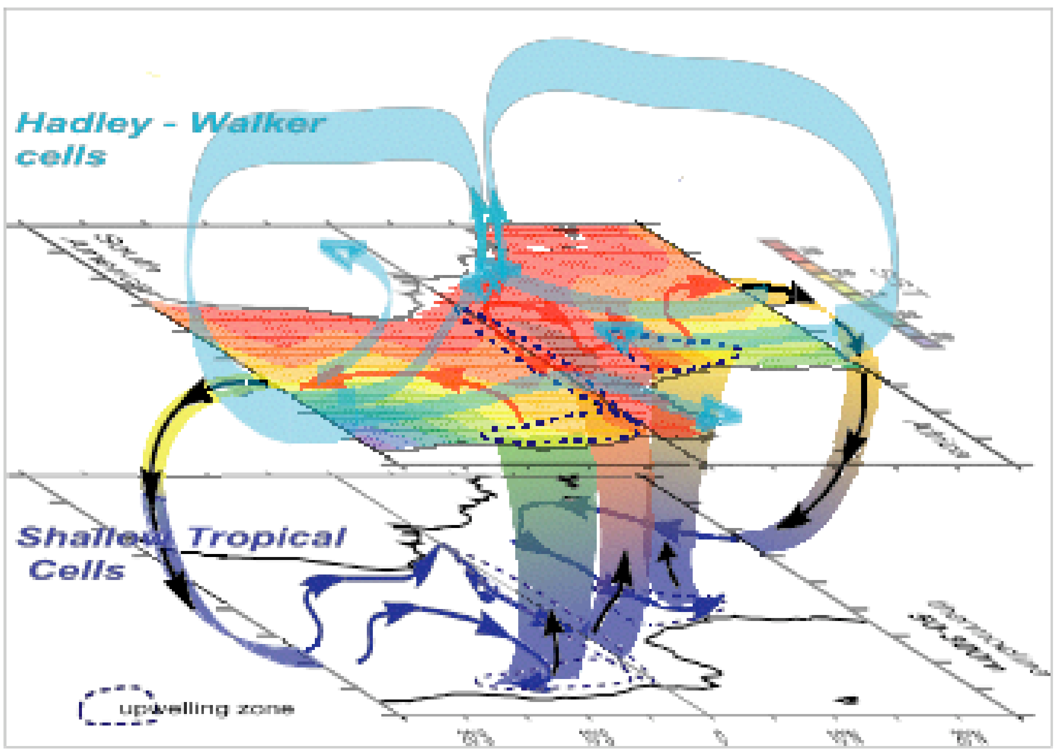

Figure 10*: A schematic 3-dimenstional drawing of the Atlantic shallow tropical cells (STCs, in purple arrows) and the overlying atmospheric, large-scale circulation (in light blue arrows). The climatological SST field is shown in colors on the interfacing surface where the regions of upwelling are also demarcated by blue dashed lines.*

Observations hint that such advective mechanisms are acting in the North and South Atlantic and are manifested in propagation of SST anomalies (Hansen and Bezdek, 1996; Sutton and Allen, 1997; Venegas et al., 1997; Mehta, 1998). Numerical modelling studies with forced ocean GCMs, show that in the trade wind regions, advection by ocean currents within the ocean mixed layer acts to damp thermal anomalies that are forced by surface fluxes (Seager et al., 2001).

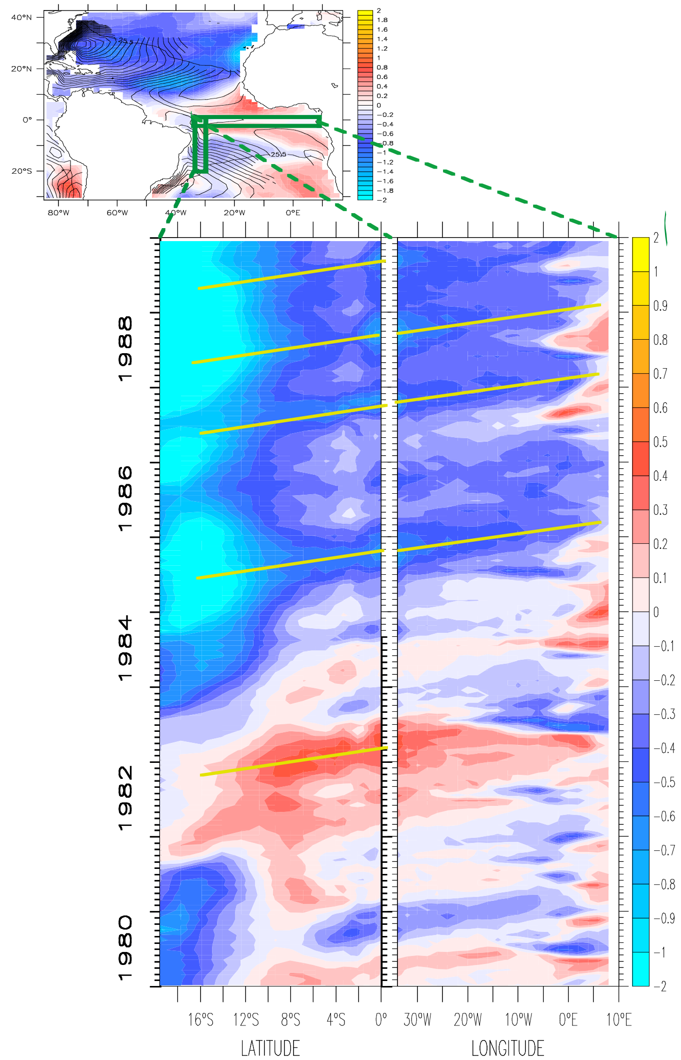

Figure 11*: Propagation of salinity compensated temperature anomalies on isopycnal 25.5, within the thermocline along the Brazilian coast and then within the equatorial undercurrent (EUC) as seen in an ocean GCM forced at the surface by NCEP-NCAR Reanalysis data (see Lazar et al., 2001).*

In the SA however, there is evidence that the geostrophic, wind driven circulation is important in generating the SST anomalies (Hakkinen and Mo, 2002). Other numerical modelling studies suggest that the Atlantic thermocline could be influenced by horizontal advection mechanisms (Huang and Pedlosky, 1999). Less attention has been given to the vertical and turbulent aspect of the problem, that is, how anomalous signals of mass flow or active tracer are passed into or out of the mixed layer with the exception of the present diagnostic study of Sterl and Hazeleger (2002). This aspect is strongly dominated by upwelling and diapycnal turbulent mixing.

Upwelling along the equator, is the main region terminating the subsurface limb of the southern (and northern) shallow tropical circulation (STC, see *Figure 10*) and therefore is the first area where anomalous subsurface signals coming from the SA have the potential to reach the surface and affect the local ocean-atmosphere interaction. It is here where the slow ocean dynamics have the potential to affect the atmosphere by creating surface signals that the ITCZ is sensitive to. Furthermore, due to the importance of atmospheric heating and convection in the ITCZ, it seems plausible this part of the atmosphere that can efficiently transmit anomalous oceanic signals to the global atmosphere (e.g., Xie and Tanimoto, 1998; Robertson et al., 2000).

Several recent model studies looking at particle trajectories or the role of the MOC in the tropical Atlantic through a $V'\overline{T}$ mechanism (Blanke et al., 1999; Fratantoni et al., 2000; Malanotte-Rizzoli et al., 2000, Inui et al., 2002) underscored that in agreement with estimates from observations (Stramma and Schott, 1999) the Southern Atlantic is by far the main source of water of the equatorial thermocline and the surface. MacPhaden and Zhang (2002) have identified within the Pacific a clear control of the equatorial upwelling rate by the tropical and subtropical winds through the STC overturning. A similar or even stronger role for the southern Atlantic winds is plausible, but attempts to detect it has been unsuccessful (McPhaden, personal communication) maybe due to the extreme paucity of data in the region. In addition to these mass sources and their variations, an Atlantic ocean GCM forced with NCEP-NCAR Reanalysis data indicates that there exists a subsurface advective bridge of heat or salt anomalies from the southern subtropics to the equator (a $\overline{V}$ T' mechanism, see Lazar et al., 2001). Zhang et al. (personal communication) found evidence for this mechanism in observations.

This said it is however still unknown if a subsurface heat anomaly can surface at the equator with a significant intensity, especially after its passage through intense mixing region within the North Brazil Current. Nonetheless, a recent model study by Lazar et al. suggests that the spiciness (the temperature and salinity characteristics of an isopycnal layer) of the equatorial undercurrent can be inferred from tropical SST and sea surface salinity (SSS) a few years earlier, thanks to the long-term propagation of salinity compensated heat anomalies (see *Figure 11)*.

As discussed above in Section 2, in the tropical Atlantic, it is the inter-hemispheric SST gradient that is the most influential factor in determining ITCZ variability, only to be seconded by SST anomalies in the equatorial upwelling region (Xie et al. 1998; Sutton et al., 2000). Recent studies (e.g., Lazar et al. 2001) identify two large upwelling/obduction regions in the tropics, which terminate the subsurface limbs of the STCs and therefore are areas where anomalous subsurface signals coming from the subtropics can interact with the surface, through mixing (see *Figures 10 and 12*). These regions coincide in latitude with the two African *coastal upwelling* regions (the Guinea and Angola Domes) commonly invoked for eastern tropical sources of thermocline water, but extend beyond them. In particular, these two large upwelling areas appear to coincide with the Northern and Southern Hemisphere centres of action associated with the variability of the inter-hemispheric SST gradient and thus can influence the ITCZ (compare the extension of the warm SST regions in *Figure 5* and *6* with the upwelling extension *Figure 12*). Even though the strong equatorial upwelling is also providing water to these two regions, their subsurface properties are likely to be intensely modified during the mixed layer transit from the equator and can transmit directly and efficiently the subsurface anomalous signals to the eastern tropical mixed layer. We hypothesize therefore that the STCs, through their influence on SST in the subtropical upwelling areas, provide a way by which *slow ocean processes* modulate the meridional gradient of tropical SST, particularly on long (decadal) time scales.

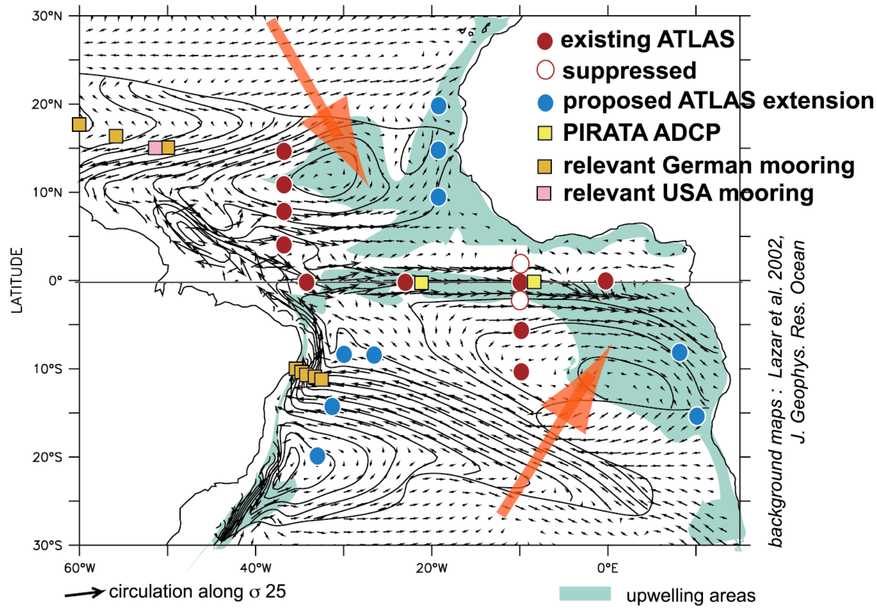

Figure 12: A schematic representation of the returning limb of the tropical Atlantic STCs and the regions of surface upwelling. Red arrows point at the two broad off-equatorial upwelling regions discussed in the text above (after Lazar et al., 2002).

Other potential remote influences of the SA ocean are the *Agulhas Rings*, which arrive at the western equator from the southern tip of Africa and could modify the SST and the Air-sea interactions in the western basin and the *meridional overturning circulation (MOC)* that changes in intensity and could produce effects in a comparable way in its western tropical Atlantic surface limb.

4. SUMMARY AND FUTURE WORK

In the TA the ITCZ is highly sensitive to changes in SST gradients, particularly in the meridional direction and during the boreal spring, but also in summer when the zonal gradient along the equator is strong. There is now sufficient evidence to imply that the SA plays a role in shaping these gradients through surface ocean atmosphere interactions and subsurface, oceanic pathways.

In the realm of the atmosphere interaction with the oceanic mixed layer, a phenomenon akin to the NAO appears to be acting in the SA in which extratropical wintertime variability imprints its signature in the SST field, which through its persistence preconditions the coupled system for a positive feedback interaction in the deep tropics two to three seasons later. On a much slower time scale, thermocline water the SA STC are carrying anomalous temperature and salinity signals to affect the properties of upwelling water in the tropical regions off western Africa and the equatorial cold tongue. These processes are set to affect the variability of the AMI on interannual and decadal time scales and should be considered in addition to the more familiar effects of the NAO and ENSO.

The study of SA influences on the ITCZ is in its early stages and more work needs to be done to understand and quantify the processes described above. Fore example, the link between the chaotic extratropical SA atmosphere and the ITCZ provides potential predictability because of the long delay between the phenomenon's inception in the austral winter and its impact on the ITCZ in the austral fall (boreal spring). Can this potential be materialized? Can the precursors be identified in a robust manner to allow a long-lead prediction? How does this phenomenon interact with other external influences on the ITCZ? How does it interact with the apparently large austral summer variability?

The role of the ocean circulation in affecting longer-term variability needs to be better quantified and understood. The importance of the STCs in setting up the mean properties of the upper ocean is unequivocal, but the impact on variability is not quantified. It is obviously difficult to detect such effects when they underlie the impact of the direct and highly variable surface interactions. Long-term observations need to be available to allow such analysis. The issues that need to be addressed are the effects of the SA STC on upwelling in the eastern part of the basin off Benguela and to what extent this intervenes with the strong surface forcing by the atmosphere there and influences the inter-hemispheric SST gradient. Another important influence occurs in the tropical thermocline and the effect on SST and the formation of Atlantic warm events should be studied.

There are ongoing long-term observation efforts in planning for the SA. It is important to coordinate them with existing efforts in the TA to achieve maximum benefit with addressing the causes of ITCZ variability.

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