White paper
(First Draft) on a

“Tropical Atlantic Climate Experiment“ (TACE)

by

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Executive Summary (to follow)

Preamble

The eastern Tropical Atlantic cold tongue and ITCZ regime is a region of high interest for a better understanding of climate variability, both regional and remote. In particular, its SST variability is significantly correlated with land rainfall and thus of importance for studying the potential for predictability of West African Monsoon variability. The heat budget in the eastern TA is to a large degree determined by non-local exchanges, i.e. by advection via the large-scle circulation, by equatorial and coastal upwelling and associated mixing processes at the underside of the shallow cold tongue, and by equatorial waves. Most coupled models that are not using flux corrections show large deviations from observed SST in that region, displacing the cold tongue far to the west and in fact having a warm eastern equatorial regime **(Fig. 1.0.1)**. While other factors, such as cloud representation, are presumed to contribute to this coupled model deficit, more realistic model representation of the important ocean mechanisms in affecting SST is certainly a major requirement for achieving progress here.

In association with the planning for AMMA, first discussions were held at a Tropical Atlantic Workshop at NOAA/AOML (Miami, March 2003) to outline an ocean program that might take place within the time frame of the AMMA observational periods and focus on improving our understanding of the ocean's role in SST variability. Based on the Miami Workshop results, the CLIVAR Atlantic Implementation Panel, at its last meeting in Nice (France, April 2003) charged a subgroup of the Panel and other experts with developing a "White Paper" for a "Tropical Atlantic Climate Experiment" (TACE), which is presented in sections 1-4. An important prerequisite for a potential observational program is the continued existence of the PIRATA moored station programs, and ACE also relies on the French EGEE program (section 5) with observations in the Gulf of Guinea, that is already ongoing and expected to last throughout the AMMA period.

1. Scientific Basis

1.1 Background on Tropical Atlantic Variability (TAV) and relation to climate anomalies of the region

1.1.1 Geographical definition and importance

The tropical Atlantic (TA) region can be defined as the section of the Atlantic Basin that lies between the tropics of Cancer and Capricorn. It includes the continental areas of Central and eastern South America in the west and West Africa in the east, and the marine areas surrounded by them. The dominant climatic features of this region are the massive convection centers over Africa and South America, the relatively narrow marine Intertropical Convergence Zone (ITCZ) that stretches between them, just north of the equator, and the trade wind systems that converge into the ITCZ from north and south in which shallow convection areas are embedded. At the ocean surface, there are features that roughly resemble the main features of the equatorial Pacific Ocean, that is: an eastern ocean "cold tongue" area and a matching warm pool region on the western side of the basin. These surface features are linked to a three-dimensional circulation, the meridional (Hadley) and zonal circulation cells in the atmosphere and the horizontal equatorial current systems of the upper ocean as well as the shallow meridional overturning cells that connect the equatorial ocean with the northern and southern subtropics. The TA climate system exhibits an intricate, mainly latitudinal, seasonal migration and contains transient components, such as easterly waves and tropical storms, which travel from east (Africa) to west (Central and North America) across the basin.

In the normal seasonal variation of the TA climate system, the migration of the marine ITCZ (hereafter the Atlantic Marine ITCZ, AMI) and the changes in its intensity are of particular importance as they are linked with the year-round variation of rainfall over the ocean and the adjacent land regions, with direct impact on the economies of some of the world's most densely populated and poorest countries, which rely heavily on agriculture. The climate of the countries surrounding the TA creates the necessary conditions for debilitating tropical diseases such as dengue, malaria, cholera, and meningitis, which are sensitive to the variability in rainfall, temperature, and humidity. The largest global source of dust lies in subtropical Africa, which in addition to its immediate impact on humans has considerable influence on the global radiative balance and, via the supply of micro nutrients like iron, on marine biological production and biogeochemical cycles. Also within this region, tropical storms of hurricane intensity regularly inflict tremendous destruction and loss of life on the northern half of the TA, with impact reaching into the Gulf of Mexico and Atlantic coasts of the USA.

The region's vulnerability to climate is underscored by its sensitivity to interannual climate variability. TA interannual-to-decadal climate variability is appreciable and variation in climate conditions (such as storminess, rainfall, humidity, temperature, and dustiness) can lead to disruption of the normal supply of water for agriculture and other human and wildlife needs, to the appearance and severity of various epidemics, and to the probability of tropical storm damages.

Understanding TA climate variability and predictability, with the goal of improving its prediction and identifying and quantifying its relationship to various societal impacts, are important research goals that were recognized by CLIVAR. However despite such programmatic emphasis, actual progress in TA climate prediction has been slow to come and concerted international effort is needed to improve the understanding of the regions' climate variability and the mechanisms that underlay its observed behavior.

1.1.3 The climate of the tropical Atlantic: mean annual cycle

The center of the TA climate system, both in terms of climatic significance and of the dynamical focal point for the annual and interannual variability is the marine ITCZ complex, which includes the wind convergence zone with its convection region and precipitation maximum, the surface low-pressure trough, and the maximum in regional SST distribution

(Fig. 1.1.1a). The climate mean annual cycle and its variability are manifested in the coherent variation of the entire complex.

Throughout the year, the coherent structure of the Atlantic ITCZ complex (AMIC) migrates north and south, staying largely parallel to the equator across the Basin with a slight inclination to the north in the eastern part of the basin (larger in some months than others, **Fig. 1.1.1a and b**). In boreal spring (April-May) the AMIC attains its southern most position with its core reaching 5°S in the west, over the northeast coast of Brazil, but staying slightly north of the equator in the Gulf of Guinea region in the east. In boreal summer (July-August) the AMIC moves furthest away from the equator to 8-10°N (**Fig.1.1.1b**).

The underlying surface conditions in the two extreme seasons are quite different (**Fig. 1.1.1a**): In boreal spring a relatively weak and broad region of marine convection, strongest in the western equatorial region, is located over a wide strip of warm SSTs with weak latitudinal gradients. In the boreal summer the band of AMI precipitation is sharp and stretches across the entire ocean basin with largest values in the east. The band of warm SST is relatively narrow surrounded by strong latitudinal gradients, particularly to the southeast, where the Atlantic cold tongue resides.

The relationship between SST, convection, and surface winds was studied by Mitchell and Wallace (1992). They emphasized the dominance of the first harmonic of the annual cycle in the pattern of ITCZ seasonal variability and proposed that the reasons for this behavior lies in the response of the tropical atmosphere-ocean system to the variations of insolation in the presence of a north-south asymmetry of the distribution of land masses around the equator (particularly in the eastern boundary from which the trade winds are blowing). In particular, they note that it is the development of the massive convection centers over land (in the Atlantic case, the west African monsoon) during late spring, early summer, that leads to the development of the cross equatorial flow in the east, which in turn forces equatorial upwelling and advection of cold water from the Southern Hemisphere and the development of the cold tongue (CT). The development of the cold water leads to rise in sea level pressure over the equator, which further enhances the northward flow, which assists in the development of the monsoon. Over the ocean, the presence of warm water at ~7°N and the airflow from the south, contribute to the creation of a strong, and well-defined region of ITCA convection. This positive feedback interaction between ocean and atmosphere, claim Mitchell and Wallace, is what keeps the marine ITCZ well to the north of the equator well in fall, when the convection over land begins to move south of the equator. In the western part of the TA Basin the northern marine ITCZ position in the fall, keeps the northeastern part of Brazil dry, despite the location of land convection.

In many of it characteristics, the AMIC is not different than its eastern equatorial Pacific counterpart. Here however, the extreme summer position of convection is somewhat south of that of the eastern equatorial Pacific system and is attained earlier in the year, making the annual migration shorter in time from boreal latitudinal minimum to summer maximum and longer from the latter to the former. In contrast the Pacific migration between peaks is even. The explanation for this is not well studied and may be related to the presence of the dry Saharan desert to the north.

1.1.3 Ocean circulation and heat storage changes

a) Western and central tropical Atlantic

In association with the annual migration of the ITCZ is a strong dynamical response in the ocean which leads to major changes in the prevailing surface currents and upper ocean heat advection and storage patterns **(Fig. 1.1.2)**. As the ITCZ migrates northward across the equator in boreal spring to its maximum meridional position in late summer, (i) the cross-equatorial flow at the western boundary accelerates leading to an intensification of the North Brazil Current (NBC) and a growing extension of the NBC retroflection, (ii) the North Equatorial Countercurrent (NECC) turns on in the western part of the basin carrying retroflected NBC flow into the ocean interior **(Fig.1.1.2a),** and (iii) Ekman transport from the tropics to subtropics reaches its seasonal minimum and the "excess" upper ocean waters are stored in the NECC ridge at 3-5º N.

During boreal winter, when the ITCZ migrates to its southernmost position, the NBC diminishes in strength, the NBC retroflection and NECC in the western basin largely vanish **(Fig.1.1.2b)**, and the warm waters stored in the NECC ridge during summer are swept northward into the subtropics by reestablished Ekman transport over the region accompanying the return of the northeast trades. The net result of these actions is a large annual redistribution of upper ocean heat content in the northern tropics and a concomitant annual signal in the meridional heat transport (of nearly 0.5 PW across 8º N), often referred to as the annual "storage and release" phenomenon (Philander and Pacanowski, 1980, Fratantoni et al., 2000).

The impact of these upper ocean circulation changes on the surface heat balance and seasonal SST patterns is not well established, though generally thought to be secondary to direct surface forcing of the upper ocean by radiative and latent heat fluxes. The role of advection for the annual cycle of SST has been analyzed by Foltz et al. (2003). This heat budget study shows that in the northwestern ITCZ the heat balance is mostly local, i.e. between net surface heat flux and heat storage change, with ocean advection of little effect. However, in the equatorial cold tongue the ocean's role is important through three mechanisms: First, mean zonal and meridional advection; second, eddy fluxes, mostly due to Tropical Instability Waves (TIWs); and third, mixing at the bottom of the shallow mixed layer. All terms show strong seasonal variations. Jochum et al. (2003) describe the heat budget of the upper layer of the tropical Atlantic in a high-resolution ocean model. They show that the vertical eddy fluxes of heat compensate the horizontal eddy fluxes of heat associated with TIWs.

b) Eastern tropical Atlantic

The Gulf of Guinea (GG) is the place where the variability of the upper ocean appears in the most obvious way through SST anomalies, which extend southward along the coast and strongly influence the hydrological conditions and coastal upwelling, areas of obvious economical interest for the surrounding countries. The SST is conditioned by advection (horizontal and vertical), purely equatorial dynamics (waves and upwelling) and by the exchanges at the ocean-atmosphere interface. Typical SST amplitudes observed in the GG of the diurnal, seasonal and interannual signal are about 0.5°C, 5°C and 2°C, respectively, and are thus of primary importance if one considers their impact on the exchanges of turbulent fluxes between the ocean and the atmosphere.

This regime involves an average circulation of the surface ocean towards the Northwest and the West, the South Equatorial Current (SEC). The large seasonal SST variability of the GG is due to the low thickness of the surface homogeneous layer and to the rising of cold sub-superficial waters along the equator and the coasts (Merle, 1983). However, the upwelling cannot be solely explained by the local wind regime (Houghton, 1976; Voituriez, 1981) but also by that of the winds over the whole basin, via the equatorial Kelvin waves (Houghton, 1983; Katz, 1987; McCreary et al., 1984). Furthermore, the equatorial upwelling maintenance depends upon the trade wind regime, but also upon vertical and horizontal advection, vertical mixing and fluxes associated with the instability waves (Gouriou and Reverdin, 1992; Foltz et al., 2003).

In the north of the GG, the surface circulation is dominated by the eastward Guinea Current (GC), often considered as a continuation of the North Equatorial Countercurrent (NECC) in boreal summer-fall, when the CCEN is fully developed across the whole basin (Fig. 1.1.2a; Richardson and Walsh, 1986; Arnault, 1987; Stramma and Schott, 1999). In the southern hemisphere along the African coasts, the surface current is southward up to about 5°-8°S, and is mostly supplied by equatorial counter and under currents waters (Gordon and Bosley, 1991; Wacongne and Piton, 1992). Around 7-8°S, the South Equatorial Countercurrent (SECC) joins in the northern branch of a tropical cyclonic gyre, the Angola dome (Shannon et al., 1986), the coastal part of which is associated with the southward Angola Current (AC), that disappears around 15°S at the latitude of the Benguela Front (Meeuwis and Lutjeharms, 1990).

This coastal zone exhibits noticeable interannual events, with SST anomalies exceeding 8°C as observed in March 1995, similar to the Pacific El Niño events, the "Benguela Ninos" (see section 1.3). These events, characterized by anomalies of SST and of meridional gradient of pressure and dynamic height in the GG (Reverdin, 1985), are also correlated with anomalies of the ITCZ latitudinal position (Servain et al., 1985; Picault et al., 1985). The observed SST anomalies and heat content anomalies within the mixed layer may also be linked to the SSS. The strong water freshening observed in the GG, mostly due to the dominant precipitation (exceeding evaporation) in the deep GG (Biafra Bay) and to the Congo River discharge around 6°S, are able to induce barrier layer effects, as observed off the Amazon River in the western basin (Pailler et al., 1999), and thus to affect these parameters.

c) Subsurface flows

Equatorial and off-equatorial upwelling in eastern boundary domes determines SST in a large part of the TA. The supply of the upwelling regions is foremost by the EUC but also, at somewhat deeper thermocline levels, by the North and South Equatorial Undercurrents (NEUC, SEUC) at 3-5^0 latitude (**Fig. 1.5.1**). The NEUC and SEUC diverge poleward on their way east and rise, partially supplying the upwelling of the Guinea and Angola domes, respectively (see section 1.5). The respective potential roles of the eastward undercurrents in TAV need further exploration.

1.1.4 Land feed backs

As the tropical Atlantic has a relatively small basin, land effects on climate received quite a lot of attention. Desertification and human influence on the Sahel region has been widely studied. However, recent studies show that the land effects on low frequency climate variability are relatively small. A summary of feedbacks can be found in Rowell et al. 1995 and is shortly repeated here.

Reduced soil moisture during drought lessens supply for evaporation needed for convection and can act as positive feedback on rainfall. Similarly, surface albedo changes through lack of vegetation leads to reduced rainfall (Otter man 1994). However, both observations (Corel et al. 1984) and modeling efforts (Folland et al. 1986) show that this effect is smaller than the effect of SST anomalies on rainfall variability. Reduced vegetation also affects surface roughness and thereby low level convergence and generates more dust (Xue and Shukla 1993, Nicholson 1988). Rowell et al. 1995 suggest that all land processes have a smaller effect than was concluded by earlier papers, partially because of unrealistic prescribed land property perturbations in the models. The effect of dust, however, has not been studied well yet in climate models.

Convection over the continents is important for the local hydrological cycle. While the marine ITCZ received a lot of attention of modellers, convection over land has received less attention. However, the convection over land does seem to play an important role in setting up teleconnections from the Tropical Atlantic to the extratropics. As shown by Terray and Cassou (2002), Okumura et al. (2001) and Robertson et al (2003) the diabatic heating anomaly associated with anomalous convection over the Amazon basin can affect North Atlantic climate modes.

1.1.5. Interannual variability

The pattern of seasonal AMIC migration varies from year to year and these changes are the predominant source of climate variability in the TA region. The most notable climate impacts in the region, the variability of rainfall over NE Brazil and the coastal regions surrounding the Gulf of Guinea, and the fluctuations in rainfall and dustiness in sub-Saharan Africa (Sahel), are tied in with anomalies in the ITCZ seasonal position and intensity. There is a close link between anomalies in the ITCZ position and intensity and anomalies in SST within the TA Basin (Moura and Shukla, 1981; Hirst and Hastenrath; 1983; Hastenrath and Greischar, 1993; Nobre and Shukla, 1996; Uvo et al., 1998; Ward, 1998; Ruiz-Barradas et al., 2000; Giannini et al. 2003). The tropical ocean-atmosphere interaction is strongly seasonally dependent with the boreal spring variability different from that during boreal summer.

During the boreal spring, rainfall variability is well correlated with the difference between north TA (NTA) and south TA (STA) SST, associated with broad, basin-size SST anomalies in the tradewind belts (**Fig. 1.1.3a**). This SST pattern has been referred to as the "meridional gradient mode" (section 1.2). The meridional SST gradient seems to be determined by the atmospheric forcing of SST anomalies in the trade wind regions, north or south of the equator, that do not appear to be synchronized with one another (Houghton and Toured, 1992; Nobre and Shukla, 1996; Wagner, 1996; Rajagopalan et al., 1998; Enfield et al., 1999).

The ocean's role in this process is not clear at present: In the short term (interannual) the ocean creates a damping effect on the emerging large-scale SST anomalies by horizontal advection (mean and Ekman, Seager et al., 2001). There is also a possibility of feedback from shallow overturning cells that link the tropics and subtropics (the so-called "STCs" – see section 1.5). Close to the African coasts however, upwelling is working in the same direction as the surface heat flux effect, cooling the mixed layer when the trades are stronger than normal and warming when they are weaker.

During boreal summer the variability is well correlated with SST anomalies along the equator and to the south in the eastern equatorial cold tongue region (**Fig. 1.1.3b**). In this case too, the rainfall increases on the anomalously warmer side of the mean ITCZ leading to an impact on the Guinean coast. This phenomenon has led to speculation that a Bjerknes mechanism, akin to the one acting in the eastern equatorial Pacific, can be at play, with ocean and atmosphere dynamics leading to create a positive feedback and oscillatory interannual behavior (Houghton, 1991; Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas et al., 2000). This variability has been referred to as Atlantic ENSO (**Fig. 1.1.4** and section 1.3). However, observations and models suggest that there is no self-sustained behavior here and that the phenomenon is damped rather quickly throughout a single summer season making the prediction of this phenomenon rather tricky, particularly in the boreal fall. It is not clear why this is so, but some studies suggested that this is due to the physical properties of the region and the mean properties of the equatorial thermocline structure in the east.

The AMIC also exhibits a strong sensitivity to remote influences from outside the TA region, in particular from the eastern equatorial Pacific region. ENSO warm events weaken the Atlantic marine ITCZ rainfall in the boreal late winter and spring (Hastenrath, 2000; Chiang et al., 2002). This influence works in part through a direct atmospheric bridge (through the equatorial waveguide) and in part through the influence exerted by ENSO via the Pacific/North American (PNA) teleconnection pattern on tradewind intensity and SST in the western NTA (Covey and Hastenrath, 1978; Enfield and Mayer, 1997; Chiang et al, 2002). The sensitivity of the Atlantic marine ITCZ to SST variability north and south of the equator exposes the AMIC to influences of other phenomena that affect the intensity of the trades, such as the North Atlantic Oscillation (NAO) and similar low-frequency oscillations in the Southern Hemisphere (Czaja and Marshall, 2002; Bareiro et al, 2003).

Global climate models display various levels of systematic errors in the TA region. These errors tend to be largest in coupled models **(Fig. 1.0.1),** but their sources can be detected in the uncoupled versions of those models. The most disturbing error is the tendency of most coupled GCM to reverse the sign of the east-west SST gradient on the equator (i.e., put the colder SST on the western side of the basin). Associated with this error is a tendency for coupled models to move the AMI considerably south of the equator in the boreal spring. Part of the problem may arise from the inability of many atmosphere models to simulate sufficient stratus cloud cover. Another problem may arise from too coarse resolution in most ocean models, which cannot resolve adequately the coastal upwelling region. These problems are amplified in the coupled mode by air-sea interactions.

1.2 The Meridional Gradient Mode

1.2.1 Observational evidence

Early interest in the relationship between tropical rainfall and anomalies of sea surface temperature (SST) was motivated by observational studies of rainfall fluctuations in Northeast Brazil (Hastenrath and Heller, 1977; Moura and Shukla, 1981). The Nordeste, a part of Brazil dependent on agriculture, has a strongly seasonal cycle in which much of the annual rainfall occurs in the months of March through May when the ITCZ is at its southernmost position. The great drought of 1958 forced 10 million people to emigrate from the Nordeste (Namias 1972). These studies found by matching wet and dry years in the Nordeste with patterns of SST that drought associated with an anomalous northward shift of the ITCZ occurred in conjunction with an anomalous northward gradient of SST, an association often referred to as the Atlantic dipole. An intense example of this circumstance occurred in 1993 (Rao et al., 1995). Somewhat weaker relationships have also been identified between the northward gradient of SST and rainfall anomalies in West Africa (Hastenrath, 1990; Lamb and Pepper, 1992).

Many observational studies that followed can be roughly divided into those limited to examining oceanic variables and those looking for covariability between the atmosphere and ocean. Early principal component analyses of SST variability followed (Weare, 1977; Servain, 1991) seemed to confirm the presence of a pattern of variability in SST that was geographically stationary, with decadal time-scales. However, it was pointed out by Houghton and Tourre (1992) and confirmed by Enfield and Mayer (1997), Mehta (1998) and

Rajagopalan et al. (1998) that when the assumption of spatial orthogonality of the principal components is dropped then the northern and southern hemispheres tend to act independently. However in observational studies in which both atmospheric and oceanic variables were included such as Nobre and Shukla (1996), Ruiz-Barradas et al (2000), or Wang (2001; **Fig. 1.2.1**) the results again indicated the presence of a stable pattern of variability across the equator.

The meridional mode identified by Ruiz-Barradas et al (2000) is most pronounced in spring when it is the primary principal component. The SST pattern is most pronounced in the northern hemisphere and is accompanied by meridional wind anomalies along the equator heading down the pressure gradient and thus into the warmer hemisphere. Away from the equator the pattern of anomalous wind stress corresponds to an increase in surface winds in the cool hemisphere and a decrease in the warm hemisphere. A dipole pattern of diabatic heating is in its positive phase, reflecting enhanced convection, in the warm hemisphere, also associated with anomalous deepening of the mixed layer.

1.2.2 Factors affecting the gradient mode

While observational studies in the literature disagree on how to characterize TA climate variability in terms of empirical modes, they agree on the following points:

i) the meridional position of the Atlantic ITCZ is sensitive to the anomalous cross-equatorial SST gradient (CESG), especially in February-April when the ITCZ is at its southernmost position and the seasonal CESG is weak;

ii) a meridional dipole configuration of SST anomalies, although it rarely occurs, maximizes the anomalous CESG;

iii) off-equatorial SST anomalies are associated with changes in the strength of the easterly trades on either side of the equator/ITCZ.

Point i) is supported by ocean GCM studies (Carton et al. 1996; Seager et al. 2001) while point iii) by atmospheric GCM results (Moura and Shukla 1981; Chang et al. 2000; Sutton et al. 2000). Coupled modeling studies generally support the notion that interannual variability in cross-equatorial SST gradient and the Atlantic ITCZ is coupled and involves some degree of their mutual interaction (Chang et al. 1997; Xie and Tanimoto 1998; Xie 1999; Chang et al. 2001; Kushnir et al. 2002). The anomalous trades, the wind-induced changes in surface evaporation and SST anomalies are organized into spatial patterns indicative of a positive wind-evaporation-SST feedback. These findings were mostly obtained with hybrid coupled models. As stated above, fully coupled ocean/atmosphere models have strong biases in the tropical Atlantic. However, it appears that with careful tuning for the climate in the tropical Atlantic with a relatively coarse resolution model good results could be obtained (Haarsma and Hazeleger in prep.).

The tropical Atlantic is subject to strong external forcing. Through the PNA teleconnection and an anomalous Walker circulation subsidence **(Fig. 1.2.1b),** ENSO warming in the equatorial Pacific reduces the northeasterly trades and gives rise to a delayed warming in the northern tropical Atlantic (Enfield and Mayer 1997; Klein et al. 1999; Saravanan and Chang 2000; Huang et al. 2002). The impact of ENSO is strongest in boreal winter. Chiang et al. (2002) reports that when the tropical Pacific is anomalously warm, a stronger than normal Walker Cell suppresses precipitation in the tropical Atlantic. NAO also modulates the strength of the northeasterly trades and hence SST in the subtropical North Atlantic. Such external forcing of the northeasterly trades explains a large percentage of observed SST variability in the northern tropical Atlantic, which subsequently triggers the

CESG-ITCZ interaction in the deep tropics and induces changes on and across the equator (Xie and Tanimoto 1998; Chang et al. 2001; Wu and Liu 2002; Czaja et al. 2002). There may also be an impact via extratropical South Atlantic climate variability (Mo and Häkkinen 2001) in addition to the impact of ENSO on the Walker Cell. From the South Atlantic Robertson et al. (2003) show an impact as well. A dipole in the South Atlantic that constitutes of a warming around 20 S and cooling around 35 S (and vice versa) induces anomalously westerly low level flow in the tropical Atlantic, increased rain fall over the eastern cold tongue and southward shift of the ITCZ. The response is strongest in boreal spring. The response of the ITCZ in the Sahel to SST in the South Atlantic, tropical Pacific and Indian Ocean has been highlighted recently by Giannini et al. (2003).

All these TAV mechanisms are highly seasonal: ENSO and NAO forcing is strongest in boreal winter; the CESG-ITCZ interaction in March-May when the equatorial Atlantic is uniformly warm; and the equatorial mode is most pronounced in the boreal summer coinciding with the season of the cold tongue and the shallow thermocline in the east.

While the CESG-ITCZ interaction almost certainly exists, it remains to be determined how far this meridional air-sea interaction extends toward the poles. There is observational evidence for a positive SST-low cloud feedback over the subtropics (Tanimoto and Xie 2002), indicative of atmospheric reaction in the planetary boundary layer. This mechanism received relatively little attention, but is also visible in models (e.g. Okumura et al. 2001). Over the ITCZ, the cloudiness is associated with surface wind convergence and the radiation feed back acts as negative feedback, but broader stratiform cloudiness anomalies in subtropics are not associated with surface wind convergence and a positive feedback can operate. Surface cooling increases static stability at top of planetary boundary layer and hence the amount of stratus. These enhance reflection of sunlight and amplify the surface cooling. Little is known about the magnitude of this effect and AGCMs have trouble simulating it associated with bad representation of stratiform cloudiness.

Some modeling studies suggest that the CESG-induced shift of the Atlantic ITCZ and resultant upper-tropospheric divergence force a barotropic response modulating the strength of the North Atlantic subtropical high and the northeasterly trades (Okumura et al. 2001; Sutton et al. 2001; Terray and Cassou 2002). This suggests that CESG/ITCZ in the deep tropics might interact with the subtropical Atlantic, a mechanism that may give rise to a pan Atlantic pattern of anomalies of SST, SLP and surface wind that seems most pronounced on decadal timescales in instrumental observations (Xie and Tanimoto 1998; Tourre et al. 1999). This pan Atlantic pattern has been used to explain a link between the tropical and high-latitude North Atlantic observed in paleoclimate records (Peterson et al. 2000; Chiang et al. 2003).

Predictability studies for TAV generally show improved hindcast skills in models initialized with SSTs in the eastern equatorial Pacific and tropical Atlantic (Hastenrath and Greischar 1993; Penland and Matrosov 1998; Chang et al. 2003). Initial SST anomalies in these regions allow inclusion of ENSO teleconnection and the CESG/ITCZ interaction within the tropical Atlantic, respectively. This result is consistent with the diagnostic/modeling studies showing the importance of both ENSO forcing and local air-sea feedback, as expressed in **Fig. 1.2.2b** from Sutton et al. (2000). It shows the partition between internal, i.e. stochastic and non-predictable, and SST-induced variance in an atmospheric model driven by observed SSTs. The meridional equatorial wind is highly dependent on the SST gradient ("dipole" in Fig. 1.2.1b), but also dependent on ENSO (for the winter months), and on the equatorial Atlantic SST ("atl3").

More recently, Joyce et al. (2003) studied the wind stress curl variations associated with the cross-equatorial wind stress variations and found that northward (southward) stress anomalies are accompanied by substantial negative (positive) wind stress curls. These curl variations then drive cross-equatorial Sverdrup transport variations from the warm to the cold side of the equator, i.e. against the wind stress, with a time lag of a few months. The Sverdrup transports thus serve to dampen the SST dipole.

1.3 Atlantic ENSO and the Benguela Nino

Superimposed on the primarily annual cycle of SST are anomalies during the boreal summer months (JJA) that frequently exceed 1^{o}C during the peak month. Warm anomalies are generally maximum in the zone of the boreal summer cold tongue, while cool anomalies tend to have broader temporal and geographic variability. During some years, but not all, the warm anomalies appear along the southwestern coast of Africa as well and is known as the Benguela Nino. The associated reduction in equatorial upwelling leads to a weakening of nutrient supply to the otherwise productive surface waters and to an effective temporal shutdown of the equatorial outgassing of CO_2 into the atmosphere.

Here we review the historical observational record of these phenomena and corresponding changes in meteorology as well as current ideas about their underlying dynamical causes.

1.3.1 Atlantic ENSO

The anomalous warmings of the equatorial and southern Gulf of Guinea occur at approximately two to three year intervals with 14 such warm events having occurred in the four decades since 1963 (\`63, \`66, \`68, \`73, \`74, \`81, \`84, \`87, \`88, \`93, \`96, \`97, \`99, and \`03). The first well-documented event in 1963 occurred during the EQUALANT program. The coincidence of warming sea surface temperatures, a relaxation of the trade winds and shifts in convection during that summer caused Merle (1980) and Hisard (1980) to dub this phenomenon the 'Atlantic Nino'. Further observational results by Servain et al. (1982) made clear the connection between changes in the trade winds and changes in SST. The 1984 event occurred during another observational program, SEQUAL/FOCAL and just after the massive 1982-3 Nino. The extensive array of subsurface observations showed that the warming of the mixed layer occurred in conjunction with an anomalous deepening of the oceanic thermocline in the eastern basin (Philander, 1986), which resulted from an eastward shift of anomalous heat within the equatorial waveguide (Carton and Huang, 1994).

Ruiz-Barradas et al (2000) present an empirical examination of the evolution of the Atlantic Nino based on combined ocean and atmospheric reanalyses. They find that typically during a warm event the excess of warm water originates from either northwestern or southwestern basin in response to relaxations of the trade winds. Associated with the warming SSTs are changes in the overlying atmosphere. The equatorial trade winds relax west of 20^{o}W while further eastward the northward winds associated with the North African summer monsoon also weaken (Horel et al., 1986; Zebiak, 1993). Corresponding increases in diabatic heating in the mid-troposphere occur along with a southward shift of tropical convection (Wagner and da Silva, 1994; Carton et al., 1996; Giannini et al. 2003), causing flooding in the coastal cities of the Gulf of Guinea.

The periodicity of the Atlantic Nino seems to vary considerably from decade to decade. The decade beginning in 1974 had few warm events relative to the surrounding decades of the 1960s and 1980s. The reasons for these changes are still poorly understood. Key parameters such as the heat content of the tropical thermocline have only recently come to be measured regularly, while theoretical attention seems to be focusing on changes in the rates of subduction within the tropical thermocline (see Section 1.5). A second issue of considerable interest is the extent to which the equatorial Atlantic acts in response to ENSO in the tropical Pacific. But, while there is some indication of a phase lagged relationship for particular events, for example, 1984, averaged over multiple decades the relationship is not statistically significant (Dommenget and Latif, 2000).

These anomalous shifts in tropical convection and equatorial winds are generally well captured by atmospheric general circulation models that are forced by Atlantic Nino SST anomalies, confirming that the trade winds respond to changes in SST in ways similar to the El Nino counterpart in the Pacific (Dommenget and Latif, 2000; Wang and Carton, 2003). The similarity of conditions in the tropical Atlantic and Pacific suggests that the Bjerknes feedback mechanism of thermocline change in concert with fluctuations in the trade winds may explain the weaker interannual variability in the Atlantic as well. This idea was explored in an intermediate coupled model by Zebiak (1993) who concluded that the tropical Atlantic was likely marginally stable, but capable of damped oscillations in response to forcing external to the equatorial Atlantic. The strength of the Bjerknes feedback and possible external forcing mechanisms remain areas of active research (e.g. Wu and Liu, 2002).

1.3.2 The Benguela Niños

Benguela Niños are intermittent, acute, extreme warm events near the border between the southward flowing Angola Current and the Benguela upwelling system off southwestern Africa (Shannon et al. 1986). These anomalously warm events have dramatic effects on the fisheries and the climate of the region. They tend to induce significant rainfall anomalies (Rouault et al. 2003) and can drastically modify fish distribution and abundance (Boyer et al. 2001). Benguela Niños occurred in 1934, 1949, 1963, 1984 (Shannon et al. 1986) and more recently in 1995 (Gammelsrød et al. 1998). Such episodes are less frequent and less intense than their Pacific counterparts, and they tend to develop south of equator.

In essence, Benguela Niños express themselves as abnormally and persistent high sea surface temperatures (SST) along the coast of Angola and Namibia. Conversely, Benguela Niñas may be regarded as similar, except that the SST anomalies along the coast are cool (Florenchie *et al.*, 2003b). Smaller warm and cool events along the Angola / Namibian coast occur frequently and may be generated in a similar way to Benguela Niños and Niñas; however, their surface expression is weak due to other factors.

A combination of various observational and model analyses at different depths suggests that, despite their limited surface expression, warm and cold episodes along the coast of Angola and Namibia are in fact large-scale events spreading from the equator at different depths with a duration of several months. Analysis of altimeter, SST and OPA OGCM output (Florenchie *et al.*, 2003ab) indicates that all warm (cold) episodes in the tropical SE Atlantic over the 1992-2000 period tend to be associated with positive (negative) sea level anomalies spreading along the African coast from the equator to as far south as about 20°S **(Fig. 1.3.1).**

The 1995 and 1996 warm events show positive anomalies with respective local maxima of 12 cm and 10 cm while the 1997 cold event shows strong negative anomalies with a local minimum of -11 cm. Calculations from the slope of Hovmoeller plots of sea level anomalies suggest a poleward propagation rate of between 0.5 and 1 m/s **(Fig. 1.3.1)**. Such an estimate agrees with the poleward propagations observed in the eastern Pacific by Enfield and Allen (1980) or simulated (Clarke and Van Gorder, 1994). A coastal trapped wave propagation process is consistent with the spreading of anomalies from the equator southward. However, discrepancies between theoretical phase speeds and the slower observed ones may occur because the theory does not take into account coastal shelf and slope bottom topography or bottom friction (Clarke and Van Gorder, 1994; Pizarro et al., 2001).

Analysis of ERS wind stress and Reynolds SST in the equatorial Atlantic (Florenchie *et al.*, 2003ab) indicates that, about 3 months prior to the appearance of SST anomalies along the Angola coast, the eastern equatorial Atlantic is directly influenced by remote zonal wind stress anomalies **(Fig. 1.3.2).** through equatorial wave dynamics (there is less than a one-month lag between the two signals). As noted by Picaut (1985), equatorial oceans tend to respond clearly and coherently to wind fluctuations as seems to be the case here. Anomalies in the trades in the western to central equatorial basin excite eastward propagating Kelvin waves that depress or lift the thermocline all the way to the African coast and create subsurface temperature anomalies. On reaching the African coast, coastal trapped waves are generated which propagate southward and induce SSTA in the Angola Benguela frontal area (ABA), where the thermocline reaches the surface **(Fig. 1.3.3)**.

The strong correlation between SST anomalies in the ABA and interannual zonal wind anomalies south of the equator over the western and central Atlantic basin suggests a mechanism based on equatorial and coastal trapped waves to explain the equatorial origin of most episodes. SST anomalies become visible at the surface one to two months after the appearance of subsurface temperature anomalies at the thermocline depth. Such anomalies can be attributed to vertical shifts of the thermocline under the action of propagating Kelvin waves initially triggered by zonal wind variations. These waves are deviated poleward on approaching the African continent and temperature anomalies become more or less visible at the surface as a function of various factors like the strength of the event, the depth of the thermocline or the upwelling or downwelling-favourable winds. Temperature anomalies start interacting with the atmosphere when and where the thermocline outcrops along the coast. Seasonal variations of the thermocline depth and shape also modulate the surface expression of the anomaly pool.

Analysis of local heat fluxes (Florenchie *et al.*, 2003b) suggests that the latent heat flux seems to have a rather passive role on the evolution of events in the ABA and mainly acts as a thermostat to regulate cold and warm events at the surface. Local variations in latent heat flux definitively did not create the higher than normal SSTs in the large Benguela Niños of 1984 and 1995. Furthermore, since the local rain anomalies tend to be positive (negative), and cloud cover increased (decreased) during warm (cool) events, changes in solar radiation tend to weaken the events, i.e., they act in concert with the latent and sensible heat fluxes to moderate the surface expression of the events. Local wind-induced upwelling and offshore Ekman transports may have contributed towards producing lower SSTs during the 1992 and 1997 cool events, but, in general, the local wind regime does not seem to play the major role in the expression of Benguela Niños and Niñas.

Despite the relatively rare occurrence of Benguela Niños and Niñas, warm and cold SST anomalies tend to develop regularly off Angola and Namibia. Monthly standard deviations reveal seasonality with a maximum of surface temperature variability in March/April and a minimum in September/October. Major warm events in phase with late summer are likely to give rise to Benguela Niños since they induce extremely high sea temperatures that affect the ecosystem. By interacting with the atmosphere via moisture fluxes, high SSTs may reinforce the rainfall season of southwestern Africa with sudden flooding and devastating consequences.

Sea level anomalies in the eastern equatorial basin show a strong correlation with the southern SSTA signal. The remote forcing of the SST anomalies highlights the possibility of being able to forecast future extreme events via real-time sea level and wind observations or predictive models. The development of equatorial subsurface anomalies could also be detected in advance thanks to local measurements such as the ones performed by the PIRATA array (Servain et al. 1998). However, the non-linear response of SST anomalies in the ABA to the remote wind forcing emphasizes the need for further work to understand the way different mechanisms seem to control the development of each individual event in the tropical Atlantic basin.

b) Links with the West African monsoon

Analysis of NCEP OLR, wind and geopotential height data indicates that the winter intensification of wind-stress off the Angolan coast is linked with convective activity over equatorial West Africa (Risien *et al.*, 2003). Given that some of the moisture feeding into the West African monsoon emanates from the tropical SE Atlantic, better understanding of the teleconnections between monsoonal activity and variability in the heat budget of the eastern South Atlantic is needed. The role of modulations to the South Atlantic anticyclone, which is known to vary substantially on interannual to interdecadal scales (e.g., Venegas *et al.*, 1997; Reason, 2000) as well as respond to ENSO forcing (e.g., Venegas *et al.,* 1999; Reason *et al.,* 2000), in influencing both the SST and upper ocean heat content in the SE Atlantic as well as the moisture flux towards West Africa remains poorly understood.

1.4 Intraseasonal variability

Intraseasonal variability has been reported from all parts of the tropical Atlantic. Typical periods at the western boundary near the equator were about 1.5-2 months (Johns et al., 1990; Schott and Fischer, 1993). This variability is also observed in the interior equatorial belt (Send et al., 2002; Thierry et al., 2003). At the boundaries between reversing zonal currents of the interior and eastern TA Tropical Instability waves (TIWs) are generated (Weisberg and Weingarten, 1988) that may cause undulations of SST fronts and thus can affect the atmosphere.

Recent model studies (Jochum et al. 2003) suggest that the instabilities near the equator can cause relatively large internal variability. This may lead to significant variance of the springtime position of the SST maximum (can be north and south of the equator). This mechanism is still relatively unexplored, as high-resolution ocean models and long integrations are needed, but it might be important for the coupled climate. At least it means that there are limits to predictability due to the chaotic nature of the ocean circulation. That is, nonlinearities in the tropical Atlantic Ocean are important and should not be overlooked.

Plans for TACE fieldwork must therefore assure that energetic intraseasonal changes are properly resolved and included into the analysis.

1.5 Ocean circulation, subtropical cells (STCs) and their role in TAV

1.5.1 Introduction

The subtropical cells (STCs) are shallow overturning circulations confined to the upper 500 m. They connect subduction zones of the eastern, subtropical ocean with upwelling zones in the tropics **(Fig. 1.5.1).** The subsurface STC branches carry thermocline water to the equator either in western boundary currents after circulating across the basin in the Subtropical Gyres or directly in the ocean interior. They are closed by poleward surface currents, largely Ekman transports, that return the upwelled waters to the subtropics (*e.g.*, Malanotte-Rizzoli *et al.*, 2000).

One function of the STCs is to provide the cool subsurface water that is required to maintain the tropical thermocline. For this reason, STC variability has been hypothesized to be important for the decadal modulation of ENSO and for Pacific decadal variability, and it may affect Atlantic equatorial SST as well. STC pathways are complicated by their interaction with the other ocean currents, namely, the northward flow of warm water in the Atlantic by the Meridional Overturning Circulation (MOC) with a transport of about 15 Sv (Ganachaud and Wunsch, 2001; Lumpkin and Speer, 2003). One result of these interactions is that the southern STC is stronger than the northern one (Zhang *et al.*, 2003; Fratantoni *et al.*, 2000; Lazar *et al*, 2002).

The STCs also interact with even shallower overturning cells confined to the tropics, the Tropical Cells (TCs). They are associated with downwelling driven by the decrease of the poleward Ekman transport 4–6° off the equator (Molinari *et al.*, 2003). Interestingly, although the TCs are strong in zonal integrations along constant depths, they are much diminished in integrations carried out along isopycnal layers, indicating that they have little influence on heat transport (Hazeleger *et al.*, 2001). Their existence implies that any measure of STC strength must be defined poleward of the TC convergences, that is, closer to the dynamical intersection between the tropics and subtropics (near 8–10° say).

Finally, the off-equatorial undercurrents have to be considered, the North and South Equatorial Undercurrents (NEUC and SEUC). The fate of these currents is not clear. They diverge poleward toward the east (Arhan *et al.*, 1998; Bourles *et al.*, 2002) and hence may upwell in the aforementioned off-equatorial regions or recirculate in the North or South Equatorial Currents. Although most of their transport is located deeper than the EUC, their shallow portions overlap with the deeper part of the EUC. Thus, they can potentially impact the STCs by partially blocking the equatorward transport of thermocline water to the equator, instead carrying it eastward where it may upwell along the eastern boundary or in domes. Even more fundamentally, they can perhaps be considered part of the STCs themselves, a deep equatorial branch.

1.5.2 Wind stress fields and Ekman transport divergences

As discussed in section 1.1.1, the Atlantic ITCZ shows a marked seasonal migration, moving from its most equatorial position during February to its most northern location near 10°N during August. The Ekman transports are directed poleward in both hemispheres with a zonally integrated, annual-mean divergence between 10°S and 10°N of 26 Sv for the NCEP reanalysis stresses. The time series of the Ekman transports across 10°N and 10°S for 1990-99 from the NCEP and the ERS-1/2 scatterometer stresses as well as their divergence **(Fig. 1.5.2)** shows variations of about 2 Sv amplitude at interannual time scales with some longer-period trends superimposed. There is general agreement among both products, and it suggests that the driving of the variability is much smaller than that of the mean STC.

The Ekman-pumping velocity field, w_{ek}, for the annual-mean winds shows a band of positive (or weakened negative) w_{ek} that extends nearly across the basin in the North Atlantic, limiting the possibility of interior STC pathways in the northern hemisphere. This band shows large seasonal variation in westward extent **(Fig. 1.5.3)**. There is also an area of positive w_{ek} in the eastern tropical-subtropical South Atlantic, but allowing a larger longitude range of interior exchange with the tropics. The annual-mean area integral of positive w_{ek} in the region W, 5 N N plus the line integral of offshore Ekman transport along the coast, is 4.4 Sv, providing an upper limit for off-equatorial upwelling in the Guinea dome and offshore of NW Africa. Similarly, the area integral of positive w_{ek} for W, 20 S S plus the coastal Ekman divergence driven by the alongshore winds yields an upper estimate on coastal and Angola Dome upwelling of 5.5 Sv, yielding a total eastern boundary Ekman divergence for the Atlantic of about 11 Sv, similar to the eastern Pacific and not small in relation to the equatorial 10°N/10°S divergence of 26 Sv.

From recent satellite observations, an ITCZ has also been identified in the western tropical South Atlantic. It extends eastward from the Brazilian coast in latitude range 3–10°S during boreal summer (Grodsky and Carton, 2003), and is associated with wind convergence, high SST, reduced surface salinities, and increased precipitation. Its possible effects on the structure of the southern-hemisphere STC, if any, have not yet been investigated.

1.5.3 Observations of STC pathways and transports

a) Subduction

For the North Atlantic, Qiu and Huang (1995) estimated an annual-mean subduction rate of 27 Sv. By calculating geostrophic pathways of the subducted waters, they further showed that most of the subducted water returns northward within the NEC and Gulf Stream, but they did not specifically estimate the southward transfer that would contribute to the northern STC. For the South Atlantic, Karstensen and Quadfasel (2002) estimated a total subduction of 22.5 Sv south of about 10°S, of which 18.7 Sv was inserted into density classes lighter than 26.8 kg·m^{-3} that can upwell in the eastern tropical Atlantic. Much of the subducted water that is introduced into the southern SEC, however, reaches the western boundary south of the bifurcation latitude (12–15°S; Fig. 3), returning southward within the Brazil Current.

b) Subsurface equatorward flow

Western-boundary pathways

In the South Atlantic, the SEC carries thermocline waters subducted in the southeastern ocean toward the northwest. The bifurcation of the SEC occurs at 12–15°S, and from there the North Brazil Undercurrent (NBUC) transports the bulk of the STC waters equatorward. Based on 7 shipboard current-profiling sections at 5°S, Schott *et al.* (2002) estimated that the NBUC transports 25 Sv northward across that latitude (**Fig. 1.5.4a**). This transport is a superposition of the MOC, the South Atlantic STC, and a recirculation of the southward interior Sverdrup transport, the latter estimated to be about 10 Sv near 5°S (Mayer *et al.*, 1998). A total warmwater transport of 35 Sv (above 1100m) is crossing the equator at 44°W (Schott *et al.*, 1998). The NBC overshoots the equator and, after passing through a retroflection zone known for its intense eddy-shedding activity (Garzoli *et al.*, 2003), most of it merges into various zonal currents in the interior ocean. At 5°S the NBUC transports 12 Sv in density range = 23.4–26.2 kg m^{-3} **(Fig. 1.5.1)** that is evaluated by Zhang *et al.* (2003) for STC circulation (see below).

In the North Atlantic, thermocline water in the NEC is carried equatorward by the Guyana Undercurrent, which bends eastward to join the NECC at 5–8°N and the NEUC at 3–5°N. It is a very weak flow, transporting only about 3 Sv (Wilson *et al.*, 1994; Bourles *et al.*, 1999a; Schott *et al.*, 1998). Zhang *et al.* (2003) estimate a western boundary undercurrent transport of 3.3±1.0 Sv, in agreement with the earlier observational estimates. The weakness of the northern equatorward STC flow compared to the southern one is of course a consequence of the Atlantic MOC, transporting about 15 Sv of warm water across the equator.

Interior pathways

In the interior of the tropical-subtropical Atlantic, the water masses subducted in the eastern subtropics and inserted into the thermocline by Ekman pumping are characterized by a potential-vorticity (PV) minimum. The Ekman upwelling associated with the ITCZ, however, brings stratified waters with higher PV into the density range of the subducted waters, causing them to make a westward detour around this barrier on their way south. This is clearly seen in the PV distribution (f/_h, where _h is layer thickness per density increment _ = 0.2 kg m^{-3}) on the isopycnal surface = 25.4 kg m^{-3}, as determined by Zhang et al. (2003) from climatological hydrographic data **(Fig. 1.5.5a)**. Trajectories of the subducted waters on the = 25.4 kg m^{-3} surface show that there are indeed interior pathways in the South Atlantic, facilitated by the narrowness of the PV barrier there **(Fig. 1.5.5b)**.

For the South Atlantic, Zhang *et al.* (2003) determined the geostrophic transport across 6°S, between the African coast and 34°W for various density classes **(Fig. 1.5.5c)**, obtaining a total of 4.0±0.5 Sv for the density range = 23.5–26.3 kg m^{-3}. For the North Atlantic, they estimated 2 Sv of equatorward interior flow across 10°N **(Fig. 1.5.1)**, concentrated in the density range = 23.5–26.0 kg m^{-3}. A similar calculation was carried out by Lazar *et al.* (2002), also yielding interior pathways. It should be noted that the lightest portions of the equatorward flow (< 24.0 kg m^{-3}) are subducted at quite low latitudes and therefore do not really qualify as part of the STC, because they are not participating in a subtropical-tropical exchange.

c) Equatorial currents

Equatorial Undercurrent

The EUC draws most of its water from the NBC retroflection **(Fig. 1.5.1).** Schott *et al.* (2002) reported that the undercurrent layer (= 24.5–26.8 kg m^{-3}) is supplied by a northward flow of 13.4±2.7 Sv at 5° S, and estimated that about 80% of this flow enters the EUC. The EUC has a total eastward transport of 21.9±3.5 Sv in the mean section at 35°W, of which 8.6 Sv occur in the near-surface layer above = 24.5 kg m^{-3} (Schott *et al.*, 2003). The surface-layer eastward flow occurs predominantly during the spring when the ITCZ migrates close to the equator **(Fig. 1.1.1b),** and weakened easterly, or even westerly, wind stresses drive the near-surface flow (Bourles *et al.*, 1999b; Schott *et al.*, 1998).

Farther east, at 0°E, the transport is reduced to about 6 Sv and the bulk of the EUC transport lies below = 26.25 kg m^{-3} (Bourles *et al.*, 2002). These authors showed that during boreal summer the Atlantic EUC can even surface and terminate away from the boundary. The fate of the mean EUC at the eastern boundary and its possible supply of the Gabon-Congo Undercurrent or Angola Dome is still uncertain (Stramma and Schott, 1999; Stramma et al., 2003).

North and South Equatorial Undercurrents

The NEUC and SEUC seem to be weak in the west and to strengthen into the central ocean. Their potential roles in the STC are, with the transports carried within their shallower density layers, to supply partial inflow into the off-equatorial eastern upwelling regimes along the coasts and in the Guinea and Angola domes. The SEUC is recognizable **(Fig. 1.5.4b)** at about 100–400 m depth at 3–4°S, with its density range 26.2–27.0 kg m^{-3}; its transport at 35°W is only 3 Sv, but at 23°W it has strengthened to about 10 Sv **(Fig. 1.5.4c).** Although a supply of the SEUC out of the NBC retroflection regime has not been established from ship surveys, the SEUC is distinguishable from the low-oxygen interior thermocline waters of the tropical South Atlantic by an oxygen relative maximum (Arhan *et al.*, 1998). An interpretation is that it is supplied by a mixture of NBUC waters with interior SEC recirculations. Mean transports from sections further east have still to be composed from the sparse data base.

Since the equatorward flow of subducted water in the interior ocean occurs in the density range σ_θ = 23.5–26.3 kg m^{-3} (Zhang *et al.*, 2003), most of it can pass over the SEUC in the western basin **(Fig. 1.5.4a)** but not at 23°W in the central basin where isopycnals σ_θ = 25.4 kg m^{-3} and deeper pass through its upper part **(Fig. 1.5.4c).** Thus, the denser part of the subducted water **(Fig. 1.5.5c)** cannot reach the EUC. D. Zhang (pers. comm., 2003) estimated that about 1 Sv of the interior equatorward thermocline flow is trapped by the SEUC in this way and carried eastward with it.

The NEUC is not as clearly identifiable as an isolated current core as the SEUC is, because it is not clearly separated from the NECC in boreal summer and annual mean. Ist supply is partially out of the NBC retroflection region (Schott et al., 1998) and partially out of the NEC recirculation (Bourles et al., 1999a). In the east it presumably supplies the Guinea Dome and related upwelling and thus can be part of the STC circulation.

Tropical cells

From their ADCP section analysis, Molinari *et al.* (2003) inferred the existence of a North Atlantic TC, consisting of shallow upwelling at the equator and downwelling from 3–6°N. For the downwelling branch, Grodsky and Carton (2002) estimated a transport of 4±2 Sv for 35°W–10°E from drifter convergences. The existence of a TC in the tropical South has not yet been been quantified. In any event, the presence of Atlantic TCs has to be taken into account when evaluating the wind forcing and net meridional upper-layer STC transports, i.e. the relevant quantifications need to be carried out poleward of about 6° latitude.

As mentioned above, the subduction and equatorward recirculation of the TCs is mostly limited to the surface-mixed layer. Therefore, they play only a small role in isopycnic models and their climate relevance is considered small.

e) Upwelling

Source depth of upwelled waters

The core of the Atlantic EUC lies in density range σ_θ = 24.5–26.8 kg m^{-3}, and the SEUC and NEUC extend to densities of the order σ_θ = 27.0 kg m^{-3}. Since the σ_θ = 26.25 kg m^{-3} isopycnal within the EUC reaches the surface at the Greenwich meridian (Bourles *et al.*, 2002), only densities near and less than this value can upwell to the surface at the equator (Snowden and Molinari, 2003). The fate of the denser water is unclear, but it may upwell in the off-equatorial areas.

Equatorial upwelling

From an evaluation of early direct current sections across the equator, Gouriou and Reverdin (1992) estimated a divergence of 15 Sv for the 4–35°W band, and concluded that upwelling into the surface layer was confined to the upper part of the EUC. From an average of 12 cross-equatorial western Atlantic ADCP sections extending for 10° of longitude centered near 35°W, Molinari *et al.* (2003) estimated an upwelling transport of about 11 Sv. These estimates are comparable to upwelling transports obtained from inverse-model analyses of the divergence across off-equatorial sections. Lux *et al.* (2001), for example, derived an upwelling transport of 7.5 Sv out of the thermocline water layer (σ_θ = 24.58–26.75 kg m^{-3}) for a box closed by zonal sections at 7.5°N and 4.5°S. Earlier, Roemmich (1983) had obtained an upwelling transport of 6–10 Sv across σ_θ = 26.2 kg m^{-3} for basin-wide sections at 10°N and 10°S, the spread of values depending on model assumptions. When comparing these estimates with the Ekman divergence of about 25 Sv across 10°N/10°S it has to be considered that diapycnal upwelling velocities have a sharp profile and maximum upwelling values may not be reached when averaging over density ranges.

Off-equatorial upwelling

Off-equatorial upwelling occurs along the eastern coasts and in two cyclonic domes, namely, the Guinea and Angola domes in the northern and southern hemispheres, respectively **(Fig. 1.5.3),** which are driven by the regions of strong positive w_{ek} in the eastern tropical ocean. Upwelling estimates based on observations have not been reported for these regions, but, as noted above, their combined Ekman upwelling effect (11 Sv) is not negligible. Their effects, while not known in detail, are nevertheless included in aforementioned inverse studies for those contributions that fall inside the northern and southern boundaries of their respective analysis domains.

1.5.4 Models

a) Mean circulation

Numerical model studies on circulation in the tropical Atlantic started with studies of Philander and Pacanowski (1984). They showed the strong asymmetry in the sources of thermocline waters of the equatorial Atlantic. There is a large input from the South Atlantic, while the input from the North Atlantic is modest. A more detailed picture of ocean currents and pathways is provided by Schott and Boening (1991). They used a high resolution ocean model to show details in the currents and seasonal changes in the overturning circulation. A predominant source via western boundary currents is shown as was suggested by the boundary current meter measurements.

A number of recent observational (see section 1.5.3) and modeling studies (**Fig. 1.5.6)** show that the tropical Atlantic water originates predominantly from the Southern Hemisphere. It has been shown that the presence of the basin-wide meridional overturning circulation in the Atlantic associated with NADW formation induces the asymmetry (e.g. Fratantoni et al.; 2000; see section 1.5.6). Another source of asymmetry, as discussed above, is the ITCZ which is north of the equator. The associated upwelling generates a barrier of potential vorticity in the subsurface ocean and blocks from the north to the south.

The detailed pathways from the subduction sites in the extratropics and tropics towards the tropical thermocline have been studied in a number of ocean models. Using coarse resolution models pathways from the South Atlantic to the equator using Eulerian diagnostics are shown by Malanotte-Rizzoli et al. (2000), Lazar et al. (2002) and Inui et al. (2002). Lagrangian diagnostics have been used by Blanke et al. (1999) to study the pathways. Using the same Lagrangian diagnostics and output from an eddy-permitting model Hazeleger et al. (2003) show that for a detailed, quantitative picture on the sources of the tropical Atlantic Equatorial Undercurrent it is important to take high-frequency variations into account. High-frequency ocean motions cause dispersion of water masses while they transfer from their subduction sites to the tropics. Also, there is a profound effect of the high-frequency variations on the overturning circulation.

The so-called tropical cells, which consist of upwelling on the equator and downwelling about 5 degrees poleward of the equator are completely compensated by eddy motions. That is, they do not exist when the residual mean flow is considered. Since it is the residual mean flow that carries active and passive tracers, these cells are not important for the heat transport. A similar conclusion was drawn by Jochum et al. (2003).

The picture that emerges from all these studies is that water masses are carried from the main subduction sites at about 20 S along the South Equatorial Current towards the North Brazil current at the western boundary which feeds the tropical thermocline **(Fig. 1.5.6).** This is the lower branch of the so-called Subtropical Cell. A pathway through the interior of the ocean basin is also possible, but it transfers far less water than along the western boundary, despite the homogeneous potential vorticity structure south of the equator. Reasons for this are the strong zonal character of the currents in the tropics and the strong recirculations in the ocean which are not revealed in Eulerian diagnostics. From the north water masses subduct along the North Equatorial Current and transfer to the western boundary where they feed into the North Equatorial Counter Current. Via a complex route the water masses make their way to the equator. The ratio of water coming from the north to the south is about 1:8.

b) Sources and fate of the EUC

In the tropics, most attention has gone to the Equatorial Undercurrent. As the EUC is at thermocline level, the sources and fate of the EUC are of interest (see Hazeleger et al. 2003, and Hazeleger and de Vries, 2003; **Fig. 1.5.6**). Also, it is the EUC water that upwells on the equator. Just off the equator the South Equatorial Undercurrent and North Equatorial Undercurrent are the Atlantic equivalents of Tsuchia jets in the Pacific. Especially the SEUC and NEUC in ocean models depend on the mixing parameterizations being used (Schott and Boening 1991). The role of the NEUC and SEUC in ventilation of the tropical thermocline is still unclear.

Once at the equator, thermocline water upwells. At the surface it is expelled poleward at the surface by the Ekman drift forced by the zonal wind stresses. Upwelling of thermocline water also occurs at the eastern boundaries of the basin (Hazeleger et al. 2003b). After upwelling, water masses subduct mainly in the tropical basin in the Southern Hemisphere to become part of the STC. Some fraction of the water masses does not return to the tropics while recirculating in the STC but joins the basinwide overturning to become part of the conveyor belt.

c) Dependence of STC exchange windows upon wind fields

A variety of model studies have addressed the dependence of Atlantic STC pathways and exchange windows upon the wind fields. Malanotte-Rizzoli *et al.* (2000) used an intermediate-resolution (non-eddy resolving) OGCM, driven by COADS climatology and obtained PV distributions that looked qualitatively similar to those observed. Regarding the exchange windows, they find an interior exchange zone in the northern hemisphere, whereas all the waters subducted in the eastern South Atlantic take the western-boundary pathway. As expected, the seasonal effect of the northern PV barrier is most pronounced in late boreal summer and much reduced in winter, so that the annual mean is dominated by the summer situation. They attributed the stronger than observed equatorward northern thermocline flow in their simulation to their model's having a somewhat weak MOC.

It is obvious that there must be a dependence of the exchange windows on the patterns and intensity of the wind stress climatology, since that determines upwelling and the PV barrier. Inui *et al.* (2002) studied the differences in STC pathways between the commonly used Hellermann and Rosenstein (1983) and da Silva *et al.* (1994) forcing fields. They found that for the stronger Hellerman and Rosenstein (1983) forcing the interior exchange window is much reduced in comparison to the weaker da Silva *et al.* (1994) forcing. This calculation was confirmed by a similar model study with both wind stress climatologies carried out by Lazar *et al.* (2002).

d) Effect of the MOC

Several investigators have studied the interaction of the Atlantic MOC with the STC. Fratantoni *et al.* (2000) compared two solutions to a 6-layer isopycnal model of the tropical Atlantic, one forced by wind alone and the other by winds and a 15-Sv MOC imposed at the model's northern and southern boundaries. A similar calculation was carried out by Boening (pers. comm., 2002) using an Atlantic-basin GCM. Both studies found that the STCs were nearly symmetric about the equator in the solution forced only by winds. In contrast, for the solution forced by wind and the MOC, the northern cell was so weak that there was only about 2 Sv of equatorward thermocline flow in the northern hemisphere, similar to the observational evidence presented above.

The transports of solutions forced by both winds and the MOC are approximately a linear superposition of the transports from solutions forced by MOC and wind forcing alone. This linearity, however, is not at all true for the mesoscale variability. The eddy kinetic energy at the western boundary is greatly enhanced with the addition of the MOC. Fratantoni *et al.* (2000) attribute this to the combined effects of increased current shear, advection of potential vorticity from the equatorial waveguide by the strengthened NBC and enhanced potential vorticity gradient. In their model, NBC eddy shedding only occurs in the combined forcing case.

1.5.5 Variability

a) Observations or lack thereof

Different aspects of the Atlantic STCs exhibit interannual-to-decadal variability. As documented in previous sections, equatorial SSTs and sealevels vary at interannual periods, corresponding to Atlantic ENSOs and Benguela Ninos **(Figs. 1.1.4, 1.3.2)** , but also at decadal periods **(Fig. 1.2.1)**. The Ekman divergence between 10°N and 10°S has amplitudes of several Sverdrups **(Fig. 1.5.2)**, and temperature variability at thermocline levels has been documented. The transfer of South Atlantic thermocline waters by North Brazil Current rings also undergoes longer period variations (Goni and Johns, 2003), which likely has consequences for the distribution of water masses and STC pathways. To date, though, there is no Atlantic counterpart to the McPhaden and Zhang (2002) calculation for the Pacific that points toward these individual variations being collectively associated with STC variability.

b) Models

The ocean’s role in low-frequency variability has been discussed mainly in terms of variations in the STCs. Basically, two mechanisms are proposed. Firstly, anomalous temperature anomalies generated at the surface in the extratropics may subduct and advect to the tropical thermocline following a subsurface pathway in which diffusion is small (the so called mechanism, first proposed by Gu and Philander, 1997). Secondly, the strength of the STCs may vary and cause a change in the tropical thermocline (the mechanism, Kleeman et al. 1999).

Both mechanisms have been extensively studied in the Pacific (e.g. Schneider et al. 1999, Hazeleger et al. 2001a). The advection of thermal anomalies to the equator on interannual to decadal time scales does not seem to work. The ocean circulation is too dispersive. It is expected that this even increases when eddy resolving models are used. In the South Atlantic, propagation of thermal anomalous around the subtropical gyre has been shown (Lazar et al. 2002), but the anomalies cannot be traced to the equator. At long, interdecadal time scales, the ventilation of the thermocline is important as anomalous water masses originating from the extratropics will slowly fill up the thermocline, but one needs persistent surface heat flux anomalies for that (for the Pacific this can be seen in Hazeleger et al. 2001b). Variations in the strength of the STCs and their impact on the thermocline has been hardly studied. Scaling analysis shows that it is easier to change tropical thermocline by "pull" from the tropics than "push" from the extratropics.

1.6 Potential MOC Effects on Tropical Atlantic SST

The Atlantic meridional overturning circulation (MOC) plays an important role in the global climate system through its large transport of heat northward across the equator. Significant changes in the strength of the MOC would cause widespread climate changes particularly over the North Atlantic (e.g., Manabe and Stouffer, 1995). Recently there have been several studies focused on the impact of variations in the Atlantic MOC on longer term variations in tropical Atlantic SST, using both uncoupled ocean models (Yang, 1999; Johnson and Marshall, 2002) and coupled models (Dong and Sutton, 2003). These studies have shown that an important aspect of the ocean's dynamical response to a change in the MOC is the so-called "equatorial buffer" (Kawase, 1987; Johnson and Marshall, 2002), which limits the rapid communication of MOC signals across the equator between the two hemispheres.

The initial response to change in deep water production in the North Atlantic consists of an equatorward-propagating baroclinic Kelvin (or more generally, coastally trapped) wave along the western boundary. Upon reaching the equator it is transmitted to the eastern boundary via an equatorial Kelvin wave and poleward in the basins by coastal Kelvin waves, and then westward into the ocean interior by long Rossby waves. As a result of this process, a rapid adjustment of the MOC occurs throughout the northern hemisphere ocean, via the Kelvin wave response (on the order of a few months), while the response in the southern hemisphere is much slower, set by the Rossby wave time scale (of order several years). Consequently, there is a "mismatch" in the strength of the MOC in the two hemispheres for a number of years following a significant change in the MOC, and a corresponding convergence or divergence of heat in the equatorial region during this period. The feedback of this process to the atmosphere can lead to coupled changes in SST and atmospheric circulation patterns and an amplification of the response.

Yang (1999) suggested a relation between changes in Labrador Sea Water thickness in the subpolar North Atlantic between 1950-1990 and the tropical Atlantic SST "dipole" index with a lag of 5 years. The SST anomaly was concentrated along the western boundary and was shown by a model simulation to be consistent with the pattern produced by anomalous meridional heat advection following a change in the MOC strength. Dong and Sutton (2003) consider the coupled problem and show that a significant reduction of the MOC introduced by freshwater input to the subpolar North Atlantic leads to important SST changes in the tropical Atlantic. The response in the northern hemisphere is rapid, and significant upper ocean circulation anomalies appear in the tropics after several months **(Fig 1.6.1).** Initially the MOC imbalance between the hemispheres results in a broad cooling over the North Atlantic and warming over the tropical South Atlantic. Corresponding changes in the surface atmospheric pressure fields lead to strengthening of high pressure over the subtropical North Atlantic and lowering of pressure over the tropical South Atlantic, which amplifies the response through air-sea fluxes and leads to the development of a dipole SST anomaly in the tropics after about 6 years **(Fig 1.6.2).**

The anomaly in cross-equatorial SST (cooler water north of the equator and warmer water south of the equator) in turn leads to a southward shift of the ITCZ and an associated precipitation anomalies over the tropics **(Fig. 1.6.3).** The SST and precipitation anomaly patterns that develop under such a scenario are quite similar to patterns that have been associated with other interannual forcing mechanisms, such as NAO variability or ENSO-related atmospheric teleconnections. The conclusion to be drawn from these studies is that sufficiently large changes in the strength of the MOC can have important consequences for tropical Atlantic SST. The equatorial region, though remote from the probable forcing regions of MOC variability, is in fact a focal point for oceanic heat storage changes as a consequence of the delayed adjustment of the MOC throughout the basin.

Other effects of a change in the MOC on tropical circulation patterns can be more subtle but still potentially important in oceanic feedbacks to the atmosphere. For example, it is likely that a change in the MOC would substantially impact the structure of the shallow overturning cells that link the tropics and subtropics – the "STCs" (see section 1.5). The present pattern, in which the southern cell of the Atlantic STC is dominant over the northern cell, is believed to be a direct result of the MOC, which cuts off most of the supply of thermocline waters to the equator from the northern subtropics. A decrease in the MOC would lead to a greater symmetry of the cells and an increase in northern hemisphere waters supplied to the EUC that feed equatorial upwelling. Conversely, an increase in the MOC would likely shut down the northern cell altogether and force a redistribution of its upwelling branch to areas farther north of the equator.

Finally, it is possible that changes in the vertical structure of the MOC, independent of any change in the MOC strength itself, could lead to important changes in the stratification and heat storage patterns in the tropics that could have eventual impacts on SST. It is presently believed that about half of the upper limb of the MOC is supplied by thermocline waters from the Indian Ocean and that the other half is supplied by subpolar mode waters originating in the Pacific. For example, Drijfhout et al., (2003), using the Lagrangian trajectory method to backtrace particles from the Atlantic Equator to Drake Passage, Indonesian Throughflow and a section south of Tasmania, found that particles crossing the Atlantic equator and originating from those sections all arrive at the western boundary below 100 m.

However, this distribution could be highly sensitive to climatic anomalies, including changes in the zonal winds over the southern ocean. Increased flow of the warm branch of the upper MOC from the Indian Ocean would likely result in important stratification changes in the southeastern South Atlantic, which is the source region for the southern hemisphere STC waters that are supplied to the equator. An increased strength of this branch and a more rapid advection time of these waters to the tropics could lead to changes in the properties of the upwelled waters to the equator as well as their zonal distribution along the equator.

2. Dedicated model studies of the Tropical Atlantic

2.1 Introduction

In the last decade much progress has been made in understanding of the tropical Atlantic circulation by use of numerical models. Model studies of the atmospheric circulation revealed the impact of sea surface temperature variations on the atmosphere (e.g. Dommenget and Latif 2000, Saravanan and Chang, 2000, Sutton et al 2000, Giannini et al 2003 a.o.), they showed how remote regions are connected to tropical Atlantic variability and they gave insight in mechanisms of the variability in the tropical Atlantic. Ocean model studies showed details on the oceanic circulation, revealed sources of tropical Atlantic water masses (e.g. Malanotte-Rizzoli et al 2000; Hazeleger et al. 2003; Jochum et al 2003) and showed how atmospheric variability plays a role in generating oceanic variability (e.g. Carton et al. 1996, Seager et al. 2001). Hybrid coupled model studies have been performed to gain insight in coupled ocean-atmosphere variability (e.g. Chang et al. 1997). Studies with fully coupled ocean-atmosphere models are sparse, the reason being that coupled models, if not flux corrected fail to simulate basic features of the tropical Atlantic circulation (Davey et al. 2001, **Figs. 1.0.1., 2.1**).

Most attention of the modellers has gone to explanation of sea surface temperature (SST) variability in the tropical Atlantic. The was motivation by the relation between rainfall in the Sahel and in northeast Brazil to SST in the tropical Atlantic (Nobre and Shukla 1991). Variations in the ITCZ have a huge impact on the people living in West Africa and northeastern South America. Furthermore, tropical Atlantic SST may affect extratropical variability. Evidence has been found that the North Atlantic Oscillation is affected by tropical Atlantic SST (Okumura et al. 2001, Terray and Cassou 2002).

As discussed in detail in previous sections, the tropical Atlantic exhibits a strong seasonal cycle on top of which interannual and decadal variability has been found. Two main modes of variability are identified. That is, the interhemispheric meridional gradient mode **(Fig. 1.2.1)**. and the equivalent of El Nino in the eastern cold tongue of the equatorial Atlantic (**Figs. 1.1.4,** e.g. Ruiz-Barradas et al., 2000). The interhemispheric mode is most pronounced during boreal spring while the eastern equatorial cold tongue variability is strongest in boreal summer and fall. The processes that determine these modes range from local thermodynamic feedbacks, radiation feedbacks, dynamical ocean feedbacks to remote influences (see previous chapter).

The relation between the tropical Atlantic and extratropical climate can go two ways. Tropical Atlantic SST may affect the NAO, while there is also an impact of ENSO, NAO and South Atlantic variability on the tropical Atlantic. Modelling results have helped to identify mechanisms of these tropics-extratropics teleconnections.

In all previous sections it was shown that numerical models provided usefull tools to determine relevant processes for tropcial Atlantic variability. Here we will focus on recommendations for studies after atmosphere and upper ocean processes. Since the ocean can provide the long term memory in the tropical climate system and SST can be used as predictor for rainfall.

2.2 Challenges ahead

A prerequisite for a successful dynamic prediction is the use of an air-sea-land coupled model that is capable of realistic simulations. However, nearly all coupled models show poor results for the tropical Atlantic. It is essential to remove biasses in the fully coupled models. Chief among these biases are the failure to keep the mean ITCZ north of the equator and to maintain the equatorial cold tongue. In these models, the ITCZ moves back and forth across the equator following the sun, and stays far too long south of the equator. The zonal SST gradient on the equator is opposite to observations in many models, with higher SSTs in the Gulf of Guinea than east of South America. Peculiarly, this reversal of SST gradient occurs despite prevailing easterly winds on the equator in some models (**Fig. 1.0.1,** Davey et al. 2002). The northward-displaced ITCZ and the equatorial cold tongue are key features of tropical Atlantic climate that are conceivably important for the space-time structure of observed TAV. It is a high priority to reduce and remove these biases in climate models.

The attention of coupled modelers has mainly gone to the tropical Pacific and models have been tuned to represent that as good as possible. As a result the tropical Atlantic is poorly simulated. It has been hypothesized that vertical mixing in the ocean is important for this east west gradient in the Atlantic. In a coupled model study of he tropical Atlantic (with SPEEDO: Micom-Atlantic-1 degree coupled to Speedy T30L7 model, Hazeleger et al. 2003c), it was indeed found that relatively small changes in the parameterization of the oceanic mixed layer has a profound impact on the SST gradient on the equator (Haarsma and Hazeleger, in preparation, see Figure 4). Also, it appears that with careful tuning of the vertical mixing and radiation parameters in the model, the tropical Atlantic mean state could be well represented. This work shows that there is scope for improvement of simulations of Atlantic climate in coupled models.

2.1.1 Technical improvements to improve simulation of tropical Atlantic climate

1) Improvement of coupled models with emphasis on the tropical Atlantic in order to remove biasses in the mean climate and annual cycle. Specific attention should go out to vertical mixing parameterizations (mixed layer and deep ocean diapycnic mixing), atmospheric boundary layer processes, and cloud-radiation interaction (maritime stratus and deep convection over continents).
2) Bridge the gap between ocean models used in "ocean-studies" and ocean models used in "coupled-studies". The main difference between these models is the resolution. The success of linear models in the tropical Pacific and the lack of computer power led to the use of coarse resolution ocean models. However, the elaborate eddy-resolving ocean models start to simulate current structures as found in observations and show that nonlinearities and chaotic behavior is important. Therefore, there should be an effort in coupling these models to atmosphere models.

2.1.2 Dedicated ocean model studies for TACE

Before observation period:

1) Dedicated design study to estimate requirements for observations in eddy resolving ocean models. For example to estimate rms-errors of different oceanic variables derived from floats in order to determine the amount of floats needed and the location of deployment.

During and after observation period:

2) Assimilate the obtained observations (hydrography, floats, fluxes) into ocean models which are forced with surface fluxes over the observation period to improve the simulated circulation and determine a detailed SST budget. Identify the different processes at work (advection, lateral and vertical mixing and surface fluxes), their seasonal and interannual variability. The observations will give time series at fixed locations and along trajectories of floats. This model output will put the observations in a large scale context.

2.1.3 Dedicated coupled model studies for TACE

A prerequisite of process studies is that the models simulate the gross features of tropical Atlantic climate credibly (SST, thermocline depth, radiation, clouds, rainfall, wind stress, and surface heat fluxes).

1) *Studies after eastern cold tongue variability and eastern boundary upwelling areas*. As these areas show a role of ocean dynamics on SST varriability and the rainfall responds to the SST in these areas, there is a potential for predictability beyond NWP time scales. Proposed coupled studies are: (a) detailed studies after sources, pathways and fate of water masses in the eastern cold tongue (b) studies after equatorial trapped waves and their impact the atmospheric circulation (c) studies after feed back processes at play to clarify wether a Bjerkness feed back (marginally stable or unstable) and/or shortwave radiation feedback is operating and (d) study the impact from remote regions by coupling in the Atlantic only compared to a globally coupled model. A hierarchy of ocean models can be used in these studies to investigate the effect of different processes. Such studies should also include the role of biogeochemical feedbacks, e.g. involving biotically affected changes in the absorption profiles of solar radiation,

2) *The role of extratropical-tropical and tropical-tropical climate connections.* The meridional gradient mode and associated ITCZ variability is strongly influenced by climate in remote regions (ENSO, NAO, Indian Ocean, South Atlantic). Also, tropical Atlantic variability could have an impact on extratropical climate (e.g. the NAO). In general, the role of convection over land, associated diabatic heating, wave propagation and the impact on meridonal and zonal cells should get specific attention. Recommended studies are: (a) as mentioned in (1), clever coupling strategies can separate remote impact from local variability in the tropical Atlantic, (b) multi-model ensembles should be used to look for robust features of the impact of remote regions on the tropical Atlantic (c) dedicated coupled model studies should be used to identify the impact of the Subtropical Cell variability on the tropical Atlantic by the v'Tbar or vbarT' mechanism. Partial coupling experiments and the use of a hierarchy of ocean models could be used to achieve this.

3) *Exploration of new perspectives.* Processes that have not yet been considered very well but that might play a role include interaction of tropical Atlantic circulation with the basin-wide MOC, biogeochemical feedbacks and effects of dust. Studies should be made to investigate the strength of these feedbacks.

4) *Climate change*. Experiments with climate change scenarios to study possible changes in tropical Atlantic variability.

2.1.4 Studies needed for models

1) Reanalysis data of the atmosphere helped enormously in the research of tropical climate. However, the quality of reanalysis data in the tropics is not sufficient. Attention should go out at improving this. A climate reanalysis, focussed on climate and less on NWP time scales should be considered. This includes ocean reanalysis products that can be of great use in validating models and studying variability.

2) As vertical mixing/upwelling are key processes in the tropical Atlantic ocean and badly represented in ocean models, a process study including detailed observations (hydrography, microstructure measurements and surface fluxes) with regard to these processes will be valuable for model improvement. The aim of such an observational study should be to determine the heat budget in the upper ocean and its seasonal variability.

3. Observations required for TACE

The questions to be addressed with TACE are based on the results presented and conclusions drawn in section 1 as well as on the deliberations in a number of Tropical Atlantic workshops held during the past four years (Miami, Venice, Paris and Kiel) and including those put forward in the French EGEE project (see section 5).

The overall objective is determination of improved predictability that can be accomplished by a better understanding of the ocean mechanisms involved. This objective requires the following observational studies:

1) *Importance of the surface and subsurface oceanic circulation for SST variability*
The role of advection in determining SST is still poorly understood, but has been found important in the cold tongue region at the intraseasonal to interannual time scale. This task requires box budget studies in key regions to delineate the respective contributions of horizontal advection vs. upwelling, mixing and fluxes by intraseasonal variability in the eastern tropical Atlantic and in the ITCZ regime. An essential component for this endeavor will be time series of advection of surface- and thermocline currents along several longitudes, which will be complemented by water mass and circulation anomaly observations based on drifters, isopycnic and profiling floats. Also to be included are mixing studies in the shallow cold tongue for improved parametrization and model representativeness in the region

2) *Role of shallow overturning variability of the STCs in affecting upwelling and SST*
As discussed in previous sections (e.g., section 1.5.5), it is the variability of wind forcing, i.e. of the Ekman transport divergence across the tropical-subtropical boundary (the process) rather than the equatorward advection of temperature anomalies subducted in the subtropics (the process) that determines the interaction of the tropics with the subtropics. Pathways and variability of the STC should observed during TACE by Lagrangian studies and by extending the observational program to the west and poleward from the immediate study region, for example by observing the variability of equatorward thermocline water transports.

3) *Eastern coastal and dome upwelling and role of off-equatorial eastward currents*
The combined Ekman upwelling along the eastern boundary and in the Guinea and Angola domes is estimated to contribute a total of about 10 Sv (section 1.5.3) to the total upwelling of the equatorial zone. Very little is known about how the relevant branches of the mean circulation, in particular the EUC and the North and South Equatorial Undercurrents, supply this upwelling. More observations are needed to explore how remote causes may affect eastern upwelling and local climates.

4) *Role of wave dynamics*
The interannual variability in the eastern tropical Atlantic has been compared with the Pacific ENSO as regards the role of signal propagation by equatorial waves. Equatorial Kelvin waves generate coastal trapped waves that propagate southward and cause SST anomalies as far south as 20S (the Benguela Ninos; section 1.3.2). Intensified equatorial observations may therefore help, as proven by the Pacific TAO array, to explain and forecast such events.

5) *Improved surface flux fields*
   Routine observations of SST in regions such as the ITCZ that are frequently under cloud cover suffer from biases, in particular as regards representation of intraseasonal variability. One requirement therefore are moored (Atlas type) stations for flux calculations and calibration studies for improved applicability of satellite SST measurements in the eastern and ITCZ region

6) *Role of salinity at the eastern boundary and barrier layers*
   It needs to be studied to which degree low salinities due to precipitation at the easternmost part of the GG and due to river discharges farther south (mainly Congo-Zaïre) influence the SST, and possibly the surface currents, via the phenomenon of the "barrier layer"; can they contribute to a positive " feedback " during years characterized by strong positive anomalies of SST?

7) *CO_2-uptake and storage changes*
   TACE is mostly oriented by physical questions, but the CO_2- problem can be addressed in conjunction with planned observations. Of particular interest is the sensitivity of TA air-sea CO_2 exchange to locally and remotely forced changes in upwelling intensity and source water properties. Observational requirements include pCO_2 sensors on surface flux moorings. Box budget studies of nutrient and carbon inventories will further help to identify pathways and fate of the respective biogeochemical elements.

8) *Satellite observations*
   Altimetry (combined with GRACE referencing) will be essential for continued monitoring of anomaly propagation in the TACE region. SST and, if accuracy increases sufficiently, also salinity will be required, as is ocean colour for observational guidance and larger-scale pattern studies.

4. Field Experiment proposed in conjunction with AMMA

The elements for a tropical Atlantic climate experiment that were to be closely associated with AMMA were discussed in a Workshop in Miami (March 2003; Molinari et al., *CLIVAR Exchanges* 2003) and are summarized below.

4.1 Objectives

The objectives of the proposed field work are developed on the basis of the Observing Period framework defined by AMMA, and are building on the existing French observational program EGEE (section 5). The three different AMMA periods and their objectives are:

4.1.1 During the Long-term Observing Period

To characterize the mechanisms by which anomalous structures (currents and stratification) in the eastern tropical Atlantic (i.e., the equatorial cold tongue, the ITCZ and north of the ITCZ) affect sea-surface temperature variability.

4.1.2 During the Enhanced Observing Period

To characterize the annual cycle of air-sea interaction, in particular the interactions and feedbacks between the ITCZ and the regional (i.e., north as well as south of the ITCZ) SST field.

4.1.3 During the Special Observing Period

To estimate the terms in the heat budget including mixing at the base of the mixed layer and surface fluxes in the Atlantic cold tongue and under and north of the ITCZ during different phases of the monsoon (i.e., onset, peak and late).

4.2 Methodology

The following ingredients of a field program with shipboard or moored observations and associated remote sensing are required for the three observing periods

4.2.1 Long Term Observing Period

Sustained observations including:

- **PIRATA network**
  The PIRATA stations (blue in Fig. 4.1) form the backbone of the long-term observing period and all efforts should be made to extend them throughout the AMMA and TACE observational period.
- **Subsurface current and temperature/salinity moorings**
  Mooring sections are proposed along 23 W, 10W and 6E, to be maintained for several years during the LOP, for studying the role of advection in SST variability. They are to be equipped with ADCPs at the top to measure the shallow flows as well as with current meters and T/S recorders at sufficient vertical resolution below.
- **Surface drifters** (partially funded)
  Pursue pathways of upwelled waters from equatorial and eastern upwelling to subduction regions, ground truthing for satellite observations, in particular under cloud cover; Carton/Garzoli are funded to deploy drifters along cross-equatorial meridional lines (Fig. 4.1)
- **Lagrangian floats** (partially funded)
  Pursue STC thermocline pathways from subtropical subduction zones via zonal equatorial undercurrents to upwelling regions (see, e.g., Fig. 1.5.1) along density surfaces, using ispopycnic RAFOS floats; German RAFOS deployments at shallow isopycnals (IfM Kiel) are beginning in 2004;
- **VOS** (funded)
  XBT transects across the TA have been over the past decades, and will continue to be, an essential element for determining heat storage variability and thermocline signal propagation; they are operated as part of the NOAA/OGP program and also by IRD. In addition, surface salinity observations are carried along some lines, yielding essential information for satellite calibration and mixed layer modeling.
- **Profiling floats** (partially funded);
  As part of the international ARGO program, APEX and PROVOR (see section 5.1.2) floats are being deployed during 2002-4 (**Fig. 4.2;** http://www.coriolis.eu.org) and will be operational during the proposed TACE period; floats are either drifting at 1900 m

(PROVOR) or 1000m (typical ARGO depth), or in the IfM Kiel deployments at shallower STC levels, and profile to 2000m at preselected time intervals (typically 10 days); the frequent profiles from the tropical and subtropical Atlantic over a period of 4-5 years will allow tracing of water mass anomalies and calculation of geostrophic circulation anomalies for the TACE region. US deployments started already, more French deployments are planned as part of EGEE and the German participation in TA deployments is also funded to begin in 2004, so we may see a significant addition to ARGO coverage compared to **Fig. 4.2**.

- **Coastal and island stations (funded)**
- **Surface flux moorings**

for calibrating routine flux products; satellite–based products are known for serious biases in regions of frequent cloud cover. It is therefore proposed to expand into the eastern ITCZ area with two stations, located at 5N, 23W and 12N, 23W; adding pCO_2 sensors will allow to better delineate physical impacts on air-sea CO_2 fluxes.

- **Satellite observations (partially funded)**

As described in several of the science sections satellite products of altimetry, SST, wind stress, color and cloud cover are of prime importance for achieving TACE objectives and will be analyzed for the entire duration of the project

4.2.2 Enhanced Observing Period

Three seasonal cruises (during onset, peak and late monsoon) along the moored sections (23W, 10W and 6E) for at least two years to include

- intensive atmospheric as well as oceanographic observations crossing the ITCZ and cold tongue;
- small deliberate tracer experiment injections into the EUC to quantify mixing and the fate of EUC water;
- microstructure profiling of the upper 300m during the cruises in the cold tongue.

4.2.3 Special Observing Periods

Heat budget calculations and process studies for determining the mixing physics from research vessels at selected locations within the cold tongue and under and north of the ITCZ (see also section 6.2).

5. EGEE

The French Program « Etude de la circulation oceanique et de sa variabilite dans le Golfe de Guinee" **(EGEE)** is closely tied to the observational periods of AMMA. The EGEE cruises will be closely linked to the AMMA Extended Observation Period (EOP) as well as with the PIRATA program.

They also complement the data set obtained from the SVP drifters and PROVOR profiler network (actually, the PROVOR deployment in the Eastern tropical Atlantic and GG, especially via the EGEE cruises, constitutes a French commitment to the ARGO program in the framework of CORIOLIS). EGEE will include the following observations:

5.1 Ship surveys

In order to sample the GG during contrasting climatic situations, the cruises have to be carried out every six months during extreme seasons, i.e.

- boreal spring-summer (equatorial upwelling onset, in phase with the monsoon onset, around May-June), and
- boreal fall-winter (absence of equatorial upwelling, harmattan period over the coastal countries of west Africa, the ITCZ being in its southernmost position).

The cruises will be carried out for three consecutive years of the AMMA program, and during the two opposite seasons in each year. For comparability, the cruise tracklines will be systematically repeated. They will also have to run through the PIRATA buoys located in the region (to save R/V sea time, if PIRATA buoys are maintained).

5.1.1 Selection of shipboard sections

The 10°W meridional section has already been carried out several times during the PIRATA and both EQUALANT cruises, and is therefore selected (Fig. 5.1). The cruises will also "close" the GG area (for example to estimate mass balances) towards the African coast along a section at 6°S, already done during CITHER 1 and EQUALANT 2000. The cruises will have to pass by São Tomé, for maintainance of the meteorological station (see 5.3). The small meridional section at 6°E, between São Tomé and Nigeria, shows a section carried out during EQUALANT 2000, in very warm and fresh waters. During the cruise associated with the SOP of AMMA (2006), a section around 4°E between the equator and the coast will also be carried out (either in the south of the box " CATCH "), which could be retained for all the cruises (see the possible tracklines on Figures 6.a below, suitable for modifications).

5.1.2 Shipboard hydrography and current profiling observations

For most of the dynamical studies mentioned above, the measurements that have to be continuously acquired are the sea surface temperature and salinity (thermosalinograph), surface and subsurface currents (VM-ADCP), navigation parameters (GPS) and meteorology. To get the subsurface salinity and temperature structures, XBTs will be launched "en route" alternatively with "in station" CTD-O2 profiles (from the surface down to about 500-1000m depth; eventually with L-ADCP measurements if VM-ADCP does not allow to get current measurements below 200m depth); with a resolution of about every 1/2 degree in latitude/longitude. Profiles will also be done close to the PIRATA buoys, for comparison and validation of the ATLAS buoys oceanic data, and if possible (at least just before or after their deployment) close to the PROVOR floats for inter-calibration with the T/S profiles provided by these profilers. Sea surface water samplings will be regularly done for the salinity (thermosalinograph data calibration/validation), and nutrients.

During the first phase of the SOP, during late boreal spring 2006, a cruise similar to EQUALANT 2000, along most of the same trackline with measurements and samplings over the whole water column along with transient tracers (CFCs), will be carried out. Additional meteorological measurements are also planed, especially thanks to a turbulent flux measurement system of INSU/Météo-France, like the instrumented mast used during EQUALANT 1999. The final trackline of this particular "deep" cruise will be discussed with the meteorologists of AMMA, mostly for the repetition of the meridional section south of Benin, at 2°50'E (see Figure 6.b).

During the third phase of the SOP, *ie* during the late boreal summer 2006, a particular cruise will be dedicated to the late monsoon fluxes toward the Atlantic Ocean, and to the oceanic conditions during the cyclogenesis period. It will consist in a zonal section off Dakar (Senegal), a section across the Guinea Dome, and a repetition of the meridional section at 10°W. This cruise could be carried out simultaneously to an US/CLIVAR cruise in the center and the west of the Atlantic basin, around the same latitudes (see Figure 6.b).

5.1.3 Other shipboard work

The EGEE cruises are a component of AMMA but are also closely linked to different other programs:

PIRATA: the buoys of the PIRATA network deserve to be maintained at least till 2005 (consolidation period of this program), with a minimum of one cruise per year.

Surface drifter and ARGO float deployments: In the framework of the ARGO international program (and of its French component CORIOLIS), the EGEE cruises will be used to launch surface drifters (SVP) and subsurface drifting buoys (IFREMER PROVOR ones and eventually NOAA ALACE ones.

The maintenance of the moorings (current meters and Yoyo) of a different national program (see below).

Part of the measurements (from the thermosalinograph and CTD profiles) will be as far as possible sent in quasi-real time through satellite transmission in order to get them useable to the scientific community and for assimilation in the numerical models, particularly in the framework of MERCATOR (French component of GODAE). The numerical results of MERCATOR will be systematically compared to the measurements obtained during the cruises, for validation (if data are not yet assimilated in the model…).

The data will also be used for "ground truth" (SST, SSS, T and S profiles, winds, nutrients, currents…) of satellite data calibration, in this region where the cloudiness is particularly important.

5.2 Moored station work at 10W on equator

Current meter (ADCP) and cycling CTD (Yoyo) moorings located at 10°W on the equator, which were deployed during the EQUALANT program and continuation of which is planned in the framework of an other French national program.

5.3 A meteorological station at São Tome

This station in the eastern part of the GG has been installed by EGEE in fall 2003.

6. Relation to AMMA and other programs

6.1 AMMA

One of the motivations for this “white paper” was to present the basis for a Tropical Atlantic Climate Experiment (TACE) as an oceanographic underpinning of the mainly meteorologically oriented AMMA program. Therefore the TACE field work schedule as outlined here is closely related to the three observational periods of AMMA, and close cooperation of field work should be coordinated between both groups. For details of AMMA we refer to the AMMA Implementation document (ref…)

6.2 Cold tongue studies

As discussed at various places in this document, the physics of the eastern TA and cold tongue is very similar to the eastern tropical Pacific. Recently, the Eastern Pacific Investigation of Climate (EPIC 2001; Raymond et al., 2003) was carried out in the cold tongue regime. Plans are also underway, coordinated by the US Pacific CLIVAR Implementation Panel, for a Pacific Upwelling and Mixing Physics (PUMP) study. The objectives of both studies relate directly to answering important open questions also in the eastern Tropical Atlantic; implementation of PUMP may therefore reduce the observational requirements in the Atlantic regarding process studies on cold tongue physics.

6.3 Biogeochemical programs

As mentioned in some of the science sections of this document, there are opportunities for cooperations with biogeochemical objectives within the context of TACE. Both SOLAS and the International Ocean Carbon Coordination Project (IOCCP, a joint activity of the IOC CO2 Panel and the Global Carbon Project) have an interest in the TACE region, and relevant observations might be added to the TACE observational program, such as measuring oxygen on profiling floats or pCO_2 on flux buoys.

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8. List of Acronyms

ABA	**Angola Benguela Frontal Area**
ABF	**Angola-Benguela Front**
AC	**Angola Current**

AD	**Angola Dome**
AG	**Angola Gyre**
AGCMs	**Atmospheric GCMs**
AMI	**Atlantic Marine ITCZ**
AMIC	**Atlantic marine ITCZ complex**
AMIP	**Atmospheric Model Intercomparison Project**
AMMA	**Atlantic Monsoon Multidisciplinary Analysis**
APEX	**Autonomous profiling Explorer**
ARGO	**Array for Realtime Geostrophic Oceanography**
ATLAS	**Autonomous Temperature Line Acquisition System**
BC	**Brazil Current**
CATCH	**Couplage avec l'Atmosphère en Conditions Hivernales experiment**
CGCMs	**Coupled GCMs**
CESG	**Cross-Equatorial SST Gradient**
CITHER 1	**Etude de la CIrculation THERmohaline**
CLIVAR	**Climate Variability and Predictability**
COADS	**Comprehensive Ocean-Atmosphere Data Set**
ECT	**Eastern Cold Tongue**
EGEE	**Etude de la circulation océanique et de sa variabilité dans le Golfe de Guinée**
ENSO	**El Niño Southern Oscillation**
EOP	**Extended Observation Period**
EPIC	**Eastern Pacific Investigation of Climate**
EUC	**Equatorial Undercurrent**
ERS	**European Remote Sensing Satellite**
EQUALANT	**EQUAtorial atLANTic study**
GC	**Guinea Current**
GCM	**General Circulation Model**
GCUC	**Gabon-Congo Undercurrent**
GD	**Guinea Dome**
GG	**Gulf of Guinea**
GODAE	**Global Ocean Data Assimilation Experiment**
ITCZ	**Intertropical Convergence Zone**
IOCCP	**International Ocean Carbon Coordination Project**
IRD	**Institut de Recherche et Development**
LOP	**Long-Term Observing Period**
MOC	**Meridional Overturning Circulation**
MLD	**Mixed-layer Depth**
MPI OPYC	**Max Planck Institute**
NADW	**North Atlantic Deep Water**
NAO	**North Atlantic Oscillation**
NBC	**North Brazil Current**
NBUC	**North Brazil Undercurrent**
NEC	**North Equatorial Current**
NECC	**North Equatorial Countercurrent**
NCEP	**National Center for Environmental Prediction**
NEUC	**North Equatorial Undercurrent**
NTA	**North Tropical Atlantic**

NWP	
OCCAM	**Ocean Circulation and Climate Advanced Modelling Project**
OGCM	**Ocean General Circulation Model**
OISSTA	**Optimal Interpolation Sea Surface Temperature Anomaly**
OLR	**Outgoing Longware Radiation**
OPA	**Océan Parallélisé**
PIRATA	**Pilot Research Moored Array in the Tropical Atlantic**
PNA	**Pacific/North American**
PUMP	**Pacific Upwelling Modelling Physics**
PV	**Potential-Vorticity**
RAFOS	**SOFAR (Sound Fixing And Ranging) spelled backwards**
SEC	**South Equatorial Current**
SECC	**South Equatorial Countercurrent**
SEUC	**South Equatorial Undercurrent**
SLA	**Sea Level Anomaly**
SLP	**Sea Level Pressure**
SOLAS	**Surface Ocean Lower Atmosphere Study**
SOP	**Special Observing Period**
STA	**South Tropical Atlantic**
STCs	**Subtropical Cells**
STOIC	**Study of Tropical Oceans in Coupled GCMs project**
SSS	**Sea Surface Salinity**
SST	**Sea Surface Temperature**
SSTA	**Sea Surface Temperature Anomaly**
TA	**Tropical Atlantic**
TACE	**Tropical Atlantic Climate Experiment**
TAV	**Tropical Atlantic Variability**
TCs	**Tropical Cells**
TIWs	**Tropical Instability Waves**
VOS	**Voluntar Observing Ships**
WES	**Wind Evaporation-SST**
WSA	**Wind Speed Anomaly**

9. References

Arhan, M., H. Mercier, Y. Gouriou and B. Bourlès, 1998: Hydrographic sections across the Atlantic Ocean at 7°30'N and 4°30'S. *Deep-Sea Res. I*, **45**, 829–872.

Arnault, S., 1987: Tropical Atlantic geostrophic currents and ship drifts. *J. Geophys. Res.*, **92**(C5), 5076-5088.

Binet, D., B. Gobert and L. Maloueki, 2001: El Niño-like warm events in the Eastern Atlantic (6°N, 20°S) and fish availability from Congo to Angola (1964-1999), *Aquat. Living Resour.*, **14**, 99-113.

Blanke, B., M. Arhan, G. Madec, and S. Roche, 1999: Warm water paths in the equatorial Atlantic as diagnosed with a general circulation model. *J. Phys. Oceanogr.*, **29**, 2753-2768.

Blanke B. and P. Delecluse, 1993, Variability of the tropical Atlantic ocean simulated by a general circulation model with two different mixed layer physics, *J. Phys. Oceanogr.*, **23**, 1363-1388.
Bourlès, B., R.L. Molinari, E. Johns, W.D. Wilson and K.D. Leaman, 1999a: Upper Layer Currents in the Western Tropical Atlantic (1989-1991). *J.Geophys. Res.*, **104** (C1), 1361-1375.
Bourlès, B., Y. Gouriou and R. Chuchla, 1999b: On the circulation in the upper layer in the western equatorial Atlantic. *J. Geophys. Res.*, **104** (C9), 21,151-21,170.
Bourlès, B., M.D. D'Orgeville, G. Eldin, Y. Gouriou, R. Chuchla, Y. DuPenhoat, and S. Arnault, 2002: On the evolution of the thermocline and subthermocline eastward currents evolution in the Equatorial Atlantic. *Geophys. Res. Lett.*, **29**(16) 32-1-4.
Boyd, A.J., J. Salat and M. Masó, 1987: The seasonal intrusion of relatively saline water on the shelf off northern and central Namibia, The Benguela and Comparable Ecosystems, *S. Afr. J. Mar. Sci.*, **5,** 107-120.
Boyer, D. C., H. J. Boyer, I. Fossen and A. Kreiner, 2001: Changes in abundance of the northern Benguela sardine stock during the decade 1990-2000, with comments on the relative importance of fishing and the environment. *S. Afr. J. Mar. Sci.,* **23,** 76-84.
Boyer, D. C., and I. Hampton, 2001: An overview of the living marine resources of Namibia. *S. Afr. J. Mar. Sci.*, **23,** 5-35.
Carton, J.A., and B. Huang, 1994: Warm events in the Tropical Atlantic, *J. Phys. Oceanogr.*, **24,** 888-903.
Carton, J.A., X. Cao, B.S. Giese and A.M. da Silva, 1996: Decadal and interannual SST variability in the tropical Atlantic Ocean. *J. Phys. Oceanogr.*, **26**, 1165-1175.
Chang, P., L. Ji and H. Li, 1997: A decadal climate variation in the tropical Atlantic Ocean from thermodynamic air-sea interactions. *Nature*, **385**, 516-518.
Chang, P., R. Saravanan, L. Ji and G.C. Hegerl, 2000: The effects of local sea surface temperatures on atmospheric circulation over the tropical Atlantic sector. *J. Climate*, **13**, 2195-2216.
Chang, P., L. Ji and R. Saravanan, 2001: A hybrid coupled model study of tropical Atlantic variability. *J. Climate*, **14**, 361-390.
Chang, P., R. Saravanan and L. Ji, , 2003: Tropical Atlantic seasonal predictability: The roles of El Nino remote influence and thermodynamic air-sea feedback, *Geophys. Res. Lett.* **30**, 1501, doi: 10.1029/2002GL016119.
Chiang, J.C.H., M. Biasutti and D.S. Battisti, Sensitivity of the Atlantic ITCZ to conditions during Last Glacial Maximum. *Paleoceanogr.*, submitted, 2003
Clarke, A. J., and S. Van Gorder, 1994: On ENSO coastal currents and sea level, *J. phys. Oceanogr.*, **24,** 661-679.
Courel, M.F., R.S. Kandal, and S.I. Rasool, 1984: Surface albedo and Sahara drought. *Nature*, **307**, 528-531.
Czaja, A., P. van der Vaart and J. Marshall, 2002: A diagnostic study of the role of remote forcing in tropical Atlantic variability. *J. Climate*, **15**, 3280-3290.
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,** 15835-15847.
da Silva, A. M., C.C. Young and S. Levitus, 1994: Atlas of Surface Marine Data 1994. Vol. 3: Anomalies of Heat and Momentum Fluxes; Vol. 4: Anomalies of Fresh Water Fluxes NOAA Atlas NESDIS 8, U.S. Department of Commerce, NOAA, NESDIS, 413 pp.
Davey, M.K., et al., STOIC: a study of coupled model climatology and variability in tropical ocean regions. *Clim. Dyn.*, **18**, 403-420, 2002.
Delecluse, P., J. Servain, C. Levy, K. Arpe and L. Bengtsson, 1994: On the connection between the 1984 Atlantic warm event and the 1982-1983 ENSO, *Tellus*, **46,** 448-464.
Dommenget, D., and M. Latif, 2000: Interannual to decadal variability in the tropical Atlantic.

J. Climate, **13**, 777-792.
Drijfhout, S.S., P. de Vries, K. Doos, and A.C. Coward, 2003: Impact of eddy-induced transport on the Lagrangian structure of the upper branch of the thermohaline circulation. *J. Phys. Oceanogr.,* **33,** 2141-2155.
Enfield, D. B., and J. S. Allen, 1980: On the structure and dynamics of monthly mean sea level anomalies along the Pacific coast of North and South America, *J. Phys. Oceanogr.*, **10,** 557-578.
Enfield, D.B. and D.A. Mayer, 1997: Tropical Atlantic SST variability and its relation to El Nino-Southern Oscillation. *J. Geophys. Res.,* **102**, 929-945.
Florenchie P., J.R.E. Lutjeharms, C. J. C. Reason, S. Masson and M. Rouault, 2003a: The source of Benguela Niños in the South Atlantic Ocean, *Geophy. Res. Lett.*, **30**, 10, 1505, doi: 10.1029/2003GL017172.
Florenchie, P., Reason, C.J.C., Lutjeharms, J.R.E., Rouault, M., Roy, C. and S. Masson, 2003b: Evolution of Interannual Warm and Cold Events in the South-East Atlantic Ocean, *J. Climate*, under review.
Folland, C. K., T. N. Palmer and D. E. Parker, 1986: Sahel rainfall and world-wide sea temperatures. *Nature*, **320**, 602-607.
Fratantoni, D.M., W.E. Johns, T.L. Townsend and H.E. Hurlburt, 2000: Low-Latitude Circulation and Mass Transport Pathways in a Model of the Tropical Atlantic Ocean. *J. Phys. Oceanogr.*, **30**, 1944–1966.
Ganachaud, A. and C. Wunsch, 2001: Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. *Nature*, **408**, 453–457.
Gammelsrød, T., C.H. Bartholomae, D.C. Boyer, V.L.L. Filipe and M. J. O'Toole, 1998: Intrusion of warm surface water along the Angolan-Namibian coast in February-March 1995: the 1995 Benguela Niño, Benguela Dynamics, *S. Afr. J. Mar. Sci.*, **19,** 51-56.
Garzoli, S.L., A. Ffield, and Q. Yao, 2003: North Brazil Current Rings and the variability in the latitude of retroflection. In: Malanotte-Rizzoli, P., and G. Goni (eds.). *Interhemispheric Water Exchange in the Atlantic Ocean.* Elsevier.
Giannini, A., R. Saravanan, and P. Chang, 2003: Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science express*, **Oct**, 10.1126.
Gilson J.K., P. Kallberg, S. Uppala, A. Hernandez, A. Nomura and E. Serrrano, 1999: ECMWF Re-Analysis project report series. 1. ERA-15, (version 2), *ECMWF*.
Goni, G.J. and W.E. Johns, 2003: Synoptic study of warm rings in the North Brazil: Current retroflection region using satellite altimetry. In: *Interhemispheric Water Exchange in the Atlantic Ocean.* Eds. Goni, G.J. and P. M. Rizzoli. *Elsevier Oceanography Series*, **68**, 335-356.
Gouriou, Y. and G. Reverdin, 1992: Isopycnal and Diapycnal Circulation of the Upper Equatorial Atlantic Ocean in 1983-1984. *J. Geophys. Res.*, **97 (C3)**, 3543–3572.
Gray, W., and J.D. Schaeffer, 1991: Teleconnections linking worldwide climate anomalies. M.H. Glantz, R.W. Katz, and N. Nicholls, *Es. Cambridge University Press*, 535 pp.
Grodsky, S.A. and J.A. Carton, 2002: Surface drifter pathways originating in the equatorial Atlantic cold tongue. *Geophys. Res. Lett.*, **29 (23)**, 62-65.
Grodsky, S.A. and J.A. Carton, 2003: The intertropical convergence zone in the South Atlantic and the Equatorial Cold Tongue. *J. Clim.*, **16(4)**, 723–733.
Gu, D., and S.G.H. Philander, 1997: Interdecadal Climate Fluctuations that depend on Exchanges between the Tropics and Extratropics. *Science*, **272**, 805–808.
Hastenrath, S., and L. Heller, 1977: Dynamics of climate hazards in Northeast Brazil, *Q. J. Roy. Met. Soc.*, **103**, 77-92.
Hastenrath, S., 1990: Decadal-scale changes of the circulation in the tropical Atlantic sector associated with Sahel drought, *Int. J. of Climatology*, **10**, 459-472.

Hastenrath, S. and L. Greischar, 1993: Further work on the prediction of northeast Brazil rainfall anomalies. *J. Climate,* **6**, 743-758.
Hazeleger, W., M. Visbeck, M. Cane, A. Karspack and Naomi Naik, 2001a: Decadal upper ocean temperature variability in the tropical Pacific. *J. Geophys. Res*, **106**, 8971-8988.
Hazeleger, W., R. Seager, M. Visbeck, N. Naik, and K. Rodgers, 2001b: On the impact of the midlatitude storm track on the upper Pacific Ocean". *J. Phys. Oceanogr.*, **31**, 616-636.
Hazeleger, W., P. de Vries, and Y. Friocourt, 2003: Sources of the Equatorial Undercurrent in the Atlantic in a high-resolution ocean model. *J. Phys. Oceanogr.*, **33**, 677-693.
Hazeleger, W., and P. de Vries, 2003: Fate of the Equatorial Undercurrent in the Atlantic., W., C. Severijns, R. Haarsma, F. Selten, and A. Sterl, 2003: SPEEDO. Model description and validation of a flexible coupled model for climate studies. *KNMI TR*, 38pp.
Hellerman, S., and M. Rosenstein, 1983: Normal monthly wind stress over the world ocean with error estimates. *J. Phys. Oceanogr.*, **13**, 1093–1105.
Hirst, A. C., and S. Hastenrath, 1983: Atmosphere-ocean mechanisms of climate anomalies in the Angola-tropical Atlantic sector, *J. Phys. Oceanogr.*, **13,** 1146-1157.
Horel, J.D., V. E. Kousky and M.T. Kagano, 1986: Atmospheric conditions in the Atlantic sector during 1983 and 1984, *Nature*, **322,** 248-251.
Houghton, R.W. and Y.M. Tourre, 1992: Characteristics of low-frequency sea surface temperature fluctuations in the tropical Atlantic. *J. Climate,* **5**, 765-771.
Huang, R. X. and B. Qiu, 1994: Three-dimensional structure of the wind-driven circulation in the Subtropical North Pacific. *J. Phys. Oceanogr.*, **24**, 1608-1622.
Huang, B., and J. Shukla, 1997: Characteristics of the Interannual and Decadal Variability in a General Circulation Model of the Tropical Atlantic Ocean. *J. Phys. Oceanogr.*, **27 (8)**, 1693–1712.
Huang, B., P.S. Schopf and Z. Pan, 2002: The ENSO effect on the tropical Atlantic variability: A regionally coupled model study. *Geophys. Res. Lett.* **29**, 2039, doi: 10.1029/2002GL014872.
Inui, T., A. Lazar, P. Malanotte-Rizzoli, A. Busalacchi, 2002: Wind Stress Effects on Subsurface Pathways from the Subtropical to Tropical Atlantic. *J. Phys. Oceanogr.*, **32 (8)**, 2257–2276.
Jochum, M., P. Malanotte-Rizzoli, and A. Busalacchi, 2003: Tropical Instability Waves in the Atlantic ocean. *Ocean Mod*. In press.
Jochum, M., R. Murtugudde, P. Malanotte-Rizzoli, and A.J. Busalacchi, 2003: Intrinsic variability in the tropical Atlantic Ocean. Subm. to "*Ocean-Atmosphere Interaction and Clim. Variability*" AGU monograph.
Johns, W.E., T.N. Lee, F. Schott, R. Zantopp and R. Evans, 1990: The North Brazil Current retroflection: seasonal structure and eddy variability. *J. Geophys. Res*., **95**(C12), 22,103-22,120.
Joyce, T. M., C. Frankignoul, J. Yang, H.E. Phillips, 2003: Ocean response and feedback to the SST dipole in the Tropical Atlantic. *J. Phys. Oceanogr*., subm.
Karstensen, J. and D. Quadfasel, 2002: Formation of Southern Hemisphere Thermocline Waters: Water mass Conversion and Subduction. *J. Phys. Oceanogr.*, **32 (11)**, 3020–3038.
Katz, E. J., 1987: Waves along the equator in the Atlantic, *J. Phys. Oceanogr.*, **27,** 2536-2544.
Kleeman, R., J.P. McCreary Jr. and B.A. Klinger, 1999: A mechanism for generating ENSO decadal variability. *Geophys. Res. Lett.*, **26 (12)**, 1743–1746.
Klein, S.A., B.J. Soden and N.-C. Lau, 1999: Remote sea surface temperature variations during ENSO: Evidence for a tropical atmospheric bridge. *J. Climate,* **12**, 917-932.
Kostianoy, A. G., and J. R. E. Lutjeharms, 1999: Atmospheric effects in the Angola-Benguela Frontal Zone, *J. Geophys. Res.*, **104,** 20962-20970.

Kushnir, Y, R. Seager and J. Miller, 2002: A simple coupled model of tropical Atlantic decadal climate variability. *Geophys. Res. Lett.,* **29**, 2133, doi: 10.1029/2002GL015874.
Lamb, P.J., 1978: Large–scale tropical Atlantic surface circulation patterns associated with sub-Saharan weather anomalies. *Tellus*, **30**, 240-251.
Lamb, P.J., and R.A. Peppler, 1992: Further case studies of tropical Atlantic surface atmospheric and oceanic patterns associated with sub-Saharan drought, *J. Climate,* **5**, 476-488.
Lass, H.U., M. Schmidt, V. Mohrholz and G. Nausch, 2000: Hydrographic and current measurements in the area of the Angola-Benguela front, *J. Phys. Oceanogr.*, **30,** 2589-2609.
Latif, M., and T.P. Barnett, 1995: Interactions of the tropical oceans, *J. Climate*, **8,** 952-964.
Lazar, A., R. Murtugudde and A.J. Busalacchi, 2001: A model study of temperature anomaly propagation from the subtropics to tropics within the South Atlantic thermocline. *Geophys. Res. Lett.*, **28**, 1271-1274.
Lazar, A., T. Inui, P. Malanotte-Rizzoli, A.J. Busalacchi, L. Wang and R. Murtugudde, 2002: Seasonality of the ventilation of the tropical Atlantic thermocline in an ocean general circulation model. *J. Geophys. Res.,* **107 (C8)**, 3104, do.10.1029/2000JC00066*7.*
Levitus, S., 1982: Climatological atlas of the world ocean, *NOAA Prof. Pap.,* **13**, 173pp., U.S. Govt. Print. Off., Washigton D.C..
Lough, J. M., 1986: Tropical Atlantic sea surface temperature and rainfall variations in subsaharan Africa. *Mon. Wea. Rev.*, **114**, 561-570.
Lumpkin, R. and K. Speer, 2003: Large-scale vertical and horizontal circulation in the North AtlanticOcean. *J. Phys. Oceanogr.*, **33**, 1902-1920.
Lutjeharms, J. R. E., P.M.S. Monteiro, P.D. Tyson and D. Obura, 2001: The oceans around southern Africa and regional effects of global change, *S. Afr. J. Sci*., **97**, 119-130.
Lux, M., H. Mercier, and M. Arhan, 2001: Interhemispheric exchanges of mass and heat in the Atlantic Ocean in January-March 1993. *Deep-Sea Res. I*, **48**, 605–638.
Maes C., P. Delecluse and G. Madec, 1998, Impact of westerly wind bursts on the warm pool of the TOGA-COARE domain in an OGCM, *Clim. Dyn*., **14**, 55-70.
Madec G., P. Delecluse, M. Imbard and C. Levy, 1999: OPA 8.1 Ocean General Circulation Model reference manual, Internal report number XX, Institut Pierre-simon Laplace (IPSL), Paris, **XX**.
Malanotte-Rizzoli, P. K. Hedstrom, H. Arango and D.B. Haidvogel, 2000: Water mass pathways between the subtropical and tropical ocean in a climatological simulation of the North Atlantic ocean circulation. *Dyn. Atmospheres and Oceans*, **32**, 331–371.
Marshall, J., et al., 2001: North Atlantic climate variability: phenomena, impacts and mechanisms. *Int. J. Climatol*., **21**, 1863-1898.
Masson S., P. Delecluse, J. P. Boulanger and C. Menkes, 2002, A model study of the Seasonal variability and formation mechanisms of barrier layer in the eastern equatorial Indian Ocean, *J. Geophys. Res.*, In Press.
Mayer, D.A., R.L. Molinari and J.F. Fiesta, 1998: The mean and annual cycle of upper layer temperature fields in relation to Sverdrup dynamics within the gyres of the Atlantic Ocean. *J. Geophys. Res.*, **103**, 18545-18566.
McCreary, J., J. Picaut, and D. Moore, 1984: Effects of the remote annual forcing in the eastern tropical Atlantic Ocean. *J. Mar. Res.,* **42**, 45-81.
McPhaden, M. J. and D. Zhang, 2002: Slowdown of the meridional overturning circulation in the upper Pacific Ocean. *Nature*, **415**, 603-608.
Meeuwis, J. M., and J. R. E. Lutjeharms, 1990: Surface thermal characteristic of the Angola-Benguela front, *S. Afr. J. Mar. Sci.*, **9**, 261-279.
Mehta, V.M., 1998:Variability of the tropical ocean surface temperatures at decadal-multidecadal time scales. Part I: The Atlantic Ocean. *J. Climate,* **11**, 2351-2375.

Mo, K.C. and S. Hakkinen, 2001: Interannual variability of the tropical Atlantic and linkages to the Pacific. *J. Climate*, **14**, 2740-2762.

Mohrholz, V., M. Schmidt and J.R.E Lutjeharms, 2001: The hydrography and dynamics of the Angola-Benguela Frontal Zone and environment in April 1999, *S. Afr. J. Sci.,* **97,** 199-208.

Molinari, R.L., S. Bauer, D.P. Snowdon, G.C. Johnson, B. Bourles, Y. Gouriou and H. Mercier, 2003: A Comparison of kinematic evidence for tropical cells in the Atlantic and Pacific oceans. In: Goni, G. and P. Malanotte-Rizzoli (eds.). Interhemispheric Water Exchange in the Atlantic Ocean. *Elsevier Oceanographic Series*, 269-286.

Moura, A. and J. Shukla, 1998: On the dynamics of droughts in northeast Brazil: Observations, theory and numerical experiments with a general circulation model. *J. Atmos. Sci.,* **38**, 2653-2675.

Moura, A., and J. Shukla, 1981: On the Dynamics of Droughts in Northeast Brazil: Observations, Theory and Numerical Experiments with a General Circulation Model, *Journal of Atmospheric Sciences*, **38**, 2653-2675.

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.

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

Otterman, J., 1974: Baring high-albedo soils by overgrazing: a hypothesized desertification mechanism. *Science*, **186**, 531-533.

Pailler, K., B. Bourlès, et Y. Gouriou, The barrier layer in the western tropical Atlantic, *Geophys. Res. Lett.,* Vol.**26**, 14, 2069-2072*, 1999.*

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

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

Philander, G., and R.C. Pacanowski, 1980: The generation of equatorial currents. *J. Geophys. Res.*, **85**, 1123-1136.

Picaut, J., 1985: Major dynamics affecting the eastern tropical Atlantic and Pacific oceans, CalCOFI Rep., Vol XXVI.

Picaut, J., and J. Busalacchi, 2001: Tropical Ocean Variability, in *Satellite Altimetry and Earth Sciences*, Academic Press, Chap. **4**, pp. 217-236.

Pizarro, O., A. J. Clarke and S. Van Gorder, 2001: El Niño sea level and currents along the South American coast: comparisons of observations with theory, *J. Phys. Oceanogr.*, **31,** 1891-1903.

Qui, B. and R.X. Huang, 1995: Ventilation of the North Atlantic and North Pacific: Subduction Versus Obduction. *J. Phys. Oceanogr.*, **25**, 2374-2390.

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

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

Reason, C.J.C., 2000: Multidecadal climate variability in the subtropics/midlatitudes of the Southern Hemisphere oceans. *Tellus*, **52A**, 203-223.

Reason, C.J.C., R.J. Allan, J.A. Lindesay and T.J. Ansell, 2000: ENSO and climatic signals across the Indian Ocean basin in the global context: Part I, Interannual composite patterns, *Int. J. Clim.,* **20**, 1285-1327.

Reverdin, G., 1985: Convergence in the equatorial surface jets of the Indian Ocean. *J. Geophys. Res.*, **90**, 11741-11750.

Reynolds, R. W., and T. M. Smith, 1994: Improved global sea surface temperature analyses using optimum interpolation. *J. Climate*, **7,** 929-948.
Richardson, P.L. and D. Walsh, 1986: Mapping climatological seasonal variations of surface currents in the tropical Atlantic using ship drift data. *J. Geophys. Res*., **91**, 10537-10550.
Risien, C., C.J.C. Reason, F. Shillington, D. Chelton, 2003: Variability in satellite winds over the Benguela upwelling system during 1999-2000. *J. Geophys. Res.*, accepted.
Robertson, A.W., C.R. Mechoso, and Y.J. Kim, 2000: The influence of Atlantic sea surface temperature anomlies on the North Atlantic Oscillation. *J. Climate*, **13**, 122-138.
Roemmich, D., 1983: The balance of geostrophic and Ekman transports in the tropical Atlantic Ocean. *J. Phys. Oceanogr.*, **13**, 1534-1539.
Rouault, M., P. Florenchie, N. Faucherau and C. J. C. Reason, 2003: South east tropical Atlantic warm events and southern African rainfall, *Geophy. Res. Lett.*, **30, 5**, 8009, doi: 10.1029/2003GL014840.
Rowell, D.P., C.K. Folland, K. Maskell, and M.N. Ward, 1995: Variability of summer rainfall over tropical north Africa (1906-92): Observations and modelling. *Q. J. R. Meteorol. Soc.*, **121**, 669-704.
Roy, C., and C. Reason, 2001: Enso related modulation of coastal upwelling in the eastern Atlantic*, Prog. Oceanogr.*, **49,** 245-255.
Ruiz-Barradas, A., J.A. Carton and S. Nigam, 2000: Structure of interannual-to-decadal climate variability in the tropical Atlantic sector. *J. Climate*, **13**, 3285-3297.
Saravanan, R., and P. Chang, 2000: Interaction between tropical Atlantic variability and El Nino-Southern Oscillation, *J. Climate*, **13**, 2177-2194.
Schneider, N., A.J. Miller, M.A. Alexander, and C. Deser, 1999: Subduction of decadal North Pacific temperature anomalies: Observations and dynamics. *J. Phys. Oceanogr*., **29**, 1056-1070.
Schott, F.A., and C.W. Boening, 1991: The WOCE model in the western equatorial Atlantic: upper layer circulation. *J. Geoph. Res*., **96**, 6693-7004.
Schott, F., J. Fischer, J. Reppin and U. Send, 1993: On mean and seasonal currents and transports at the western boundary of the equatorial Atlantic. *J. Geophys. Res.*, **98**, 14353-14368.
Schott, F., J. Fischer and L. Stramma, 1998: Transports and pathways of the upper-layer circulation in the western tropical Atlantic. *J. Phys. Oceanogr.*, **28(10)**, 1904–1928.
Schott, F., P. Brandt, M. Hamann, J. Fischer and L. Stramma, 2002: On the boundary flow off Brazil at 5-10°S and its connection to the interior tropical Atlantic. *Geophys. Res. Lett.*, **29 (17)**, 1840.
Schott, F., M. Dengler, P. Brandt, K. Affler Fischer J., B. Bourlès, Y. Gouriou, R.L. Molinari and M. Rhein, 2003: The zonal currents and transports at 35°W in the tropical Atlantic. *Geophys. Res. Lett.*, **30 (7)** 1349-1352.
Schott, F. A., J. P. McCreary, and G. C. Johnson, 2004: Shallow overturning circulations of the tropical-subtropical oceans. In: *Ocean-Atmosphere Interaction and Climate Variability,* AGU Geophys. Monogr. ed. by C. Wang, S.-P. Xie, and J. A. Carton, AGU, Washington, D. C. (in press).
Seager, R., Y. Kushnir, P. Chang, N. Naik, J. Miller, and W. Hazeleger, Looking for the role of the ocean in tropical Atlantic decadal climate variability, *J. Climate*, **14**, 638-655, 2001.
Send, U., C. Eden, and F. Schott, 2002: Atlantic Equatorial Deep Jets: Space/Time Structure and Cross-Equatorial Fluxes. *J. Phys. Oc*., **32** (3), 891-902.
Servain, J., D. Legler, J. Picaut, V. Verbeque, et M. Seva, 1985: Oceanic variability and wind stress forcing in the tropical Atlantic Ocean. *IAMAP/IAPSO Joint Assembly, 5-16 August 1985, Honolulu, HI, Paper 06-16.*
Servain, J., 1991: Simple climate indices for the tropical Atlantic Ocean and some

applications. *J. Geophys. Res.* **96**, 15137-15146.
Servain, J., and S. Arnault, 1995: On forecasting abnormal climatic events in the tropical Atlantic Ocean. *Ann. Geophysicae*, **13,** 995-1008.
Servain, J., A. J. Busalacchi, M. J. McPhaden, A. D. Moura, G. Reverdin, M. Vianna and S. E. Zebiak, 1998: A Pilot Research Moored Array in the tropical Atlantic (PIRATA). *Bull. Amer. Meteor. Soc.*, **79,** 2019-2031
Servain, J., I. Wainer, J.P. McCreary and A. Dessier, 1999: Relationship between the equatorial and meridional modes of climatic variability in the tropical Atlantic, *Geophy. Res. Lett.,***26**, 485-488.
Shannon, L.V., A.J. Boyd, G.B. Bundrit and J. Taunton-Clark, 1986: On the existence of an El Niño-type phenomenon in the Benguela system, *J. Mar. Sci.*, **44,** 495-520.
Shannon, L.V., J.J. Agenbag and M.E.L. Buys, 1987: Large and mesoscale features of the Angola-Benguela front, *S. Afr. J. Mar. Sci.*, **5,** 11-34.
Shannon, L.V., and J.J. Agenbag, 1990: A large-scale perspective on interannual variability in the environment in the South-East Atlantic, *S. Afr. Mar. Sci.*, **9,** 161-168.
Snowden, D.P. and R.L. Molinari, 2003: Subtropical Cells in the Atlantic Ocean: An observational summary. In: Goni, G., and P. Malanotte-Rizzoli (eds.). *Interhemispheric Water Exchange in the Atlantic Ocean.* Elsevier Oceanographic Series, 287-312.
Sterl, A., and W. Hazeleger, 2003: Coupled variability and air-sea interaction in the South Atlantic Ocean. *Clim Dyn.* DOI: 10.1007/s00382-003-0348-y.
Stramma, L. J. Fischer and J. Reppin, 1995: The North Brazil Undercurrent. *Deep Sea Res*., **42**, 773-795.
Stramma, L., and F. Schott, 1999: The mean flow field of the tropical Atlantic Ocean. *Deep-Sea Research II*, **46**, 279-303.
Stramma, L., Fischer, J., Brandt, P., Schott, F., 2003. Circulation, variability and near-equatorial meridional flow in the central tropical Atlantic. In: Goni, G., and P. Malanotte-Rizzoli, (eds.). *Interhemispheric Water Exchange in the Atlantic Ocean*. Elsevier Oceanographic Series 68, 1-22.
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.
Sutton, R.T., W.A. Norton, and S.P. Jewson, 2001: The North Atlantic Oscillation— What role for the ocean? *Atmos. Sci. Lett.*, 10.1006/2000.0018.
Tanimoto, Y. and S.-P. Xie, 2002: Inter-hemispheric decadal variations in SST, surface wind, heat flux and cloud cover over the Atlantic Ocean. *J. Meteor. Soc. Japan*, **80**, 1199-1219.
Taunton-Clark, J., and L.V. Shannon, 1988: Annual and interannual variability in the south-east Atlantic during the 20th century, *S. Afr. J. Mar. Sci.*, **6,** 97-106.
Terray, L. and C. Cassou, 2002: Tropical Atlantic sea surface temperature forcing of quasi-decadal climate variability over the North Atlantic-European region. *J. Climate,* **15**, 3170-3187.
Torrence, C., and G.P. Compo, 1998: A practical guide to wavelet analysis, *Bull. Amer. Meteor. Soc*., **79,** 61-78.
Tourre, Y.M., B. Rajagopalan and Y. Kushnir, Dominant patterns of climate variability in the Atlantic Ocean during the last 136 years. *J. Climate*, **12**, 2285-2299, 1999.
Venegas, S.A., L.A. Mysak, and D.N. Straub, 1997: Atmosphere-ocean coupled variability in the South Atlantic. *J. Climate,* **10**, 2904-2920.
Venegas, S.A., L.A. Mysak and D.N. Straub, 1998: An interdecadal climate cycle in the South Atlantic and its links to other basins. *J. Geophys. Res.,* **103**, 24723-24736.
Voituriez, B., 1981: Les sous-courants équatoriaux nord et sud et la formation des dômes thermiques tropicaux, *Oceanol. Acta, 4, 497-506.*

Wacogne, S. and B. Piton, 1992: The Near-Surface Circulation in the Northeastern Corner of the South Atlantic Ocean. *Deep-Sea Res.*, **39**, 1273-1298.
Walker, N.D., 1987: Interannual sea surface temperature variability and associated atmospheric forcing within the Benguela system, *S. Afr. J. Mar. Sci.*, **5,** 121-132.
Weare, B. C., 1977: Empirical orthogonal analysis of Atlantic Ocean surface temperatures. *Quart. J. Roy. Meteor. Soc.*, **103**, 467-478.
Weisberg, R.H. and T.J. Weingartner, 1988: Instability Waves in the Equatorial Atlantic Ocean. *J. Phys. Oceanogr.*, **18** (11), 1514-1533.
Wilson, W.D., E. Johns and R.L. Molinari, 1994: Upper layer circulation in the western tropical North Atlantic Ocean during August 1989. *J. Geophys. Res.*, **99 (C11)**, 22,513-22,523.
Wu, L. and Z. Liu, 2002: Is tropical Atlantic variability driven by the North Atlantic Oscillation? *Geophys. Res. Lett.*, **29**, 10.1029/2002GL014939.
Xie, S.-P. and Y. Tanimoto, 1998: A pan-Atlantic decadal climate oscillation. *Geophys. Res. Lett.,* **25**, 2185-2188.
Xie P. and P. A. Arkin, 1998: Global monthly precipitation estimates from satellite-observed outgoing longwave radiation, *J. Climate*, **11**,137-163.
Xie, S.-P., 1999: A dynamic ocean-atmosphere model of the tropical Atlantic decadal variability. *J. Climate*, **12**, 64-70.
Xie, S.-P. and J.A. Carton, 2004: Tropical Atlantic Variability: Patterns, mechanisms, and impacts. In: *Ocean-Atmosphere Interaction and Climate Variability*, AGU Geophys. Monogr., edited by C. Wang, S.-P. Xie, and J.A. Carton,subm.
Xue, Y, and J. Shukla, 1993: The influence of land surface properties on Sahel climate. Part I: Desertification. *J. Climate*, **5**, 2232-2245.
Yang, J., 1999: A linkage for decadal climate variations in the Labrador Sea and the tropical Atlantic Ocean. *Geoph. Res. Lett.*, **30**, 1070-1073.
Zhang, D., M.J. McPhaden, and W.E. Johns, 2003: Observational evidence for flow between the subtropical and tropical Atlantic: The Atlantic subtropical cells. *J. Phys. Oceanogr.,* **33**, 1783-1797.
Zebiak, S.E., 1993: Air-sea interaction in the equatorial Atlantic region, *J. Climate*, **6**, 1567-158.

10. Figure captions

Fig. 1.0.1: *Annual mean SST zonal section for equatorial strip 2N-2S, as observed for non-flux corrected CGCMs (with permission of M. Davey, see Davey et al 2002)*

Fig. 1.1.1 *a) 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 demarking 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.*

b) Annual migration of the Atlantic marine ITCZ as depicted in two longitude-sector averages (indicated on figure) climatological mean rainfall (in mm/day, see color scale) as a function of the calendar month (abscissa) and latitude (ordinate). Also shown is the first EOF of the variability around the climatology in the latitude – calendar month space (also in mm/day). Data are from the NASA GPCP (Hofman et al., 1997) spanning the years 1999-2001.

Fig. 1.1.2*: Schematic maps showing the horizontal distribution of the major tropical currents for the Tropical Surface Water layer at about 0-100 m depth for a) the northern spring and for b) the northern fall. Shown are the North Equatorial Current (NEC), the Guinea Dome (GD), the North Equatorial Countercurrent (NECC), the Guinea Current (GC), the South Equatorial Current (SEC) with the northern (nSEC), equatorial (eSEC), central (cSEC) and southern branches (sSEC), the Equatorial Undercurrent (EUC), the North Brazil Current (NBC), the Gabon-Congo Undercurrent (GCUC), the Angola Gyre (AG), the Angola Current (AC), the Angola Dome (AD), the South Equatorial Undercurrent (SEUC), the South Equatorial Countercurrent (SECC) and the Brazil Current (BC). The Angola-Benguela Front (ABF) is included as a dashed line. "Up" marks possible areas of upwelling, but not the exact places (see also Fig. 1.5.2); (from Stramma et al., 2003).*

Fig. 1.1.3: *a) Typical boreal spring variability in the tropical Atlantic region presented in terms of the first EOF (explains 33% of the variance) of March-April rainfall from GPCP 1979-2001 (contours in mm/day). The March-April SST anomaly (colors, in °C & white contours, every 0.2°) and surface wind anomaly (vector, in m/sec) are determined through regression on the time series of the rainfall EOF.*

b) Typical boreal summer variability in the tropical Atlantic region presented in terms of the first EOF (explains 23% of the variance) of the June-August rainfall from GPCP 1979-2001 (contours in mm/day). The June-August SST anomaly (colors, in °C & white contours, every 0.2°) and surface wind anomaly (vector, in m/sec) are determined through regression on the time series of the rainfall EOF.

Fig. 1.1.4: *Anomalies associated with the Atlantic Nino principal component from a five-component rotated principal component analysis of observed heat content ($\times 10^{8}\, Jm^{2}$, upper left), wind stress ($dyn\, cm^{-2}$, upper right), SST (^{o}C) middle left, diabatic heating at 500mb ($^{o}C\ dy^{-1}$, middle right) and the 12-month smoothed PC time series (bottom). From Ruiz-Barradas et al. (2000).*

Fig. 1.2.1 : *a) Time series of SST in NW (5-25N, 55-15W) and SE (0-20S, 30W-10E) tropical Atlantic and their difference; b)comparison of zonal wind anomaly and SST differences between NW and SE Atlantic and with Hadley cell Index of Wang (2001)*

Fig. 1.2.2 : *Seasonal variance partition between internal and SST-forced of three wind indices a) NE trades, b) cross-equatorial, c) zonal equatorial in an atmospheric model driven by observed SST; lower panels show fractions of SST-forced variance due to ENSO, Alt. equatorial SST (atl3) and SST dipole index (from Sutton et al., 2000)*

Fig. 1.3.1 : *Sea surface temperature (right) and sea level (left) anomalies along the African coast from the equator to 30°S versus time. After Florenchie et al., 2003).*

Fig 1.3.2: *Time series of zonal WSA averaged south of the equator (between 5.5S and 0.5S) in the central basin from 29.5W to 9.5W and (a) SLA averaged over the Topex box and (b) OISSTA averaged over the ABA. After Florenchie et al., 2003b.*

Fig. 1.3.3 *a) Subsurface propagation of warm and cold subsurface anomalies along the coast in March 1984 and 1997;*

b) Time – depth evolution 1984 over 1 month, shoaling from 100 to 45 m across tropical Atlantic. After Florenchie et al. (2003b).

Fig.1.5.1: *Schematic representation of the Atlantic STC circulation with subduction (blue), upwelling (green) zones and Ekman transports (red) that participate in the STC. For identified current branches participating in STC flows see Fig. 1.1.2. Interior equatorward thermocline pathways dotted, transport estimates marked for interior and western boundary pathways; see text for details.*

Fig. 1.5.2: *Time series of Ekman transport divergence across 10^0 N /10^0 S of monthly (thin) and lowpassed (heavy) anomalies, calculated from NCEP reanalysis (blue) and ERS-1/2 scatterometer (red) wind stresses.*

Fig. 1.5.3: *Ekman upwelling distribution for tropical Atlantic (outside $\pm 3^0$ latitude belt) from ERS-1/2 scatterometry wind stresses 1991-99 for a) February, b) August.*

Fig. 1.5.4: *Mean zonal current distributions a) across the North Brazil Undercurrent at 5^0S (7 sections), b) across the zonal equatorial current system at 35^0W (11 sections) and at 23^0W (1 section); after Schott et al.,(2002, 2003).*

Fig. 1.5.5: *Distribution of potential vorticity and b) geostrophic currents on the isopycnal surface = 25.4 kg m^{-3} for the tropical Atlantic, based on climatological hydrographic data. c) Interior net meridional transports by density layers across 10^0N (African coast to 60W, left) and 6^0S (African coast to 35^0W, right). From Zhang et al. (2003).*

Fig. 1.5.6: *Subduction sites of water masses that ventilate the Equatorial Undercurrent at 20W (data from eddy-permitting OCCAM model, see Hazeleger et al. 2003a).*

Figure 1.6.1: *Upper ocean circulation anomlalies for (a) 15 days and (b) 120 days after perturbation of the MOC by introduction of a freshwater anomaly in the subpolar North Atlantic, and time variability of (c) meridional flow off Florida and (d) zonal currents in the equatorial Atlantic, between perturbed and control simulations (from Dong and Sutton, 2003).*

Figure 1.6.2: *Global SST anomalies in years 2, 4, and 6 after reduction of the MOC by a subpolar freshwater anomaly (from Dong and Sutton, 2003).*

Figure 1.6.3: *Global precipitation anomalies (mm/day) in (a) year 6 and (b) year 7 after a reduction in the MOC (from Dong and Sutton, 2003).*

Fig. 2.1: *Annual mean SST zonal section for equatorial strip 2N-2S, as observed and for flux corrected CGCMs (compare with **Fig. 1.0.1** for non-flux corrected; with permission of M. Davey, see Davey et al 2002)*

Fig. 2.2: *Summary of seasonality of the dominant elements of climate variability in the tropical Atlantic region (see Sutton et al. 2000)*

Fig. 2.3: *Temperature along the equator in a coupled primitive equation model (SPEEDO, see Hazeleger et al. 2003c). Continuous lines: with rather strong turbulent wind mixing*

coefficent in the Kraus-Turner mixed layer parameterization. Dashed lines: weak turbulent wind mixing coefficent.

Fig. 4.1: *Schematic diagram of proposed observations for TACE with moored currentmeter arrays, flux moorings, drifters and floats; also shown is regression of SST on 1st EOF of July-Sept. W.African rainfall.*

Fig. 4.2: *Trajectories of profiling floats in the TACE region at shallow STC levels (200 m red; 400 m green) and deep (1900 blue), Nov. 2003* (http://www.coriolis.eu.org).

Figure 5.1 *proposed EGEE alternative repeat cruise tracklines, including a meridional section around 3°E, south of Benin and of the "CATCH" area. The option on the r.h.s. is slightly more expensive in vessel time. The position of the four ATLAS buoys of the PIRATA network are also represented (black dots). The passing through the 10°S-10°W buoy will depend upon the PIRATA program demand (visit of the buoy at least once a year), but the zonal section at 6°S could be shifted at 10°S.*

Figure 5.2. *proposed EGEE cruises tracklines in 2006, during the AMMA SOP 1 (May-July, on the left) and SOP 3 (end of August-September, on the right). The position of the four ATLAS buoys of the PIRATA network are also represented (black dots). The passing through the 10°S-10°W buoy will depend upon the PIRATA program demand (visit of the buoy at least once a year), but the zonal section at 6°S could be shifted to 10°S.*