Dominant Patterns of Climate Variability in the Atlantic Ocean Region During the Last 136 Years

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Abstract

Dominant spatio-temporal patterns of joint sea surface temperature (SST) and sea level pressure (SLP) variability in the Atlantic Ocean are identified using a multivariate frequency domain analysis. Five significant frequency-bands are isolated ranging from the quasi-biennial to the quasi-decadal. Two quasi-biennial bands are centered around 2.2 and 2.7-year periods; two interannual bands are centered around 3.5 and 4.4-year periods; the fifth band at the quasi-decadal frequency is centered around 11.4-year period. Between 1920 and 1955, the quasi-decadal band is less prominent compared to the quasi-biennial bands. This happens to be the period when SLP gradually increased over the Greenland-Iceland regions. The spatial pattern at the quasi-decadal frequency displays an out-of-phase relationship in the SLP in the vicinity of the subtropical anticyclones in both hemispheres (indicative of an out-of-phase quasi-decadal variability in the North and South Atlantic Hadley circulation). The quasi-decadal frequency also displays an out-of-phase relationship in the SSTs, north and south of the mean position of the intertropical convergence zone (ITCZ). This short-lived structure, lasting for approximately two years, creates the impression that a tropical SST dipole pattern is one of the characteristics of the quasi-decadal signal. All five frequency bands represent to some extent fluctuations of the NAO and are associated with tropical Atlantic Ocean warming (cooling) with different spatial evolution. The two interannual bands show opposite SST evolution to the south of the ITCZ i.e.; southeastward evolution from the western tropical Atlantic for the 3.5-year period and westward spreading from the eastern tropical Atlantic for the 4.4-year period. Moreover, a significant coherence (with a one year phase lag) is found between the SST time series along the equatorial Atlantic obtained from the 3.5-year period, and the SST time series in the NIÑO3 area in the Pacific. It is cautiously argued that the 3.5-year period is largely associated with the global El Niño Southern Oscillation (ENSO) phenomenon, while the evolution of the 4.4-year period depends more upon Atlantic local conditions.

1. Introduction

In the Pacific Ocean, where the El Niño Southern Oscillation (ENSO) phenomenon dominates the interannual variability, there is a generally understood scenario of coupled ocean-atmosphere tropical processes (Cane and Zebiak, 1985). Consequently, understanding and predicting the Pacific El Niño phenomenon have considerably improved in the last ten years (Cane et al., 1986; Barnett et al., 1988; Neelin et al., 1994; Chen et al., 1995). The relationship between sea surface temperature (SST) and sea level pressure (SLP) is well documented (e.g., Rasmusson and Arkin, 1985) and since then distinctions have been made between the global ENSO signal (Kawamura, 1994; Lanzante, 1996; Tourre and White, 1995), the decadal climate variability (Latif and Barnett,

1994) and the "ENSO-like decade-to-century variability" (Zhang et al., 1996). The Atlantic climate variability, on the other hand, is characterized by a strong basin-wide and broad-banded coherence between ocean and atmosphere anomalies. For example, in the equatorial Atlantic "conditions at times (1963, 1973 or 1984) resemble El Niño" (Philander, 1990) and a "significant part of observed variability can be described by an equatorial mode akin to ENSO" (Zebiak, 1993). In the North Atlantic, surface atmospheric circulation is dominated by the semi-permanent Icelandic Low and Azores High. The interannual fluctuations of these two cells are anti correlated and contribute to the North Atlantic Oscillation (NAO). The NAO displays quasi-biennial, quasi-decadal and multidecadal variability (Hurrell, 1995). Quasi-decadal to interdecadal variability has been identified in the Atlantic ocean-atmosphere system as well (Deser and Blackmon, 1993; Kushnir, 1994; Levitus et al., 1994; Houghton, 1996; Reverdin et al., 1997). Mehta and Delworth (1995) present evidence that SST in the tropical Atlantic Ocean displays low-frequency variability. These characteristics of the Atlantic climate spectrum may cause analysis of time and space-limited domain to be misleading.

Work on interannual variability in the Atlantic Ocean can be found in the seminal study of Bjerknes (1964), where large scale SST and corresponding SLP anomalies were first identified at the interannual time scale. Bjerknes suggested that the rapid year-to-year SST variability arises from ocean-atmosphere heat fluxes forced by changing winds. He also suggested that decadal or longer fluctuations are related to changes in the ocean circulation, particularly the subtropical gyre, in response to the long term changes of the atmospheric circulation (e.g., changes associated with the strength and location of the subtropical high). More recently Levitus (1989) was able to identify, from hydrographic data, a weakening in the strength of the subtropical gyre in the early 1970's compared to its state in the late 1950's. Deser and Blackmon (1993) and Kushnir (1994) using longer surface data sets, found patterns of correlated variability between the atmosphere and the ocean associated with low frequency physical mechanisms corroborating, in part, the results obtained by Bjerknes.

In general, regions with maximum sensible heat, latent heat and momentum exchanges are not always congruent in space and time (Budyko, 1982). Therefore, identifying processes associated with observed surface patterns of coherent low frequency ocean-atmosphere variability presents a serious challenge. It has been suggested that there is interannual correlation between the tropical and extra-tropical north Atlantic in both the atmosphere (Nobre, 1985) and the ocean (Lau and Nath, 1990). The robustness of these correlations with time, as well as their frequency dependence have yet to be fully tested. For example, Houghton and Tourre (1992) and Enfield and Mayer (1997) find that the leading modes of SST interannual variability north and south of the ITCZ are uncorrelated at zero time-lag. If these SST fluctuations are linked to trade wind circulation and associated changes in the strength and/or the position of the

Azores and Saint Helena Highs, then climate variability of the Atlantic is more complex than previously thought. The complexity in isolating climate signals is further enhanced since SST anomalies in the equatorial Atlantic Ocean are smaller than in the Pacific Ocean, and the equatorial cold tongue is much weaker (Philander, 1990; Zebiak, 1993; Jin, 1996).

Finally, when comparing climate variability in the Pacific and Atlantic Oceans, there is evidence of an inverse relationship between SLP in the eastern Pacific and the tropical Atlantic (Covey and Hastenrath, 1978), the Southern Oscillation and the SLP in the South Atlantic Ocean (Wolter, 1989) and El Niño and precipitation in northeast Brazil (Caviedes, 1973; Hastenrath and Heller, 1977; Enfield, 1996).

Most of the research involving the analyses of SST and SLP data sets from the Atlantic Ocean have focused on the North Atlantic region and it is only recently that the South Atlantic was studied (Venegas et al., 1998). In addition, almost all previous research has analyzed the data in the time domain generally by using principal component analyses or similar techniques. In an effort to better understand the spatial and temporal characteristics of Atlantic variability associated with band-limited signals, we depart from the time domain approach in this study and investigate the space-time variability of the Atlantic ocean-atmosphere system (80°N-30°S) using a joint SST and SLP analysis in the frequency domain over the Atlantic basin for a 136 year period (1856-1991). The data sets and the method used are described in section 2. The space-time evolution patterns of SST and SLP corresponding to the dominant frequencies are described in section 3. In section 4 we discuss the results in the context of previous work and also evaluate plausible mechanisms. Summary, remarks and conclusion are found in section 5.

2. Data and Method

From the recently gridded 136-year long (1856-1991) global SST and SLP data sets by Kaplan et al. (1998a, b), monthly subsets for the Atlantic domain on a 5° x 5°grid and 4° x 4° grid of SST and SLP, respectively are obtained. Because of the sparse coverage in the South Atlantic only data north of 30°S are analyzed in this study. Essentially the data-reduction by Kaplan et al. involves computing leading empirical orthogonal functions (EOF) from the most recent high quality data (UK Meteorological Office Global Ocean Surface Temperature Atlas or GOSTA for SST and the Comprehensive Ocean-Atmosphere Data Set or COADS for SLP). EOFs are then used for fitting a first order linear model of time transition. The optimal estimation is obtained using EOF projection of the analyzed field in order to obtain a "reduced space". From the estimation of the available data covariance patterns, the method fills gaps, corrects sampling errors and produces spatially and temporally coherent data sets.

The diagnostic method used for analysis is the frequency domain singular value decomposition (SVD) technique developed by Mann and Park (1994, 1996, hereafter MP94 and MP96, respectively). Time domain decomposition techniques such as principal component analysis (PCA) or multi-channel singular spectrum analysis (M-SSA) are best suited for broad band features. They are poor at isolating band limited signals that are spatially coherent, quasi-periodic or unstable on a slowly varying noise background. In such situations the frequency domain approach is optimal (MP94). Mann and Park (1998) compare these two approaches in great detail and also demonstrate the utility of the frequency domain approach using synthetic examples.

The paradigm that the Atlantic ocean-atmosphere system exhibits spatially coherent band-limited variability on a slowly varying noise background motivated our investigation of the joint SST-SLP variability using the frequency domain approach. The goal of this paper is to isolate dominant frequencies at which the Atlantic basin exhibits significant coherent variability, and subsequently examine the corresponding spatial patterns of SST and SLP anomalies. The latter will be in general small compared to the amplitude of climate variability in the Atlantic Ocean. However it should be kept in mind that the total amplitude, over any significant frequency band, would be an integration of the amplitudes from individual frequencies within that band. The method is briefly described below. For greater details regarding the method used, refer to MP94 and MP96.

The time series at each grid point are first transformed from the time domain to the spectral domain using the multitaper spectral method (MTM) (Thomson, 1982; Park et al. 1987). At each frequency (f) for each grid point series, a small number (K) of independent spectral estimates are computed using K orthogonal Slepian tapers. The tapers can be thought of as a kernel or wavelet function. In addition, the tapers, being orthogonal by construction, capture independent information. The tapers average over a half-bandwidth of pf_R centered on the frequency f . Here, $f_R = 1/N\Delta t$ (Δt is the sampling interval, equal to one month in this study and N is number of data points) is the Rayleigh frequency (the least resolvable frequency). Of the K tapers only the first $2p$-1 tapers are usefully resistant to spectral leakage. The parameters K and p provide spectral degrees of freedom and frequency resolution, respectively. A larger p implies averaging over a bigger bandwidth, and vice-versa. We choose p=2 and K=3, as in MP94 and MP96, which provides enough spectral degrees of freedom for a signal-to-noise ratio decomposition, while allowing reasonably good frequency resolution.

SST and SLP time series are standardized at each grid point. The anomalies are derived from long-term means (136 years) at each point, and the standard deviations are computed at each point as well. The grid point series are weighted with respect to latitude. At each frequency f, K independent spectral estimates of the standardized time series are calculated as:

$$Y_k^{(m)}(f) = \Sigma_{n=1}^{N} w^{(k)}_n x^{(m)}_n e^{i2pfn\Delta t}$$

where m=1,..., M are the grid points (for the concatenated SST and SLP fields); n=1,..,N are the time points; k=1,..,K are the number of spectral estimates and $\{w^{(k)}_n\}$ are the weights from the k^{th} data taper. Thus, we have K spectral estimates at each grid-point series, resulting in a matrix of size M x K at frequency f.

A complex SVD is performed on this matrix which gives K orthonormal left and right eigenvectors representing the spatial and frequency domains of EOFs of SST and SLP anomalies respectively, and K singular (or eigen) values representing the shared variance captured by each mode (or taper) within a given narrow-frequency band (or "local variance"). The fractional variance (or "local fractional variance") captured by each mode is then computed. The process is repeated for a number of frequencies, and a plot of the spectrum of fractional variance explained by the first mode at all the frequencies is readily obtained. Significance of peaks in the fractional variance spectrum are determined from a comparison with confidence limits obtained through a bootstrap procedure (Efron, 1990). Here the fields are permuted in time to keep the spatial structure intact. One thousand permutations are generated and in each case the local fractional variance at each frequency for all the K modes is derived. From this ensemble, the 99th, 95^{th}, 90^{th} and 50^{th} percentiles are computed. There are usually a number of peaks in a given frequency window (or cluster). Consequently, the significance at a given level is obtained from the averaged highest value of fractional variance within the broader spectral band, obtained by using a local polynomial smoothing procedure (Lall et al., 1998).

When the above analysis is performed in a moving time window, an estimation of the frequency variations over time is obtained. A 60-year moving window is used in this paper so that the evolution of the fractional variance spectrum is displayed. The left and right orthonormal eigenvectors are used to reconstruct spatial and temporal patterns at each frequency (refer to MP94 and MP96 for more details).

While the method used in this paper identifies joint SST-SLP dominant frequencies, the physical interpretation of the mechanisms associated with these frequencies can be cumbersome. This is why, prior to interpretation, the SVD joint spectrum is smoothed so that the spatial reconstructions are computed from the lead and well separated smoothed peaks.

3. Results

The spectrum of the first singular values, and its evolution for the whole time period, are displayed in Fig. 1. For the whole time period, significant peaks at the 95% level exist in three broad bands (Fig. 1a) the quasi-biennial band (skewed around a 2.7-year period), the interannual band (centered around periods of 3.5 and 4.4 years) and the quasi-decadal band (skewed around an 11.4-year period). A smoothed quasi-biennial time scale around 2.2-year period is also significant at the 90% level. A multidecadal peak (around 50-year period, significant at the 90% level) can also be seen in the unsmoothed curve of Fig. 1a.

The "evolutive" SVD spectrum (using a 60-year moving window) significant above the 90% level is presented in Fig. 1b. From approximately the turn of the century until the mid 30's there is a gap in the power of the period centered around 3.5 years. During that same period the power of the signal centered around the 4.4-year becomes more significant. MP96, in their joint SST-SLP analysis of the Northern Hemisphere also find significant spectral power in this time period. Other important features are the highly significant quasi-biennial and quasi-decadal signals around the 2.7-year and 11.4-year periods, respectively. The quasi-biennial signal dominates when the quasi-decadal signal is weaker and less significant. The quasi-biennial signal is conspicuous from 1920 to 1955 when the quasi-decadal signal is below the 90% confidence level. These results will be discussed in further detail in section 4e.

In the subsequent sections, joint spatial patterns of SST and SLP at the dominant frequencies, identified from Figure 1a, are described. The evolution of the spatial patterns through half their cycle is shown in six panels. It was decided, arbitrarily, to start the first panel (Figs. 2a, 3a, 4a and 5a) where the SLP anomalies contribute to a weakening of the Azores High. The patterns evolve through a "perfect" cycle; thus the seventh panel (not shown) is the same as the first panel with opposite polarity. The SSTs are shown in thin contours and shading and the SLPs are shown in heavy contours. The contour intervals are kept constant in all the figures for ease of comparison. The anomalies presented hereafter are in general small since they are computed over narrow bands, as explained in Section 2.

a. The quasi-biennial period

Two quasi-biennial spectral peaks are centered around periods of 2.2 years and 2.7 years (Figure 1a). The evolution of the spatial patterns at both these periods are very similar. Consequently only patterns corresponding to the 2.7-year period are shown in Fig. 2. The anomalies are from 0.2 mb (Fig. 2d) to 0.6 mb (Fig. 2b). The SLP anomalies form a dipole akin to the North Atlantic Oscillation (NAO). The maximum anomalous SLP occur during the part of the cycle in which weak, positive SST anomalies build up in the eastern equatorial ocean, in the regions where cold water usually dominate. The SST anomalies then expand westward along the equator and northward along the west African coastline. The tropical Atlantic Ocean is alternatively warmer or colder (up to 0.2°C) every 16 months. The maximum expansion of SST anomalies in the tropics (Fig. 2c) occurs 3 months after maximum negative SLP anomalies between Newfoundland and western Europe (Fig. 2b).

b. The interannual bands

Two distinct spectral peaks centered around 3.5-year and 4.4-year periods are identified at the interannual time scales. These frequencies capture more than 60% of the fractional variance in a narrow band centered around them (see Fig. 1a).

i. The 3.5-year period

Fig. 3 shows the evolution of the anomalies for half this period. Maxima of SLP anomalies in the North Atlantic form a dipole-like pattern akin to the NAO. The largest anomalies (0.6 mb) are identified first in the Denmark Strait and then west of the Azores (Figs. 3c and 3d). The anomalies persist for approximately half a year with little motion and then the pattern mentioned above collapses (Fig. 3e). The strongest anomalous SLP gradient is found between 45°N-55°N and 20°W-50°W (Figs. 3b and 3c). In the tropical Atlantic, SLP anomalies are weaker (0.1 mb).

In the north Atlantic Ocean, SST anomalies are rather weak despite the strong SLP gradient mentioned above (Fig. 3d). Weak SST anomalies (0.1°C) at 50°N and 30°W-40°W occur 3-6 months after the maximum anomalous SLP gradient. More specifically, positive SST anomalies peak (0.1°C to 0.2°C) between 50°N and 55°N, and between 10°N-20°N (Figs. 3d and 3e).

SST anomalies appear along the West African coastline, in the region where climatological features such as the cold Canaries current and coastal upwelling dominate (Fig. 3b). SST anomalies along the Senegalese coastline occur when SLP anomalies contribute to a weakening of the North Atlantic SLP dipole. From there the anomalies expand westward between approximately 10°N and 20°N, consistent with the path of the North Equatorial Current (NEC), and then southward between the West African coastline and 30°W. At 10°S a merging occurs with anomalies moving northward from the South Atlantic Ocean (Fig. 3e). Finally the anomalies find their way into the gulf of Guinea following the mean path of the North Equatorial Counter Current (NECC) and the Equatorial Undercurrent (EUC).

ii. The 4.4-year period

Fig. 4 shows the evolution of the spatial patterns associated with half this period. The largest SLP anomalies are identified between Iceland and the southern tip of Greenland (0.6 to 0.8 mb), and at 35oN between 35°W-65°W (0.2

mb, Fig. 4b), they persist for approximately 3/4 of the cycle with little motion. To the south between 25°N and 35°N and between 30°W and 50°W, the anomalies collapse (25°N, 40°W, Fig. 4e), a year later. In the western North Atlantic Ocean, maximum positive SLP anomalies are found over southern Greenland (Fig. 4c).

Positive SST anomalies (up to 0.3°C) appear in the 40°N-60°N zonal band, where the SLP gradient is such that westerly winds are weakened (Figs. 4c and 4d). This could reduce storm activity during winters occurring during the cycle. Along 30oN-40oN negative SLP anomalies are found 15-20 degrees of longitude east of the negative SST anomalies (Figs. 4c and 4d). Peak SST anomalies lag peak SLP anomalies by several months. Contemporaneously, positive SST anomalies (0.1°C to 0.3°C) are the dominant feature in the North Tropical Atlantic Ocean (Figs. 4c, 4d and 4e) where northeasterly winds must be weaker. In the South Tropical Atlantic, negative (positive) SST anomalies appear off-shore of Central Africa (Gabon, Congo and Angola, Figs. 4c). Subsequently, these anomalies expand westward (Figs. 4d and 4e).

c. The quasi-decadal period

The prominent quasi-decadal peak around the 11.4-year period can be seen in Figure 1a. The fractional variance captured at this frequency is significant at the 99% level. The spatial patterns associated with the quasi-decadal band are displayed in Fig. 5.

In the North Atlantic the SLP anomaly patterns are reminiscent of the NAO pattern. An out-of-phase relationship also occurs between the areas where the subtropical anticyclones in both hemispheres are found the Azores High to the north and the Saint Helena High to the south. The difference in signs between the SLP anomalies to the north and the subtropical Highs leads to the conclusion that quasi-decadal variability is also present in the winds at all latitudes. In Fig. 5 where the maximum SLP anomalous gradient is visible (Figs. 4a, 4b and 4c), it can be inferred that when the westerlies between 50°N and 65°N are weaker than normal, the northeasterly trades are also weaker than normal. At the same time the southeasterly trades are stronger than normal.

An in-phase relationship between SST anomalies in the 50°N-60°N zonal band and the North Tropical Atlantic Ocean (north of the mean position of the ITCZ) is readily seen (Figs. 5b, 5c and 5d). Furthermore, a coherent east-northeast oriented pattern of SST anomalies is found just south of the Gulf Stream extension and along the North Atlantic Current (NAC) (Figs. 5a to 5f). SST anomalies are also found in the vicinity of the North Atlantic Drift (Fig. 5a with reversed polarity).

SST anomalies in the North Atlantic Ocean, north and south of the Azores High have the same polarity. In contrast SST anomalies have opposite polarity north and south of the mean position of the ITCZ (Fig. 5d). This distribution of SST anomalies fits well with the in-phase relationship between mid-latitude westerlies and northeasterly trade winds and the out-of-phase relationship of Atlantic northerly and southeasterly trade winds on this time scale, as previously discussed. The evolution of tropical SST during the decadal cycle is thus marked by the appearance of a seesaw pattern of SST during the growth phase of the pattern.

4. Discussion

All spatial patterns corresponding to the four frequencies begin arbitrarily (Figs. 2 through 5), with a weaker than normal Azores anticyclone and a weaker than normal Icelandic Low. All of these SLP anomalies contribute to a negative NAO index at the beginning of each cycle. In this section we compare our results with relevant previous work, and also discuss possible physical mechanisms involved with the dominant signals that we have identified.

a. The quasi-biennial period

Wagner (1971), in an analysis of a 66-year Northern Hemisphere SLP data set (north of 20°N), and other investigators (e.g., Landsberg et al., 1963), identified tropospheric quasi-biennial signals or "pulses". Wagner concluded that the spectral power in a band centered at a period of around 2.5-year displayed maximum amplitude in both SST and SLP. The maxima associated with the spectral band were located near the centers of the Icelandic Low and the Azores High during winter. These results were corroborated later by Angell and Korshover (1974) who also argued that the quasi-biennial fluctuations and their different frequencies are linked to the southwest-northeast movement of the "centers of action", i.e., the Icelandic Low and the Azores High. They showed that the Azores High becomes weaker when displaced southward. A quasi-biennial oscillation was subsequently identified in the temporal record of the NAO by van Loon and Rogers (1978). More recently quasi-biennial signals were identified in SLP over the Northern Hemisphere (Trenberth and Shin, 1984), SLP, surface wind and air temperature data from the north Atlantic Ocean (Gordon et al., 1992; Deser and Blackmon, 1993) and from global analysis of air temperature and SST (MP94).

In our analysis we find two spectral peaks in the vicinity of the 2.5-year period. The evolution of SLP anomalies in the North Atlantic, is consistent with the results obtained by Angell and Korshover (1974). The contribution of the biennial signal to SLP variability there, is particularly evident after ~1920 (Fig. 1b). The shape and evolution of the patterns of tropical SST anomalies from the Angolan coastline (Figs. 2a to 2d) seems to indicate that the anomalous local currents (e.g., the South Equatorial Current, or SEC) play a preponderant role. It is interesting to note that Konda et al., (1996), using band-pass filtered biennial variations of SST obtained from satellite data, concluded that SST evolution in the tropical Atlantic Ocean is primarily controlled by lateral heat transport and not air-sea heat fluxes. Negative SLP anomalies from 60°N to 30°S, are associated with a warmer tropical ocean. The fact that the maxima in SLP and SST anomalies are far apart deserves further investigation.

b. The interannual band

During the last two decades there has been considerable evidence of a relationship between the negative phase of the SOI and equatorial Atlantic warm events (Covey and Hastenrath, 1978; Hastenrath et al., 1987; Wolter, 1989 among others). Modeling experiments have established links between the intensity of the tropical Atlantic trades and the ENSO phenomenon (Tourre et al., 1985; Carton and Huang, 1994; Delecluse et al., 1994). Zebiak (1993) discussed the importance of meridional SST gradients in the Atlantic domain and its association with tropical Atlantic Ocean circulation. Curtis and Hastenrath (1995), from a composite study, describe the evolution of meridional SST gradients in the tropical Atlantic Ocean. Wagner (1996) proposed physical mechanisms associated with the tropical meridional SST gradient. Spectral peaks around 5-year period were found by Venegas et al. (1998) in the South Atlantic from a joint analysis of SST-SLP in the time domain, based on the most recent 50 years of data. They associate this variability with a meridional displacement of the ITCZ and the Saint Helena anticyclone (the Southern Hemisphere subtropical high). Sperber and Hameed (1993) identify a 3.6 year period in the North tropical Atlantic SST, which they argue is one of the time scales at which the Walker circulation interacts with the tropical Atlantic circulation and modulates the track of the Atlantic ITCZ. Peterson and White (1997) identified an Antarctic Circumpolar wave, with a period of 4-5 years, transmitting climate anomalies in the Southern Oceans including the South Atlantic Ocean.

i. The 3.5-year period

For the 3.5-year period it is found that, in the North Atlantic, SST anomalies are rather weak even in the presence of large SLP anomalous gradient. The SLP anomalies are such that northeasterly trade winds are weaker in the North Tropical Atlantic Ocean and along the northwest African coastline. When a similar analysis is performed globally (not shown) we found that for approximately the same 3.5-year period, the patterns displayed in Fig. 3a corresponds to the time when a fully developed El Niño exists in the eastern Pacific. The southeasterly trade winds in the equatorial Atlantic will be at their peak strength, the negative SST anomaly being the largest in the equatorial Atlantic. The ITCZ will then be found farther north with weaker coastal upwelling and reduced heat loss between 10°N-20°N (Curtis and Hastenrath, 1995). At the equator and at the beginning of the arbitrarily defined cycle (Fig. 3a), negative SST anomalies and weak positive SLP anomalies are observed. Tropospheric subsidence and hydrostatic adjustment of the atmosphere to cold SST (Covey and Hastenrath, 1978; Lindzen and Nigam, 1987) will both increase SLP. The positive SST anomalies in the South Atlantic Ocean (at 30°S) expand equatorward following geostrophic flow patterns of the upper ocean (Reid, 1989; Stramma, 1991) and merge (west of 20°W) with positive SST anomalies from the north. These anomalies are reinforced by Ekman downwelling in those regions (Curtis and Hastenrath, 1995). The merging on the western tropical basin might be due to a tendency for the ITCZ to create, farther east, an initial potential vorticity barrier which inhibits direct flow along the thermocline layer from subtropical region towards the equator (Lu and McCreary, 1995). From there they propagate east-southeastward (Figs. 3a, 3b and 3c, all with reverse polarity). This anomalous propagation follows very closely the anomalous gradient of the sea-level topography computed from the oceanic geopotential thickness for the 0-1000 m layer (Levitus and Oort, 1977). A similar movement is noticed in the heat storage anomalies associated with the ENSO phenomenon (Tourre and White, 1998).

The tendency for a warmer North Tropical Atlantic Ocean during ENSO peaks, as in the 1980's, is seen in the global rotated EOF analysis of Tourre and White (1995). To gain further insight about this relationship, we compute the coherence and phase between the SST NIÑO3 time series (representative of the Pacific ENSO) and reconstruct time series averaged over grid points along the equatorial Atlantic. We find the largest significant coherence of 0.65 (99% level of confidence) at around 12 month lag, for the 3.5-year period. This seems to suggest, for example, that the 3.5-year signal contributes to a weakening of the Azores High (Figs. 3a to 3d) and that about one year after a fully developed El Niño in the eastern Pacific, a warming of the equatorial Atlantic Ocean is noticeable, with maximum SST anomalies (up to 0.2°C-0.3°C) occurring in the Gulf of Guinea (Hisard, 1980). This corroborates well with results by Philander (1990), Tourre and White (1995) and Latif and Barnett (1995). During the Atlantic warm event of 1984, after the 1982-83 "ENSO of the century" in the Pacific Ocean, a South Equatorial Counter Current (SECC) was monitored (Hisard and Henin, 1987). An eastward propagation of SST anomalies was observed during the Atlantic warm event of 1984 (Henin and Hisard, 1987; Reverdin and McPhaden, 1986). It was also concluded that during this period positive SST anomalies between the equator and 10°S and east of 20°W were indicative of a weaker SEC and a weaker Benguela Current system (Carton and Huang, 1994; Philander, 1986; Reverdin at al, 1991a).

The above arguments reinforce the idea of a specific Atlantic response to the Pacific ENSO signal as transferred through the "atmospheric bridge" as suggested by Lau and Nath (1996). The reason for the signal to be transmitted at that particular frequency and the potential role played by the atmosphere in the middle latitudes remain unclear.

ii. The 4.4-year period

The evolution of the negative SST anomalies south of the ITCZ indicates the presence of stronger southeasterly trade winds (Figs. 4c and 4d). The lack of SLP data south of 30°S does not permit to isolate a definite joint SLP anomaly pattern there. SLP anomalies in the vicinity of the Azores High are negative but they are somewhat weaker when compared with SLP anomalies for the 3.5-year period in the same region (Figs. 3c and 4c). South of the Azores High weaker northeasterly and stronger southeasterly, compared to the 3.5-year period, further tend to displace the ITCZ northward. As a result of the stronger local meridional winds, coastal upwelling gets enhanced along the Angolan coastline and is extended far westward (Shannon et al., 1986) while the strength of the Angola Gyre (Gordon and Bosley, 1991) tends to be reduced. When the southeasterly trade winds are relaxed, a warming occurs in the same region primarily due to heat redistribution by the SEC and SECC (Philander and Pacanowski, 1981). Finally SST anomalies are found farther north in both hemispheres in regions where there is also maximum Ekman drift in the mean (Mellor et al., 1982, Arnault, 1987). Similar conclusions from modeling studies were argued by Zebiak (1993) and Carton and Huang (1994).

c. Comparisons of the two interannual periods

From the way the interannual cycles are reconstructed in this paper and from a global analysis (not shown and already mentioned in section 4b; i), it can be deduced that the atmospheric linkage between the eastern Pacific and Atlantic Oceans is increased during the prominent 3.5-year period of the ENSO phenomenon. This is, for example, when negative SLP anomalies are found in the eastern Pacific, negative SLP anomalies are found in the vicinity of the Azores High and the SLP increases in the equatorial and South Atlantic through presumably tropospheric subsidence. Interestingly enough, when we performed additional spectrum analyses of the winter NAO and its two components, the Azores High and the Icelandic Low, a significant interannual signal (90% confidence level) is centered around 3.5 years. The linkages between the eastern Pacific SLP anomalous field associated with ENSO and the intensity of the Azores High could be associated with "an equivalent barotropic wave-like response in the North Pacific-North American" region as mentioned by Lau and Nath (1996). Lau and Nath were also able to associate warm ENSO events with a dipole-like SST pattern in the western North Atlantic (negative SST anomalies along the northeastern seaboard of North America associated with negative anomalies in latent and sensible fluxes). More work remains to be done is this domain.

The SST evolution in the tropical Atlantic is different for the 3.5 and 4.4-year periods. In the southeast Atlantic, SST anomalies propagate and/or expand in opposite directions, i.e., southeastward for the 3.5-year period and northwestward for the 4.4-year period. A fundamental difference between the two cycles is that colder equatorial SST are associated with higher tropical SLP for the 3.5-year period, while for the 4.4-year period this relationship breaks down. Since the joint SST-SLP evolution north of the ITCZ display similarities between the two cycles, the difference in the SST evolution south of the ITCZ must be linked to independent local atmospheric and oceanic conditions in the South Atlantic Ocean.

It is well-known that the position of the ITCZ in the tropical Atlantic is associated with local conditions such as the intensity of the northwest African heat low and the strength and position of the Saint Helena anticyclone. In the ocean these changes correspond to changes in the strength of the south subtropical gyre. The position of the ITCZ displays seasonal, interannual and lower frequency variability. For example when spectral analysis is performed on time series which depict the anomalous location of the ITCZ, the two intermediate frequencies discussed in this paper are clearly identified (not shown). Remote forcing, including ENSO, also influences the position of the Atlantic ITCZ (Enfield and Mayer, 1997). It has been shown that the latitude of the ITCZ is determined by the meridional SST gradient, the meridional SLP gradient and the relative intensity of the trades, northeasterly versus the southeasterly (Covey and Hastenrath, 1978; Hameed et al., 1993; Wagner, 1996; Wainer and Soares, 1997). We propose that the differences in the SST evolution south of the ITCZ between the 3.5 and 4.4-year periods reside in the relative difference in the strength of the trade winds system and the changes in the oceanic circulation (Peterson and White, 1997) due to both local and remote forcing. For both cycles when the NAO is weaker the northeasterly trade winds are weaker. Accordingly the ITCZ is found further north. Furthermore when El Niño is fully developed there is an indication of positive equatorial SLP anomalies (Figs. 3a to 3c). For the 4.4-year period (Fig. 4) weak positive SLP anomalies are found further south (Figs. 4d and 4e) and could be associated with a stronger and/or northward displacement of the Saint Helena anticyclone. This movement has been mentioned by Venegas et al., (1998). Resulting equatorial warming is then influenced by local conditions with weaker southeasterly trade winds. From the south, SST anomalies could also penetrate the tropics by entering the Benguela Current system and subsequent northward spreading from the SEC. Finally the possible link with the 4-5 year period of the Antarctic Circumpolar Wave (ACW) and northward Ekman layer flow propagation identified by White and Peterson (1996) and Peterson and White (1997) deserves further investigation.

d. The quasi-decadal period

The Atlantic climate variability is characterized by a strong coherence associated with this cycle. Deser and Blackmon (1993) find that SST quasi-decadal variability is associated with the Western Atlantic oscillation (WA), a geopotential pattern in the upper troposphere (Wallace and Gutzler, 1981). Molinari et al., (1997) were able to identify the same time scale variability in the subsurface temperature (down to 400m) in the mid-latitudes of the western north Atlantic Ocean. From coupled-model runs, Grotzner et al., (1998) identify low-frequency variability of around a 17-year period in their joint SST-SLP of the Atlantic basin, and attribute it to unstable air-sea interactions involving the subtropical gyre and the NAO. In their experiment the ocean responds to low-frequency wind stress curl variations and serves as a memory in the coupled system. The intensity of the Gulf Stream extension and the subtropical gyre respond accordingly and keep the coupled system oscillating. Similar findings were made by Hansen and Bezdek (1996) and Halliwell, Jr. (1995). North of 30°N the SST anomalies closely follow the mean transport stream function derived from wind-stress and subsurface density fields (Mellor et al., 1982). The amplification of SST anomalies to the north of 50°N, the region where the barotropic flow associated with the subpolar gyre (Hogg et al., 1986) separates from the Sargasso gyre. Sutton and Allen (1997) find a similar evolution in SST anomalies when they analyzed SST for a shorter time-period (1945-1989).

In our joint SST-SLP analysis, the spatial distribution of SLP anomalies is a reminder of the NAO standing index characteristics. The joint SST and SLP patterns associated with an 11.4-year period (as shown in this paper) points to a strong middle latitude NAO-SST linkage and a strong tropical-extra tropical connection at decadal time scales. This result emphasizes a time scale defined by the ability of upper ocean anomalies to persist, even after the atmospheric anomalies decay, and during which ocean-atmosphere interaction is maintained on the Atlantic basin scale.

Hurrell (1995), Halliwell (1995) and Kushnir (1996) suggest that positive (negative) SST anomalies in the westerlies and trade-wind belts appear simultaneously when the anomalous SLP gradient is decreased (increased). This is corroborated by the relative distribution of SST and SLP anomalies in Fig. 5.

In the tropical Atlantic, quasi-decadal variability is associated with an out-of-phase relationship in the SST to the north and south of the ITCZ. This is consistent with previous analyses from Houghton and Tourre (1992) and Servain (1991) who identified a decadal time scale for a SST tropical "dipole index". Mehta and Delworth (1995) mention that SST variability south and north of the ITCZ occurs at time scales between 8 and 11 years. Recently, Chang et al. (1997), using a hybrid coupled general circulation model (HGCM), show that a dipole structure with a decadal time scale can be attained through unstable thermodynamics between wind induced heat fluxes and SST anomalies. Northern and southern tropical components of the SST anomaly structure might be associated with the same mechanisms, but a true dipole structure is retained only for a relatively short time period (for at most 2 years) in the quasi-decadal evolution. When SVD and cross-spectral analyses are performed, with the NAO and subtropical SST normalized time series, significant coherence is found north and south of the equator. Nonetheless the cross spectral analysis between SST in the two regions identified in Fig. 5d reveals very little coherence (Rajagopalan et al., 1998). This leads us to believe that SST in the two subtropical regions are independently coherent with NAO (Kushnir et al., 1998).

The trade winds are out-of-phase on this time scale since the subtropical anticyclones are also out-of-phase. The meridional SST gradient across the ITCZ associated with the decadal pattern has a profound effect on the Northeast Brazil and Sahelian rainfall (Hastenrath and Heller, 1977; Mechoso et al., 1990; Sperber and Hameed, 1993; Wainer and Soares, 1997).

e. Comparison of quasi-biennial and quasi-decadal periods

As shown in Fig. 1b, the statistical significance of the quasi-biennial and quasi-decadal bands vary with time. This was first mentioned by Kutzbach (1970). In particular, we can note from this figure that, when the quasi-biennial period was significant during the 1920-1955 period, the quasi-decadal period was much less prominent. The interannual period, especially the 3.5-year signal also seems to be less significant during this period. Gu and Philander (1995) note that between 1915 and 1950 the intensity of the Southern Oscillation was relatively small and increased rapidly thereafter. Similar observations were found by Rajagopalan et al. (1997) in their analysis of Darwin SLP data. The relationship, as a function of time, between the quasi-biennial and quasi-decadal periods is still unclear. According to Hurrell and van Loon (1996), the NAO "standing index" has exhibited considerable variability at quasi-biennial and decadal time scales during the past 130 years. This is not the case in Fig. 1b, where the biennial signal is not significant prior to 1920. They also show that the NAO decadal variability becomes more pronounced after the 1950's. Similarity and differences between their results and ours could be due to several facts: 1) our study emphasizes the evolution of joint (both SST and SLP) spatial patterns as opposed to amplitudes of individual standing indices such as the NAO index for the North Atlantic only; 2) our analysis emphasizes joint coherent frequency bands, and not necessarily spectral power, in which the power is shared between two fields over the entire Atlantic domain; 3) Our analysis is based on monthly data as opposed to the winter time values of the NAO index used by Hurrell and van Loon (1996). Indeed, a moving-window spectrum on monthly time series of the NAO index (not shown) revels a prominent quasi-decadal band prior to ~1920 as in Fig. 1b. We can infer from Rogers (1984) and Hurrell (1995) that between ~1920 and ~1960 there was a downward trend in the NAO index. Also, during the 1920-1960 interval both the Azores High and the Icelandic Low moved southwestward (Glowienka, 1985). During that period the SLP over the Greenland-Iceland regions gradually increased (Rogers,1984) and the winter storm tracks in the North Atlantic were found further south (Tinsley, 1988). This year-interval is roughly related to the north Atlantic "warm period" identified by Kushnir (1994). The physical link between these observations and the concomitant changes in frequency characteristics of the joint SST-SLP patterns is not clear at this time. In a joint SST-SLP spatio-temporal analysis for the North Hemisphere, MP96 found that the quasi-decadal signal appeared as a distinctly cold-season phenomenon while the quasi-biennial signal is significant during both independent warm and cold seasons and stronger during the cold season. The effect of seasonality, while studying tropical-extratropical interactions for the entire Atlantic Ocean, needs further research.

In this paper it is shown that there is a relationship between localized SLP variations and dominant cycles in the joint SST-SLP variability in the North Atlantic. Generally, during the period when the NAO index diminishes, the quasi-biennial cycle is enhanced, which indicates a possible quasi-decadal modulation on shorter time-scale variability. It is also believed that while NAO variability is tightly linked with SST variability (e.g., Kushnir, 1994), the biennial variability as described in this analysis is largely atmospheric (compare the amplitudes of SST and SLP anomalies in Fig. 2). This is beyond the scope of this paper.

5. Summary

Five significant frequency bands for the joint SST-SLP evolution are identified in the Atlantic basin. They range from the quasi-biennial to the quasi-decadal. Two quasi-biennial bands are centered around 2.2 and 2.7-year periods; two interannual bands are centered around 3.5 and 4.4-year periods ; the fifth band at the quasi-decadal frequency is centered around 11.4-year time period. The spatial patterns corresponding to the five periods display large SLP variability in the North Atlantic. With the quasi-decadal period, the SLP patterns and their time evolution are not only linked with the NAO but also with an out-of-phase relationship between the subtropical anticyclones in both hemispheres. This indicates a dominant quasi-decadal variability in the Atlantic Hadley circulation.

The five significant frequency bands are all associated with tropical warming with different intensities and evolution along the equatorial wave guide. For the quasi-biennial bands the entire tropical Atlantic gets simultaneously warmer (colder). For the other three bands there are out-of-phase relationships north and south of the mean position of the ITCZ. For the quasi-decadal band, this relationship leads to a short-lived dipole indicative of an independent SST time-evolution north and south of the ITCZ. The 3.5-year band displays eastward evolution along the equator and seems to be linked to the Pacific ENSO through the atmosphere (warm/cold Atlantic events lag the Pacific by 15 to 18 months). The warm (cold) events associated with the 4.4-year band are weaker, propagate westward along the equator and seem to be depending upon local Atlantic conditions such as the locus of the ITCZ.

From a modeling study, Grotzner et al. (1996) suggested that the Atlantic decadal climate cycle is a coupled ocean-atmosphere phenomenon as it is in the

Pacific (Latif and Barnett, 1994; Tourre et al., 1998). The apparent difference between the two low-frequency signals in the Atlantic and the Pacific must then be attributed to the size difference of the two basins and the dominance of the NAO in the North Atlantic. The latter implies a potential for atmospheric feedback. It is interesting to note that while Wohlleben and Weaver (1995) identified an interdecadal signal in the SST of the subpolar north Atlantic gyre, Houghton (1996) found a quasi-decadal fluctuation in the upper portion of the Labrador Current where vertical mixing is weaker due to freshwater input from the Arctic (Reverdin et al., 1977).

The relationship between the dominant joint evolution of SST-SLP and their appropriate frequencies requires further investigation (with additional data sets including subsurface temperature, salinity and precipitation). For example, preliminary reconstruction of the joint SST-SLP patterns corresponding to the 50-year peak in Fig.1b (not shown) seems to be consistent with the results from Delworth et al. (1993) and Kushnir (1994), suggestive of variability in thermohaline circulation as a plausible driving mechanism at this time scale. Isolating the main physical mechanisms involved by comparing results from diagnostic analyses with modeling output (reproducibility) will have immediate predicting application (Pittalwala and Hameed, 1991). Nevertheless complexity could arise since the different signals (their relative intensity, stability and evolution) modulate each other, as evidenced in this paper, when the quasi-biennial and the quasi-decadal signals are compared (Fig. 1b). Finally more insight can be gained on the mechanisms associated with the patterns presented in this study by exploring the role of seasonality.

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Figure 1a. Spectrum of the fractional variance as a function of cycle-per-year explained by the first MTM/SVD joint SLP-SST mode. The thicker solid line represents a local polynomial smoothing of the frequency distribution after Lall et al., (1998). Dashed horizontal lines represent the 99%, 95%, 90% and 50% confidence limits obtained through bootstrapping procedures after Efron (1990).

Figure 1b. Frequency variations over time from MTM/SVD analysis performed in a 60-year moving time window. Only the fractional variance spectrum above the 90% confidence limit is shown in Figure 1b. The 90% to 95% confidence limit is in yellow, while the 95% confidence limit and above is in red.

Figure 2. Space and time evolution (top to bottom and left to right, labeled a to f) of the 2.7-year period for both SST and SLP anomalies. Only half the cycle is presented. There is a 2.7-month time difference between each frame. Thin contours (solid and dashed) represent SST anomalies (positive and negative) every 0.1°C, the zero-contour line being omitted. Shading is for values larger than 0.1oC. Thick contouring (solid and dashed) represents SLP anomalies (positive and negative) every 0.2 mb.

Figure 3. Space and time evolution (top to bottom and left to right, labeled a to f) of the 3.5-year period for both SST and SLP anomalies. Only half the cycle is presented. There is a 3.5-month time difference between each frame. Contouring convention is the same as for Fig. 2.

Figure 4. Space and time evolution (top to bottom and left to right, labeled a to f) of the 4.4-year period for both SST and SLP anomalies. Only half the cycle is presented. There is a 4.4-month time difference between each frame. Contouring convention is the same as for Fig. 2.

Figure 5. Space and time evolution (top to bottom and left to right, labeled a to f) of the 11.4-year period for both SST and SLP anomalies. Only half the cycle is presented. There is a 11.4-month time difference between each frame. Contouring convention is the same as for Fig. 2.