A Decadal Climate Cycle in the North Atlantic Ocean as Simulated by the ECHO Coupled GCM

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ABSTRACT

In this paper a decadal climate cycle in the North Atlantic that was derived from an extended-range integration with a coupled ocean–atmosphere general circulation model is described. The decadal mode shares many features with the observed decadal variability in the North Atlantic. The period of the simulated oscillation, however, is somewhat longer than that estimated from observations. While the observations indicate a period of about 12 yr, the coupled model simulation yields a period of about 17 yr. The cyclic nature of the decadal variability implies some inherent predictability at these timescales.

The decadal mode is based on unstable air–sea interactions and must be therefore regarded as an inherently coupled mode. It involves the subtropical gyre and the North Atlantic oscillation. The memory of the coupled system, however, resides in the ocean and is related to horizontal advection and to the oceanic adjustment to low-frequency wind stress curl variations. In particular, it is found that variations in the intensity of the Gulf Stream and its extension are crucial to the oscillation. Although differing in details, the North Atlantic decadal mode and the North Pacific mode described by M. Latif and T. P. Barnett are based on the same fundamental mechanism: a feedback loop between the wind driven subtropical gyre and the extratropical atmospheric circulation.

1. Introduction

The climate of the earth exhibits natural variability on a broad range of timescales. The variability on timescales ranging from years to centuries is attracting increasing attention, especially the investigation of possible anthropogenic impacts on global climate requires an advanced understanding of the natural variations on decadal and interdecadal timescales. These variations, however, are difficult to understand from observations, since instrumental climate records cover only the last hundred years. Hence, the mechanisms responsible for longer term variability are only poorly understood. But improvements in the analysis of the available data and especially the new possibility to perform relatively long experiments with comprehensive coupled ocean–atmosphere models make it possible to obtain more insight into the dynamics of longer term climate variability. In this study we present results from such a coupled ocean–atmosphere model that contribute to the understanding of the decadal variability of the North Atlantic Ocean.

Observational studies show that the North Atlantic Ocean exhibits considerable variability on interannual, decadal, and interdecadal timescales. Bjerknes (1964) described variations of the North Atlantic sea surface temperatures (SSTs) and corresponding atmospheric conditions. The relation between SST and sea level pressure (SLP) anomaly patterns led Bjerknes (1964) to the conclusion that interannual SST variability is forced by the atmosphere through air–sea flux variations. More recent studies have essentially confirmed these findings (Wallace et al. 1990; Zorita et al. 1992; Delworth 1996).

The patterns of interdecadal variability, however, are not directly linked to local air–sea interaction, and ocean dynamics seems to be important on these timescales. This supposition originally mentioned by Bjerknes (1964) has also been supported by more recent investigations. By applying different analysis techniques, Deser and Blackmon (1993) and Kushnir (1994) investigated observational records going back to the beginning of the century. They describe interdecadal variations of SST with large-scale patterns that are characterized by a uniform sign over almost the entire North Atlantic basin. The temporal evolution is characterized by a cold

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period at the beginning of the century, a rapid warming around 1930, a warm period that prevails up to the 1960s, and a following colder period that might have been terminated in the 1980s. An enhancement of the prevailing westerly winds over the positive SST anomalies does not support the picture of local air–sea interactions and indicates an active role of the ocean.

Moreover, Deser and Blackmon (1993) present evidence for a quasi-decadal cycle within the North Atlantic. This mode has a structure similar to the interannual variability, with temperature anomalies of one sign east of Newfoundland and anomalies of the opposite sign off the southeast coast of the United States. The relationship to the atmospheric anomalies is similar to that found on interannual timescales, with stronger (weaker) winds overlying colder (warmer) SSTs. This mode exhibits a significant spectral peak at approximately 12 yr in the postwar period. Before 1945 the fluctuations were less regular with a dominating period of approximately 9 yr. The occurrence of significant peaks within the red spectra (which are characteristic for many climate variables) is a strong indication for the existence of oscillatory modes. Such oscillations might result from some feedback mechanism within the coupled climate system of the North Atlantic, as shown below. Results of Levitus et al. (1994) and Molinari et al. (1997) obtained from the analysis of subsurface data, which are much less contaminated by high-frequency noise than surface data but only available for the postwar period, essentially confirm the findings of Deser and Blackmon (1993) for the time after 1945.

Besides external forcing, like variations of solar irradiance or volcanic activity, several mechanisms have been proposed to explain the climate variability of the North Atlantic. At least on interannual timescales, observational results show that atmospheric anomalies tend to force the variations in the ocean by anomalous heat fluxes and surface mixing (Wallace et al. 1990; Zorita et al. 1992). Such an active role of the atmosphere can be explained within the framework of the stochastic climate model (Hasselmann 1976). Weather noise, usually characterized by white spectra, is integrated by the much slower responding ocean, leading to spectra that are essentially red down to frequencies where the system is stabilized by some internal feedback mechanisms. Such a mechanism is able to explain many aspects of the interannual climate variability in the midlatitudes (Frankignoul and Hasselmann 1977; Frankignoul 1985; Luksch et al. 1990). The concept of the stochastic climate model has also been extended to decadal timescales by Frankignoul et al. (1997). They suggest that white noise wind stress forcing might be integrated by baroclinic Rossby waves within the wind-driven ocean gyres, resulting in red spectra up to decadal timescales. Such a mechanism, which does not yield preferred timescales, seems to be responsible for the decadal extratropical variability in a long-term integration with the Hamburg coupled ocean–atmosphere general circulation model ECHAM1-LSG (Robertson 1996; Zorita and Frankignoul 1997). It is important to mention in this context that the atmosphere might be able to create some low-frequency variability by its own by internal nonlinear feedback processes (e.g., James and James 1989).

A class of inherently coupled air–sea modes exists also, which might be either self-sustained or stochastically forced. An example of such inherently coupled oscillations is the El Niño–Southern Oscillation (ENSO) phenomenon, which originates in the tropical Pacific (e.g., Neelin et al. 1994). Especially at long timescales of decades and more the phase relationships between ocean and atmosphere are not clear, and it has been argued that in this frequency range the ocean drives the atmosphere (e.g., Bjerknes 1964; Kushnir 1994). Such an active role of the ocean has usually been proposed within frameworks that involve the thermohaline driven meridional overturning cell of the North Atlantic Ocean. Theoretical considerations indicate that multiple equilibria and self-sustained oscillations of the overturning cell are possible in the Atlantic Ocean (e.g., Stommel 1961). Various experiments with oceanic circulation models have shown variations on timescales ranging from decades to centuries that are related to anomalous freshwater forcing and deep water formation (e.g., Weaver and Sarachik 1991; Mikolajewicz and Maier-Reimer 1992; Weisse et al. 1994). Delworth et al. (1993) found an oscillation of approximately 50-yr period in a multicentury integration of a coupled ocean–atmosphere general circulation model. The mechanism responsible for this mode seems to be mostly due to processes within the ocean and involves the thermohaline meridional overturning in conjunction with anomalous upper-ocean heat and salt transports caused by an anomalous gyral circulation.

Here, we present an alternative hypothesis to explain part of the decadal variability in the North Atlantic. This type of variability is not related to the ocean or the atmosphere alone, but arises from two-way interactions between ocean and atmosphere, and it must be therefore regarded as an inherently coupled air–sea mode. It involves both the subtropical oceanic gyre circulation and the atmosphere as active components and is based on unstable large-scale ocean–atmosphere interactions in midlatitudes. The possibility of unstable air–sea interactions was originally hypothesized by Namias (1959, 1969) for the North Pacific and by Bjerknes (1964) for the North Atlantic. Latif and Barnett (1994, 1996) presented results from a century-long integration with a complex coupled general circulation model (GCM), which simulated an oscillation of approximately 20-yr period in the North Pacific. They were able to explain the oscillation consistently by unstable coupled ocean–atmosphere dynamics. Intensity anomalies of the Pacific subtropical gyre cause SST anomalies along the Kuroshio extension. The response of the atmosphere to these SST anomalies comprises changes at the air–sea interface that reinforce the original temperature anomalies so that ocean and atmosphere form a positive feedback system. The atmospheric response consists also of a wind stress curl anomaly that changes the strength of the subtropical gyre in a way opposite to the initial state. The ocean adjusts with some time lag, hence providing a delayed negative feedback and enabling the coupled system to oscillate.

In this paper we shall show that in the same numerical experiment the North Atlantic exhibits also a decadal oscillatory mode. The spatial patterns related to this oscillation are very similar to those of the quasi-decadal cycle derived by Deser and Blackmon (1993). Similar to the North Pacific, the mechanism responsible for the Atlantic mode is based on an interaction of the subtropical gyre and the North Atlantic atmospheric circulation. The latter accounts for a large amount of the model's North Atlantic decadal variability. In section 2 the coupled GCM and the setup of the experiment will be briefly described. Section 3 shows the performance of the model with respect to the mean state and interannual variability. The structure of the simulated decadal oscillation in comparison to observations is described in section 4, the underlying physical mechanism is discussed in section 5. A discussion concludes the paper.

2. The coupled general circulation model ECHO and experimental design

The numerical model applied to investigate the decadal variability of the North Atlantic is the coupled general circulation model ECHO (Latif et al. 1994). The model consists of the atmospheric component ECHAM3 and the ocean model HOPE. ECHAM3 is the Hamburg version of the European Centre operational weather forecasting model modified for climate simulation purposes (Roeckner et al. 1992; DKRZ 1992). ECHAM3 is a low-order spectral model with a triangular truncation at wavenumber 42 (T42). Nonlinear terms and the parameterizations of physical processes are calculated on a 128×64 Gaussian grid, which yields a horizontal resolution of about $2.8° \times 2.8°$. Nineteen vertical levels are defined on σ surfaces in the lower troposphere and on p surfaces in the upper troposphere and in the stratosphere.

The ocean model HOPE (Hamburg Ocean Model in Primitive Equations) is based on primitive equations [see Latif et al. (1994) and references therein]. The model covers the entire globe and includes realistic bottom topography. The meridional resolution is variable, with 0.5° within the region 10°N–10°S. The resolution decreases poleward to match the T42 resolution of the atmosphere model. The zonal resolution is constant and also matches the T42 resolution of the atmosphere model. Vertically, there are 20 irregularly spaced levels, with 10 levels within the upper 300 m. The vertical mixing is based on a Richardson-number-dependent formulation and a simple mixed-layer scheme to represent the effects of wind stirring [see Latif et al. (1994) for details].

The two models were coupled without applying flux correction. However, since no sea ice has been included so far in the ocean model, the models interact only within the band of 60°N–60°S. At higher latitudes both models are forced separately according to climatology. SSTs and sea surface salinities of the ocean model are relaxed to Levitus (1982) climatology using Newtonian formulations with time constants of about 2 and 40 days, respectively, for the upper-layer thickness of 20 m. The atmosphere model is also forced by climatological SSTs at high latitudes. Within the coupling region the ocean model is forced by surface wind stress, the net heat flux, and the freshwater flux simulated by the atmosphere model, which in turn is forced by the SSTs simulated by the ocean model. The coupling is synchronous with an exchange of information every 2 h. Limiting the coupling to the latitude band 60°N–60°S suppresses possible impacts of sea ice, oceanic convection, and the thermohaline circulation on the climate variability of the model. Nevertheless, this obvious drawback of the model can also be viewed as an interesting sensitivity experiment. It makes it possible to study the role of interactions of the atmosphere with the wind-driven circulation of the ocean in generating decadal variability in the North Atlantic.

The coupled GCM was forced by seasonally varying solar insolation over a period of 125 yr. The oceanic initial conditions are those simulated at the end of a 15-yr spinup integration with climatological forcing. Since we are mainly interested in the variability of the upper ocean, such a short spinup time is justified. The atmospheric initial conditions were taken from a control run with climatological SST.

3. Mean climate and interannual variability

The global behavior of the coupled model with respect to climate drift and interannual variability during the first 20 yr of the integration is described in Latif et al. (1994). The drift in the SST occurs mainly during the first 20 yr of the experiment. Thus, we shall consider the years 20–125 only in the following, which are characterized by a relatively small drift in SST of the order of 0.05 K decade^{-1}. A stronger drift of about 0.1 K decade^{-1} occurs in the Labrador Sea. This drift seems to be related to the slow adjustment of the deep ocean, which is not in equilibrium due to the short spinup time. The meridional overturning of the Atlantic, for example, decreases from 24 to 21 Sv (Sv $\equiv 10^6$ m^3 s^{-1}) during the course of the experiment. The adjustment is characterized by a weak damped interdecadal oscillation. Since this adjustment involves very long timescales it should not affect the study of shorter-term decadal variability on timescales of 10–20 yr.

Figure 1a shows the distribution of the SST error for the North Atlantic relative to the GISST (Global Ice and

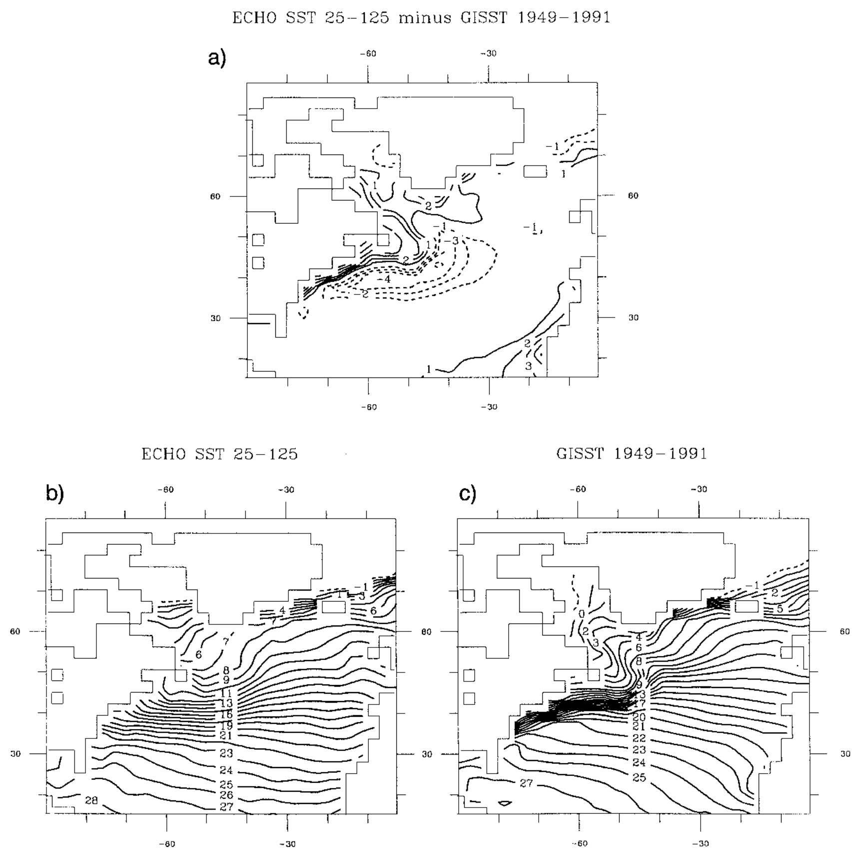

FIG. 1. (a) Annual mean SST difference (°C) between the ECHO model and the GISST dataset (1949–91), (b) annual mean SST for the coupled model, and (c) for the GISST climatology.

Sea Surface Temperature) dataset provided by the British Meteorological Office (UKMO). Although the data cover the period 1903–91, we used only the post-war data from 1949 to 1991, since these are more accurate than the pre-war data. Over large portions of the domain the SST is simulated reasonably well, with errors within ± 1 K. Large errors up to 4 K occur in the upwelling regions at the west coast of Africa. This problem is related to the inability of the atmosphere model to simulate sufficient low-level stratus clouds over the coastal upwelling regions. This causes an excess of incoming solar radiation resulting in too warm SSTs. Another area of considerable model errors is the Gulf Stream region, with errors up to ± 4 K. A comparison of the distributions of the full temperatures (Figs. 1b and 1c) reveals that the position of the Gulf Stream front is well captured but simulated much too diffuse. This is a problem that many coarse-resolution oceanic general circulation models have in common and is related to model resolution and the representation of mixing processes.

Although the ECHO-model is relaxed toward the observed climatological boundary conditions at high latitudes, it is able to reproduce a great amount of the observed interannual variability within the North Atlantic. Figure 2 shows the interannual standard deviation of SST for the period from 1949 to 1991 from the GISST dataset (Fig. 2b) and for the ECHO model (Fig. 2a). Over large portions of the ocean, especially in the subtropics, midlatitudes, and the eastern part of the basin, a rather good quantitative agreement is found. The simulation of the variability within the Gulf Stream area is similar to the observations but slightly shifted southward by approximately 5°. Considerable differences occur near the American continent only where the Gulf Stream leaves the coast. It is not surprising that the model underestimates the SST variability around 60°N, especially in the Labrador Sea, since these areas are strongly influenced by the relaxation at higher latitudes.

4. Comparison of the observed with the simulated decadal variability

To compare the dominant modes of observed and simulated North Atlantic ocean–atmosphere covariability, we performed canonical correlation analyses (CCA; Barnett and Preisendorfer 1987) of SST and SLP anomalies. CCA is a multivariate statistical analysis technique exploring the linear relationship between two sets of space–time-dependent variables. Basically, CCA decomposes the original data into pairs of spatial patterns

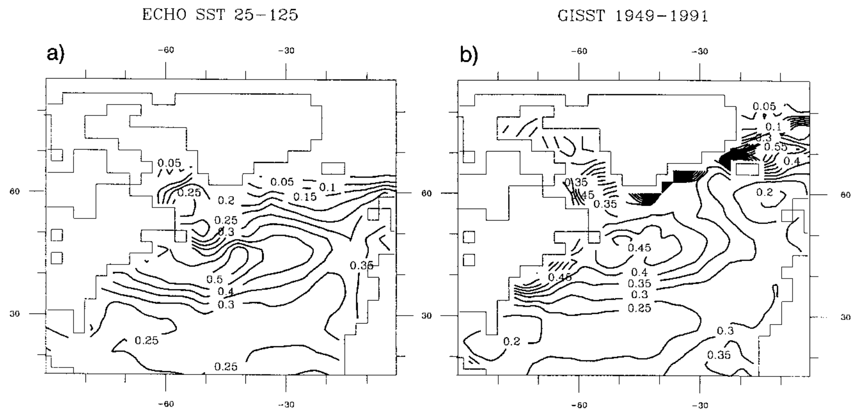

FIG. 2. Interannual standard deviation of SST (°C) (a) as derived from the coupled experiment and (b) from GISST dataset for the period 1949–91.

in such a way that their time series are optimally correlated.

Gridded instrumental observations of SST and SLP over the North Atlantic are available from the beginning of the century onward. These data reveal intense interdecadal variations and are dominated by a huge temperature increase around 1940. It has been suspected that these long variations might be related to processes in the ocean's interior, for example, the thermohaline-driven circulation. It cannot be expected from the setup of the model experiment to capture similar features in the simulation. Therefore, the CCA of observed SST and SLP anomalies was restricted to the postwar period where the interdecadal variations are less pronounced (Deser and Blackmon 1993). Moreover, the quality of the data strongly varies with time. The data coverage at the beginning of the century is very sparse, and substantial changes in measurement techniques occurred around World War II. Hence, it seems to be at least very difficult to compare the data before and after World War II directly.

We used the period 1949–88 from the observations in the CCA, while the full 100-yr period (years 25–125) was used from the model. Both CCAs were restricted to detrended wintertime anomalies. The coupling between ocean and atmosphere is strongest for this season and the SST is a good approximation for the thermal state of the upper ocean. To emphasize the low frequencies, the data were slightly smoothed with a three-point running mean filter. In addition an EOF truncation was applied prior to the CCAs. The results depend on the number of EOFs retained. It is necessary to find a balance between the need to retain as much as possible of the seeked signal and the requirement of noise reduction. Some preliminary sensitivity tests with different numbers of EOFs showed that retaining five EOFs is a good compromise in this particular case. These five EOFs explain 85% (78%) of the observed (simulated) SST variance and 91% (89%) of the observed (simulated) SLP variance.

The time series of the leading CCA mode of the observed data (Fig. 3a) have a canonical correlation of $r = 0.89$. Both the SST and the SLP time series exhibit clear variations with a decadal timescale. The corresponding spatial patterns of this mode resemble those of the interannual variability in the North Atlantic (Wallace et al. 1990; Zorita et al. 1992). A positive tongue of SST anomalies at the southeast coast of the United States extending into the central basin (Fig. 3b) is surrounded by negative anomalies in the midlatitudes south of Greenland and in the Tropics south of 30°N. This leading SST pattern explains about 17% of interannual SST variance with local extremes of more than 70%. The corresponding SLP pattern of the leading CCA mode (Fig. 3c) resembles the well-known North Atlantic oscillation (NAO), which represents an internal mode of the atmospheric circulation (Wallace and Gutzler 1981). The pattern explains about 36% of the SLP variance, with values up to 70% in the centers of action over Iceland and the Azores.

The leading CCA mode presented here resembles the quasi-decadal cycle described by Deser and Blackmon (1993), who used the EOF technique. The time evolution of the CCA mode with maxima around 1950, 1960, 1970, and 1985 and minima around 1955, in the late 1960s and the early 1980s is consistent with their results. The patterns show also a rather good agreement despite the fact that the subtropical tongue of positive anomalies off the southeast coast of America is much weaker in their analysis than in ours. The reason might be the different time interval analyzed. Nevertheless, both Deser and Blackmon's (1993) results and our findings indicate the existence of a decadal cycle in the North Atlantic.

To get further insight into the dynamics of the decadal variability in the North Atlantic, we analyzed additionally the decadal variability simulated by the ECHO CGCM. We performed a CCA between model SST and SLP anomalies. The leading CCA mode obtained from the simulation (Fig. 4) is very similar to that derived

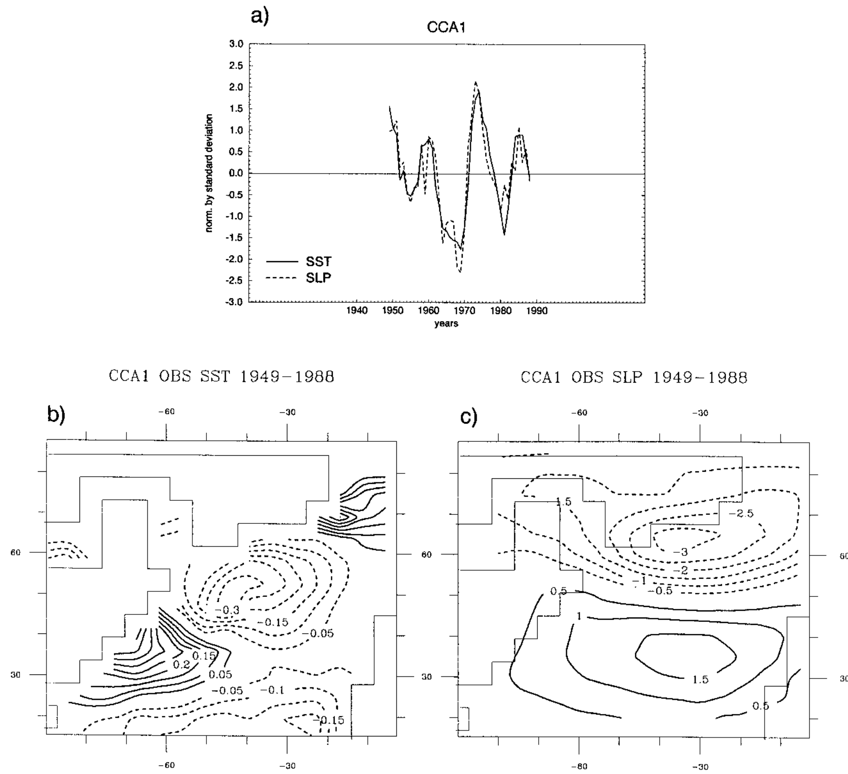

FIG. 3. The leading CCA mode of wintertime SST and sea level pressure in the North Atlantic as derived from observations (1949–88): (a) time series, (b) SST pattern (°C), and (c) sea level pressure pattern (hPa) of the first CCA mode.

from observations. The leading CCA mode of the model data explains 22.1% of the SST variance and 40.2% of SLP. Locally, more than 70% of the variance is explained in the centers of action. The time series of this mode (Fig. 4a) have a canonical correlation of $r = 0.76$. It can be clearly seen from the time series that this mode is associated with distinct variations with a decadal timescale. Spectra of these CCA time series (not shown) exhibit significant peaks at a period of about 17 yr, which is somewhat longer than the observational estimate.

The corresponding patterns of the first CCA mode derived from the ECHO model are in qualitative agreement with those derived from observations. The SST pattern (Fig. 4b) is also characterized by a tripole structure, with positive anomalies extending from the southeast coast of the United States into the central North Atlantic and negative anomalies to the north and to the south. The structure of the model pattern, however, is more zonal relative to the observations. The positive tongue extends farther to the east and attains its maximum values around 50°W while the observations show these near the American coast. The negative centers are located at the same positions but the amplitudes are relatively weak. The simulated SLP pattern (Fig. 4c) resembles also the North Atlantic Oscillation. The negative center of the dipole is located between Iceland and Greenland like in the observations but the positive extreme is simulated farther to the north at about 45°N. Thus, the midlatitude westwind anomalies related to the leading CCA mode of the model are confined to a narrower band relative to observations.

The atmospheric circulation anomalies derived both from the observations and the coupled model exhibit structures that are able to create the oceanic anomalies. The CCA–SLP patterns are related to wind anomalies that can force locally the corresponding SST patterns by anomalous surface heat fluxes and wind-induced mixing. The picture that a large portion of extratropical SST variability can be explained by a passively responding ocean has been supported by several observational and modeling studies. Lag correlations between SST indices and atmospheric parameters have shown that the atmosphere leads the ocean by several months (Davis 1976; Wallace et al. 1990; Zorita et al. 1992). Furthermore, studies by Alexander (1990) with a mixed-layer ocean model and Luksch et al. (1990) with an oceanic GCM show that most of the observed interannual wintertime SST variability of the North Pacific can be reproduced by prescribing the atmospheric forcing.

The opposite view that anomalous midlatitudinal SSTs might exert, at least under special conditions, a significant impact on the atmosphere is still rather controversial.

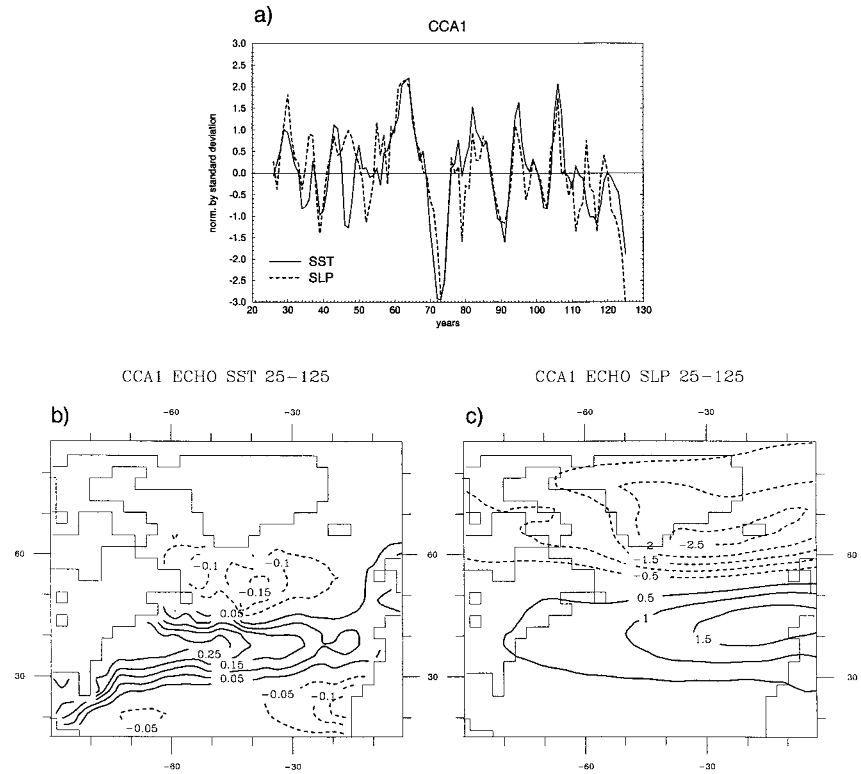

FIG. 4. The leading CCA mode of wintertime SST and sea level pressure in the North Atlantic as derived from the coupled experiment: (a) time series, (b) SST pattern (°C), and (c) sea level pressure pattern (hPa) of the first CCA mode.

Some observational studies (e.g., Ratcliff and Murray 1970; Namias and Cayan 1982) support this direction of interaction. Some evidence for a sensitivity of the atmosphere to extratropical SST variations was obtained by several experiments with atmospheric GCMs (Palmer and Sun 1985; Pitcher et al. 1988; Lau and Nath 1990; Kushnir and Lau 1992). The processes in the formation of the atmospheric response are complicated. Not only the stationary response of the atmosphere to stationary heating (Hoskins and Karoly 1981), but also the changes in the statistics of the transient eddies due to anomalous baroclinicity seem to be important (Palmer and Sun 1985), and the response might depend on the location of the zones of anomalous baroclinicity (Ting 1991).

The relative importance of tropical and extratropical SSTs for the formation of midlatitudinal atmospheric variability was addressed in several studies. Graham et al. (1994) showed by GCM experiments that in the Pacific the influence of the Tropics is clearly dominant. For the North Atlantic Hense et al. (1990) found a strong response to tropical SST anomalies and a weaker influence of extratropical SSTs with different signature. Kharin (1995) reports a NAO-like atmospheric response over the North Atlantic in an atmospheric GCM experiment with globally prescribed observed SSTs similar to that shown above. But since the NAO represents an eigenmode of the atmospheric circulation, which exists even without any external forcing, the atmospheric response in those experiments might be also triggered by remote influences from the North Pacific or the Tropics so that the results remain inconclusive. Moreover, the sensitivity experiments of Kharin (1995) show that the influences of tropical and extratropical SSTs on the atmosphere might interact nonlinearly. Thus, no final conclusion can be drawn from these studies.

The existence of a dominant period, about 12 yr in the observations and about 17 yr in the ECHO model, is a strong indication for an oscillatory process. This would require some kind of a feedback loop. If such a feedback loop would rely on ocean–atmosphere interaction, as has been originally proposed by Namias (1959, 1969) and Bjerknes (1964), a sensitivity of the atmosphere to extratropical SST anomalies would be essential.

5. The coupled decadal mode

We shall develop in the following a scenario that explains the existence of a decadal timescale in the North Atlantic climate system. Since the ECHO model reproduces the observations reasonably well, we base our explanation on the coupled model results. The existence of a dominant decadal timescale in the simulation, which was already discernible in the CCA time series (see Fig. 4), is supported by two selected Fourier spectra shown

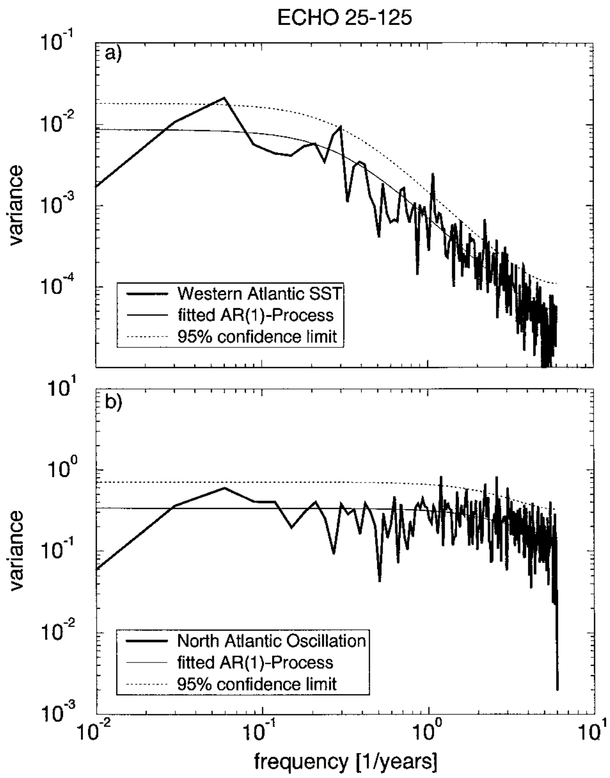

FIG. 5. Fourier spectra of anomalous western subtropical North Atlantic SST (30°–40°N, 45°–90°W) and of an index describing the model's North Atlantic Oscillation (60°–70°N, 0°–45°W minus 35°–45°N, O°–45°W).

in Fig. 5. The spectra were obtained by applying Bartlett's procedure with six degrees of freedom, yielding a longest resolvable period of 34 yr. For comparison, the spectra of equivalent red noise (first-order autoregressive) processes, as they are estimated from the respective time series, and the 95% confidence limits for accepting a red noise null hypothesis are also shown.

The spectrum of the SST anomalies (Fig. 5a) averaged over the western subtropical Atlantic (30°–40°N, 45°–90°W), the area where the CCA pattern of SST described above explains most of the variance, shows clearly a red noise behavior on the intraseasonal and interannual timescales, consistent with the stochastic climate model hypothesis. At a period of 17 yr the spectrum exhibits a clear peak that is significant at the 95% level above the red noise background. In the atmosphere the dominance of the 17-yr period is not that pronounced but still visible: a red noise behavior cannot be found in the spectrum of a NAO index (Fig. 5b), as defined by the difference of SLP anomalies averaged over 60°–70°N, 0°–45°W and 35°–45°N, 0°–45°W. The spectrum of this time series is essentially white, and the estimated AR(1)-process exhibits basically the properties of uncorrelated white noise. But in correspondence with the SST spectrum, the spectrum of the NAO index exhibits also enhanced variability at the 17-yr timescale. This peak, however, is not statistically significant above the 95% confidence limit.

The dominant 17-yr timescale is related to an oscillation that is characterized by propagating temperature anomalies within the upper layers of the North Atlantic Ocean. This was revealed by a principal oscillation pattern (POP) analysis (Storch et al. 1995). POPs are the normal modes of a first-order linear system where the system matrix is estimated from the data. POPs are in general complex with a real part, Re, and an imaginary part, Im. The complex coefficient time series satisfy the standard damped harmonic oscillator equation with a rotation period T and a damping time τ. The evolution of the system in the two-dimensional POP space can be described by a cyclic sequence of spatial patterns:

$$\ldots > -\mathrm{Re} > \mathrm{Im} > \mathrm{Re} > -\mathrm{Im} > -\mathrm{Re} \ldots.$$

The oceanic temperature anomalies at 100-m depth were used in the POP analysis, since this variable represents better the thermal state of the upper ocean and is less contaminated by noise than SST. Prior to the analysis the annual mean data were smoothed by a 5-yr running mean filter. A weaker filtering does not change the principal results of our analysis, only their statistical significance relative to the background noise becomes weaker.

The leading and only significant POP derived from the model data explains about 21% of the total variance relative to the filtered data. Locally, more than 60% of the variance can be described by that POP. The POP has a rotation period T of 17 yr and a quite long damping time τ of 21 yr. The time series of the imaginary and the real part of the POP (Fig. 6a) show a rather regular oscillation, with a strong "event" between model years 90 and 100. The two time series exhibit a 90° phase relationship at the POP period, as theoretically expected, which was confirmed by a cross spectral analysis (not shown).

The space–time evolution as described by this single POP mode can be inferred from the sequence of the imaginary and the real part patterns. The imaginary part (Fig. 6b) is characterized by a positive signal that extends in a northeastward direction from Florida into the central North Atlantic. It is surrounded by negative anomalies, with minima near 40°N, 30°W and in the central subtropics. A quarter of the rotation period later—that is, about 4 yr—during the positive phase of the real part (Fig. 6c), the positive anomaly has propagated to the northeast and intensified. Its maximum is now located at 40°N, 50°W. The surrounding negative values also changed their positions. The subtropical anomaly is reminiscent of Rossby wave activity, since the negative anomalies extend farther to the west the farther south its location is. The negative extreme of the midlatitudes intensified and moved northeastward. It is now centered at 50°N, 20°–30°W. Afterward, the sequence of events is repeated but with reversed signs. Since the highest amplitudes occur during the real part phase, it can be viewed as the extreme phase of the oscillation described by the

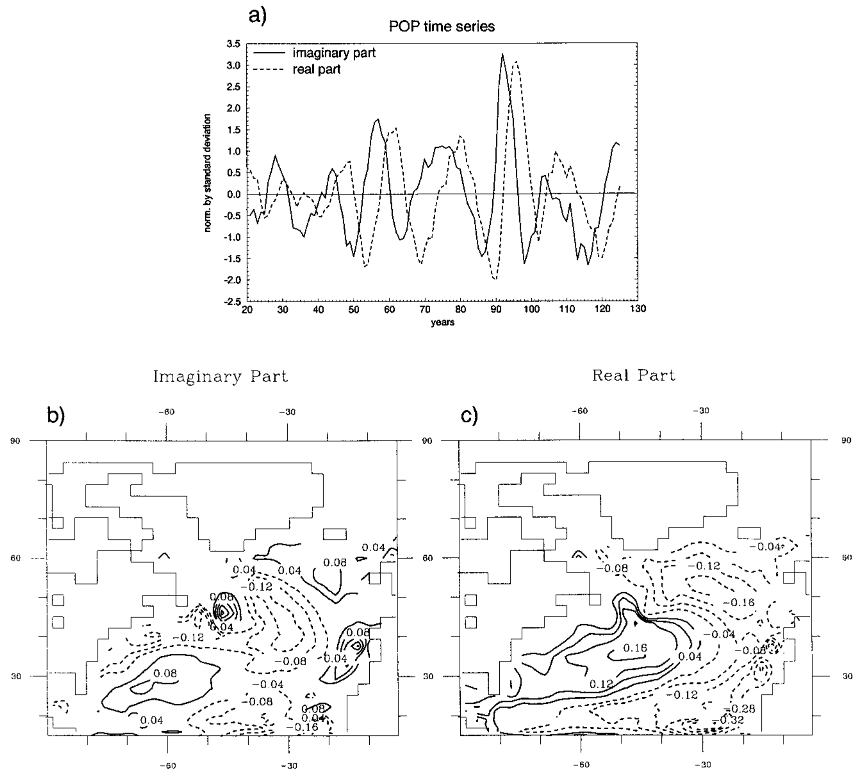

FIG. 6. Evolution of annual mean temperature at 100-m depth as derived from the first POP mode of the coupled experiment. (a) Time coefficients of the real and imaginary parts, (b) spatial pattern of the imaginary part, and (c) the spatial pattern of the real part. The time series are normalized to standard deviation one.

POP mode, while the imaginary part can be viewed as the transition phase between these extremes.

In conclusion, the following picture can be drawn from the evolution of subsurface anomalies as obtained from the leading POP mode. Positive and negative temperature anomalies travel around the North Atlantic subtropical gyre with a typical timescale of about 17 yr. They intensify as they travel along the Gulf Stream extension and weaken thereafter.

An oscillation in the North Pacific Ocean of approximately 20-yr period was described by Latif and Barnett (1994, 1996) in the same coupled experiment and documented from observations as well. The structure and evolution of this decadal mode is quite similar to that presented in this paper for the North Atlantic Ocean. Latif and Barnett (1994, 1996) explained the Pacific mode by unstable larger-scale air–sea interactions in the subtropics and midlatitudes and the dynamics of the Northern Pacific subtropical gyre. When, for instance, the gyre is anomously strong, more warm tropical water masses are transported poleward by the Kuroshio current and its extension leading to positive SST anomalies in the midlatitudes. The atmospheric response to such a SST anomaly involves a weakened storm track. The associated changes at the air–sea interface reinforce the initial SST anomaly, so that ocean and atmosphere act as a positive feedback system. The atmospheric response, however, consists also of a wind stress curl anomaly that weakens the subtropical gyre, thereby reducing the poleward heat transport and the initial temperature anomaly. The transient response of a midlatitude ocean to a variable wind stress has been described in many theoretical and modeling papers (e.g., Anderson and Gill 1975; Anderson et al. 1979; Gill 1982). The response involves the propagation of planetary waves in interaction with the mean horizontal currents. The net effect is a modification of the strength of the subtropical gyre circulation, with typical spinup times of several years to decades. This transient response of the ocean is thought to represent the memory within the feedback loop, allowing continuous oscillations within the North Pacific Ocean.

The decadal oscillation in the North Atlantic, as simulated by the ECHO model, seems to be caused also by an interaction of the subtropical gyre and the extratropical atmospheric circulation. We would like to illustrate this by means of some key components of the feedback loop. Figure 7 shows the temporal evolution of the strength of the Northern Atlantic subtropical gyre in the ECHO model. As a rather crude index for the gyre intensity we calculated the total northward mass transport of the western boundary current at 34°N. To emphasize

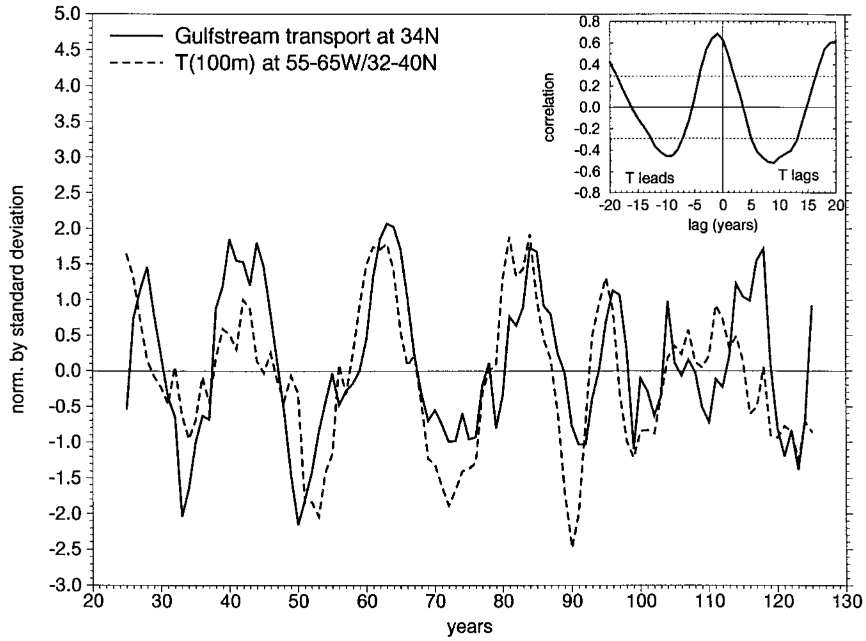

FIG. 7. Time series of the anomalous subtropical gyre intensity (as measured by the mean northward transport of the western boundary current at 34°N) and the anomalous temperature at 100-m depth averaged over the region 32°–40°N, 55°–65°W. Both index series were normalized to standard deviation one and filtered by a 3-yr running mean. The small panel shows the cross correlation function of the two time series. The thin dotted lines give the 95% confidence limits for the correlation coefficients.

the decadal fluctuations, this time series has been filtered by a 3-yr running mean. It is obvious that the model Gulf Stream exhibits distinct variations on the decadal timescale. A subsurface (100 m) temperature index computed over the area 32°–40°N, 55°–65°W is also presented in Fig. 7. This time series is a good index of the extreme phase of the cycle described above, as revealed by a comparison with the results of the CCA (see Fig. 4) and the POP analysis (see Fig. 6). The Gulf Stream intensity and the temperatures of the western North Atlantic are well correlated with each other. The cross correlation function of these two index time series (Fig. 7, small panel) exhibits a rather regular behavior, with a maximum of almost 0.7 at a lag of -1 yr and minimum of around -0.5 at lags of ± 9 yr. Another maximum is found at a positive lag of 20 yr. Although the time series are too short for precisely estimating a timescale from their cross-correlation function, the latter is indicative for a periodic (stochastic) process with an oscillation period of slightly less than 20 yr.

The oceanic temperature anomalies are associated with characteristic anomaly patterns of the atmospheric circulation. The extreme phase of the oscillation, which exhibits a strong temperature anomaly at 40°N in the western and central part of the ocean surrounded by opposite anomalies in the midlatitudes and at lower latitudes, is related to the NAO pattern (Fig. 4). Higher than normal temperatures at 40°N, which implies an enhanced meridional temperature gradient in the midlatitudes, are accompanied by an *intensified* Icelandic low and Azorian high and therefore enhanced westerlies (Figs. 4b,c). Higher than normal pressure in the subtropical high goes along also with an enhancement of the trade winds south of it. The relationship between SST anomalies and atmospheric circulation is different to that described by Latif and Barnett (1994, 1996) for the North Pacific. They found that positive SST anomalies along the Kuroshio extension are accompanied by a *weakened* Aleutian low. This opposite relationship between ocean and atmosphere in the North Atlantic relative to the North Pacific might be explained by the different location of the extratropical lows relative to the relevant SST anomalies. The Aleutian low and consistently the Northern Pacific stormtrack are located about 10° farther south than their North Atlantic counterpart. The temperature anomalies in the western subtropical Pacific of the decadal mode described by Latif and Barnett (1994, 1996) and the western subtropical SST anomalies of the Atlantic mode are located at similar latitudes around 30°–45°N. In relation to the stormtrack this implies a reduction of the relevant meridional temperature gradient in the North Pacific and an enhancement in the North Atlantic. This difference might be responsible for the opposite atmospheric responses over the two oceans.

The atmospheric circulation anomalies over the North Atlantic feed back onto the ocean's surface. In the midlatitudes (north of 40°N) and the eastern subtropics they

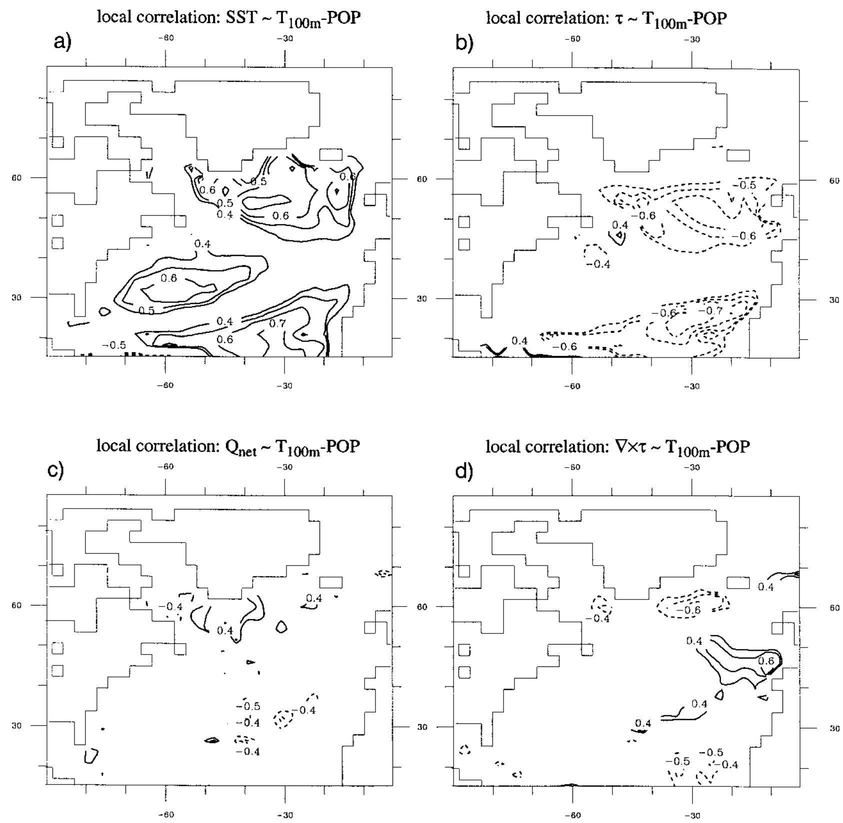

FIG. 8. Correlation patterns of the local decadal T_{100m} development (reconstructed from the POP shown on Fig. 6) with (a) sea surface temperature, (b) total surface wind stress, (c) net surface heat flux, and (d) the wind stress curl anomalies.

amplify an initial SST anomaly, thereby acting as a positive feedback. The relevant quantities related to local air–sea interaction are the net surface heat flux; the magnitude of the surface wind stress, which forces temperature anomalies by anomalous surface mixing; and the curl of the wind stress, which causes Ekman induced upwelling and downwelling. Warmer than normal temperatures are created by positive heat flux anomalies, reduced wind stresses, or negative wind stress curl anomalies. Due to the nonlinear relationship between vertical velocities and mixed layer temperatures, the latter are relevant only in upwelling regions. Correlations of these variables with the local temporal development of the Atlantic decadal mode at each grid point, as reconstructed from the T_{100m}-POP (Fig. 6), are given in Fig. 8. For comparison the correlation of the POP with the SST is also shown (Fig. 8a). The SST anomalies are closely related to the subsurface temperature anomalies associated with the decadal POP mode. Correlations exceed values of 0.7 in the three key regions, namely, the midlatitude Atlantic and the western and eastern subtropical Atlantic. It is mainly the anomalous mixing due to anomalous wind stress that is responsible for the temperature anomalies in the midlatitudes (anomalous westerlies) and eastern subtropics (anomalous trade winds) (Fig. 8b). The wind stress magnitude and 100-m temperature anomalies (as modeled by the POP) are strongly anticorrelated, with minimum correlations of less than -0.7 in those areas. The anomalous heat flux and the wind stress curl are less important for the local temperature evolution. The net heat flux correlation pattern (Fig. 8c) reveals some positive correlations south of Greenland, and there are also some indications of minor negative correlations in the wind stress curl (Fig. 8d) in the midlatitudes and eastern subtropics. Thus, in the midlatitudes and the eastern subtropics the atmosphere has a clear tendency, mainly due to anomalous wind stress, to force the respective oceanic temperature anomalies. Such a relationship is not found in the western subtropical Atlantic, the third important region of the decadal mode. None of the atmospheric forcing variables exhibits any significant correlations with the temperature anomalies in this region. We conclude that the decadal temperature fluctuations in the western subtropical Atlantic are caused by processes within the ocean and not by local air–sea interactions.

More insight into the dynamics governing the model's air–sea interactions can be obtained from the phase relationship between atmospheric forcing and the state of the upper ocean. It is well known that on monthly to interannual timescales the extratropical upper-ocean temperature fluctuations are mainly driven by high-frequency weather fluctuations (e.g., Frankignoul 1985). It is the key point of the concept of stochastic climate models (Hasselmann 1976) that the ocean acts as an integrator of these random fluctuations producing high levels of variability at low frequencies. Such simple stochastic feedback models, in which SST anomalies are driven by atmospheric noise and a feedback is considered, yield typical relationships between oceanic and atmospheric quantities (Frankignoul 1985). Lagged correlations, for example, of SST and heat flux show that the atmospheric fluxes generate the temperature anomalies at lags where the atmosphere is leading. Maximum correlations are attained when the atmosphere is leading the ocean by a few months. Usually air–sea interactions tend to dampen an SST anomaly, once it has been generated. Such a negative feedback results in a rapid drop of the correlation function at lag zero. When the atmosphere is lagging the correlations are of opposite sign to those when the atmosphere is leading. It is this kind of behavior that is usually observed when extratropical air–sea interactions are studied (Wallace et al. 1990; Zorita et al. 1992). On the other hand, a positive feedback between SST and atmospheric anomalies results in correlation functions with the same signs at positive and negative lags. This is caused by the fact that in the case of a positive feedback the atmospheric flux anomalies are still contributing to the SST anomalies when these are already decaying. It is possible to extend the stochastic climate model concept, which was originally formulated for seasonal to interannual timescales, to even longer timescales as long as the atmospheric spectra are essentially white (Zorita and Frankignoul 1996).

The character of local air–sea interactions within the ECHO model is illustrated by cross-correlation functions between the anomalies in SST and the net heat flux, the wind stress, and the wind stress curl. Figure 9 displays such functions for the midlatitudes (50°–60°N) and the eastern (15°–30°N, 0°–45°W) and the western (27°–38°N, 45°–90°W) subtropics, the regions that are relevant for the decadal mode (see Fig. 8). To highlight the long-term behavior the data were smoothed with a 3-yr running mean filter prior to the analysis.

The interaction between ocean and atmosphere is very weak in the western subtropical area (Fig. 9a). The correlations are hardly significant and no clear phase relationships can be detected. Over the midlatitude ocean and the eastern subtropics (Fig. 9b and 9c), changes in the net heat flux and magnitude of the wind stress are contributing to the creation of SST anomalies when the atmosphere is leading; the curl of the wind stress is of minor importance. The correlation functions exhibit the same sign at positive and negative lags. As explained above a symmetric correlation function is a strong indication for a positive feedback between ocean and atmosphere. If the ocean would react only passively to the atmospheric forcing, the cross-correlation function would show a sign reversal at about zero lag: at negative lags, when the atmosphere is leading, the atmospheric forcing would contribute to the generation of SST anomalies, while it would have a damping effect at positive lags. The cross-correlation function between the SST and net heat flux in the eastern subtropics (Fig. 9c) shows such a behavior. However, this is not the case for the forcing variables in the midlatitudes and the wind stress magnitude in the eastern subtropics: they still contribute to the enhancement of SST anomalies at positive lags (when the ocean is leading and the SST anomalies are already decaying) up to several years. Thus, unstable air–sea interactions exist in some regions of the North Atlantic of the ECHO model. These unstable air–sea interactions are also relevant to the generation of the decadal cycle (Fig. 8).

In the following, the phase switching mechanism responsible for the oscillatory nature of the decadal mode will be described. As mentioned above, the temperature anomalies in the western subtropical Atlantic are not generated by local atmospheric forcing. Thus, the decadal temperature fluctuations in that region must originate from processes within the ocean. The North Atlantic subtropical gyre reveals distinct intensity fluctuations with the relevant timescale. The western subtropical temperature anomalies are closely related to these circulation anomalies (see Fig. 7). Such variations of the Gulf Stream in the presence of the mean meridional temperature gradient cause anomalous meridional heat transports and thus contribute to the generation of western subtropical temperature anomalies. These temperature anomalies, however, lead slightly (by a lag of 1–2 yr) the intensity variations of the gyre. Thus, advection of mean temperature by anomalous currents cannot be the only source of the temperature anomalies.

The advection of temperature anomalies by the mean oceanic currents is also important. Temperature anomalies of opposite sign to those in the western subtropical Atlantic are generated within the eastern subtropics due to anomalous local air–sea interactions (Figs. 8 and 9c). The evolution of the T_{100m}-POP in Fig. 6 shows that these anomalies propagate to the west. They cross the ocean basin and after some years they contribute to the phase reversal. The simulated mean currents averaged over the upper 250 m (the temperature anomalies are mainly confined to this depth range) are shown in Fig. 10a. The velocities in the westward drifting branch of the gyre between 15° and 30°N range from 1 cm s^{-1} in the east to 10 cm s^{-1} in the western part. From these velocities traveling times for water parcels drifting across the subtropical ocean can be estimated. The traveling times range from 5 to 10 yr (depending on the exact choice of the starting and ending points of the trajectories), which matches reasonably well half of the period of the simulated cycle. The fact that temperature anomalies can survive for such a long time and travel over long distances has recently been supported in a study by Sutton and Allen (1997), who provide persuasive evidence from observations for advective propagation of temperature anomalies.

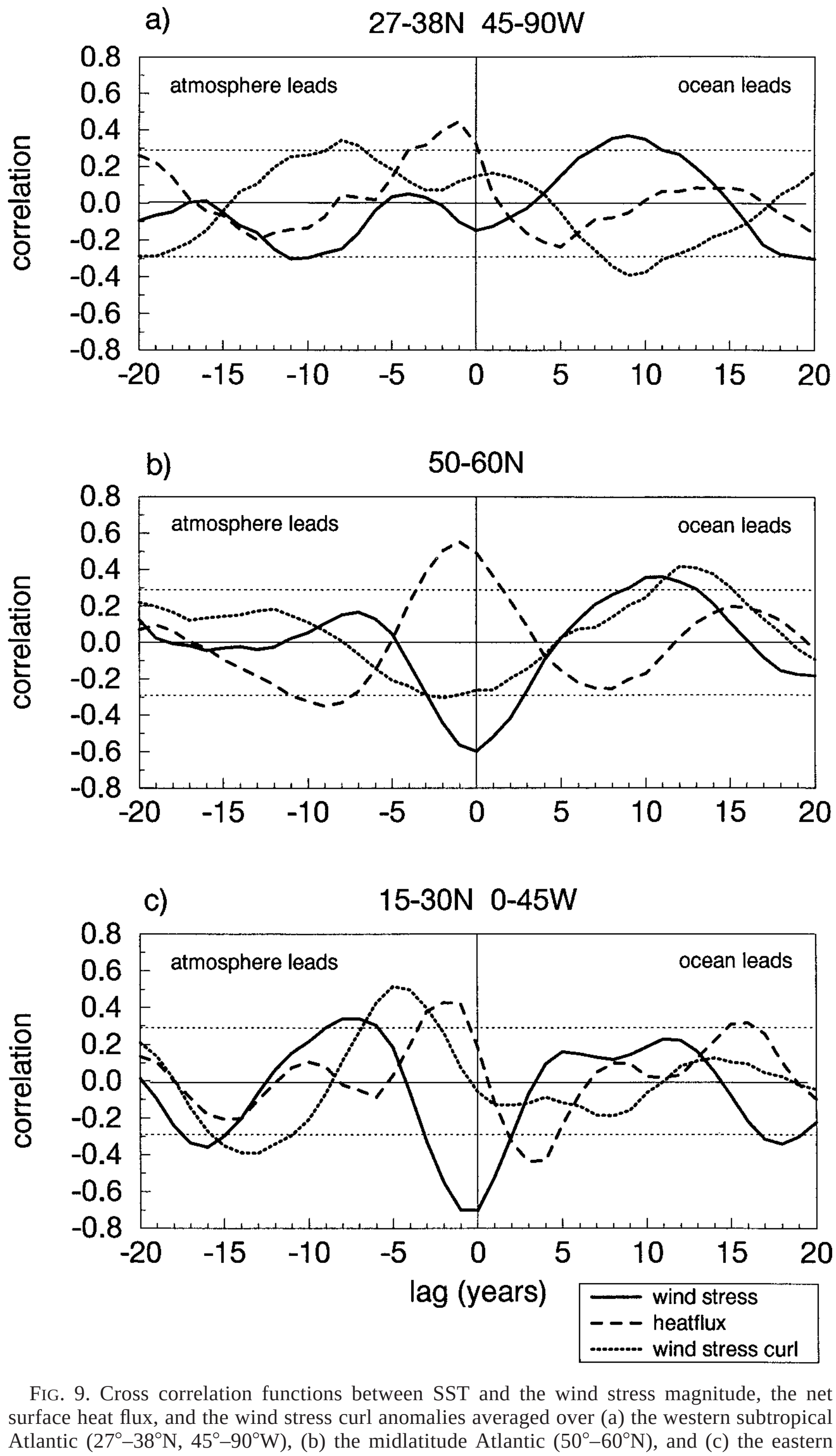

FIG. 9. Cross correlation functions between SST and the wind stress magnitude, the net surface heat flux, and the wind stress curl anomalies averaged over (a) the western subtropical Atlantic (27°–38°N, 45°–90°W), (b) the midlatitude Atlantic (50°–60°N), and (c) the eastern subtropical Atlantic (15°–30°N, 0°–45°W). The thin dotted lines give the 95% confidence limits for the correlation coefficients.

The intensity fluctuations of the subtropical gyre can be understood as an adjustment to varying wind forcing as will be shown below. This adjustment process involves westward-propagating baroclinic Rossby waves, and the adjustment timescale is determined by the traveling times of the planetary waves crossing the subtropical ocean (Anderson and Gill 1975). To give an indication of the relevant wave timescales within the ECHO model, the phase speeds of the first baroclinic, the fastest, and dominating Rossby mode as derived

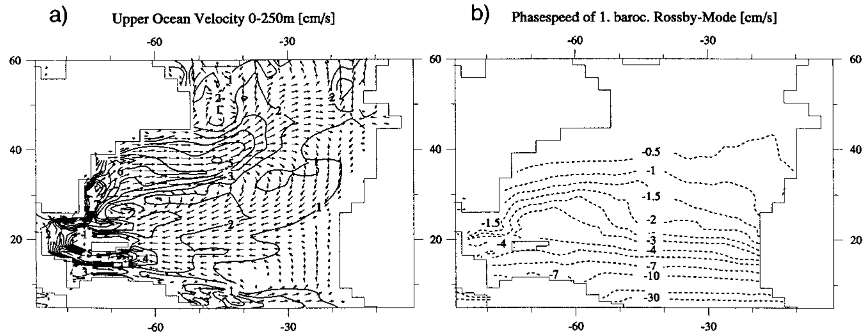

FIG. 10. (a) Mean currents (cm s^{-1}) averaged over the upper 250 m of the ECHO model and (b) phase velocities of the first baroclinic Rossby mode (cm s^{-1}). The arrows in (a) indicate the directions of the currents but not their intensity.

locally from the model's mean density stratification are displayed in Fig. 10b. Within the region of interest between 15° and 30°N they amount to 1.5–5 cm s^{-1}. Such phase speeds yield traveling times of 5–10 yr for the Rossby waves to reach the western boundary, which is also close to half of the period of the decadal mode.

As discussed, both the advection of temperature anomalies by mean currents and heat transports by an anomalous gyre contribute to the generation of temperature anomalies in the western subtropical Atlantic. The short phase lag between the time series of western Atlantic temperature anomalies and the gyre intensity with the temperatures leading by 1–2 yr (Fig. 7) can be explained by small differences in the advective and wave timescales. When the advective timescale is slightly shorter than the Rossby wave timescale the lag becomes plausible. The estimation of the traveling times given above is not precise enough to prove this rigorously. More insight could be obtained from a heat budget analysis of the upper ocean. Moreover, such an analysis would help to clarify the question of whether advection or Rossby wave adjustment is more important for the phase-switching mechanism. Unfortunately, the data necessary for such a calculation were not stored during the course of the experiment. Hence, the heat budget analysis has to be postponed to future experiments.

It remains to show the origin of the intensity fluctuations of the subtropical gyre. When, for instance, the temperatures are anomalously high in the subtropical western North Atlantic during one phase of the oscillation, this is associated with strong wind anomalies over a large portion of the North Atlantic Ocean. As shown above, this affects the local temperature balance of the ocean. Negative temperature anomalies are created by enhanced trade winds in the eastern subtropical ocean, which travel along the southern branch of the subtropical gyre to the west. They reach the western boundary within 5–10 yr and replace the original temperature anomaly. But the wind anomalies act also on the dynamics of the subtropical gyre circulation by a typical wind stress curl anomaly pattern. This pattern exhibits a dipole structure with a sign reversal between 30° and 35°N (not shown). Positive temperature anomalies and an anomalous strong gyre are related to an anomalous strong Icelandic low and Azorian high. The strengthened NAO pattern goes along with enhanced westerlies and trade winds. Extrema are located at 50° and 30°N. This causes negative wind stress curl anomalies within this latitude band. South of 30°N the negative zonal wind stress anomalies related to the strengthened trade winds gradually decrease, resulting in a positive wind stress curl anomaly in the subtropics. Hence, the wind stress curl anomalies north of 30°N have a tendency to accelerate the gyre while those to the south tend to spin the gyre down.

However, the gyre does not respond instantaneously to the changes in the wind stress curl forcing. The wind stress curl anomalies force westward-propagating baroclinic Rossby waves by local Ekman pumping. A new Sverdrup balance is established in the wake of those Rossby waves. The first baroclinic Rossby mode needs about 5–10 yr to cross the basin in the southern part of the gyre between 15° and 30°N. To the north the Rossby waves are much slower. At 40°N their phase speeds amount to 0.5 cm s^{-1} only (Fig. 10b). The corresponding traveling times in the northern part of the gyre are of the order of 20 yr and longer. Thus, the intensity of the subtropical gyre cannot adjust within half of the period of the cycle under consideration to wind stress curl anomalies imposed in its northern parts. Only the wind stress curl anomalies over the southern branch of the gyre impose a forcing that can modify the gyre within the appropriate time frame. The southern pole of the wind stress curl anomalies is centered over the eastern subtropics. Therefore, the time evolution of the wind stress curl anomalies averaged over the region 15°–30°N, 0°–45°W and the intensity anomalies of the gyre are displayed together in Fig. 11. It is obvious that the wind stress curl anomalies oscillate like the gyre intensity on the decadal timescale under consideration. Since a negative wind stress curl anomaly tends to accelerate the gyre, it is the minimum negative correlation at lags around −12 yr that is indicative for the adjustment time

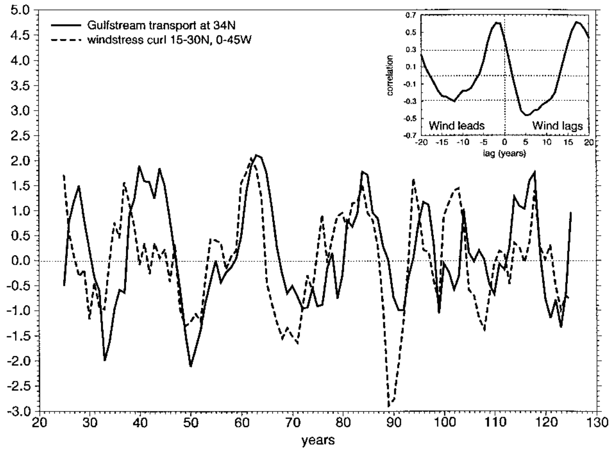

FIG. 11. Time series of wind stress curl anomalies averaged over 15°–30°N, 0°–45°W and the anomalous intensity of the subtropical gyre (as measured by the mean northward transport of the western boundary current at 34°N). Both index series have been normalized to standard deviation one and filtered by a 3-yr running mean. The small panel shows the cross-correlation function of the two time series. The thin dotted lines give the 95% confidence limits for the correlation coefficients.

of the gyre to varying wind forcing. This time lag is somewhat longer than the timescale estimated from the first baroclinic Rossby mode. However, the first Rossby mode represents only a lower limit for the gyre adjustment time since higher, slower propagating Rossby modes are also involved in the adjustment to slowly varying wind forcing.

The fact that the phase lag between the anomalies of the wind stress curl and the gyre intensity is longer than half of the period of the cycle apparently leads to an inconsistency. However, the wind stress curl and the subtropical gyre do not strictly vary in phase. Their cross-correlation function (Fig. 11) reveals that the curl is slightly leading by 1–2 yr. The western subtropical Atlantic temperature anomalies also lead the gyre intensity by a similar lag (see Fig. 7), which was explained by the advection of temperature anomalies by the mean currents. The wind anomalies force both the eastern subtropical temperature anomalies, which are advected westward, and the westward-traveling Rossby waves. Since the advective timescale appears to be faster than the Rossby wave adjustment timescale, the phase reversal of the western subtropical temperature anomalies and therefore also of the atmospheric circulation happens a few years before the gyre is completely adjusted.

The decadal cycle as simulated by the ECHO model shares many similarities with the observed decadal variability of the North Atlantic Ocean. However, the model yields an oscillation period of about 17 yr, while the observations indicate a timescale of about 12 yr. The mechanism causing the oscillation in the model results relies on advection by mean currents and on Rossby wave propagation. The adjustment time of the subtropical gyre to anomalous wind forcing depends on the basin size and on the phase speeds of the baroclinic Rossby waves. Hence, the stratification of the ocean is a crucial parameter in determining the adjustment timescale. The stratification of our ocean model is more diffusive relative to observations (Frey et al. 1997), which results in too slow Rossby waves. At Bermuda (32°N, 64°W), for example, observational estimates yield phase speeds for the first baroclinic Rossby mode ranging from 2 to 2.3 cm s^{-1} (Sturges and Hong 1995; Frankignoul et al. 1997), while phase speeds of only 1.4 cm s^{-1} are calculated from the model. This discrepancy might explain partially the too-long oscillation period in the model.

Another reason may be attributed to the mean currents in the upper ocean, which are simulated too weakly. The transport of the Gulf Stream at Cape Hatteras, where it leaves the American coast, amounts to about 85 Sv, while the model simulates only 40 Sv. This is a problem that most of the global oceanic general circulation models have in common, and it can be mainly attributed to the relatively coarse resolution in the models. Thus, both errors in the representation of the wave propagation and mean currents might account for the relatively long period of the model's decadal oscillation.

6. Discussion

It is well known from observations that the North Atlantic Ocean exhibits considerable interannual and interdecadal variability. At least the interannual variations can be mainly explained by the low-frequency response of the ocean to random weather fluctuations (e.g., Wallace et al. 1990; Zorita et al. 1992). However, the observational records are too short to decide whether the observed decadal fluctuations occur due to stochastic processes (Frankignoul et al. 1997), or if they could be viewed as either self-sustained or stochastically forced oscillations of the North Atlantic climate system. Experiments with oceanic general circulation models have indicated that long-term variations with timescales of several decades to centuries might occur as a result of self-sustained oscillations of the thermohaline circulation, possibly in interaction with the subpolar and subtropical gyres (e.g., Weaver and Sarachik 1991; Delworth et al. 1993). The observed decadal climate anomalies within the North Atlantic, however, can also result from ocean–atmosphere interactions. Latif and Barnett (1994, 1996) presented evidence for the existence of a decadal coupled mode of approximately 20-yr period in the North Pacific, which was shown to arise from such large-scale air–sea interactions.

Here, we showed by investigating the same extended-range integration with the ECHO CGCM that an interaction of the wind-driven ocean gyre circulation and the extratropical atmospheric circulation might also cause decadal climate fluctuations in the North Atlantic. Due to the chosen boundary conditions (both the ocean and the atmosphere were forced by climatology poleward of

60°N) possible interactions with the thermohaline circulation were effectively suppressed. Using CCA and POP analyses a coupled mode of approximately 17-yr period was identified, with relatively strong anomalies along the western boundary current and its extension, in the midlatitude and subtropical oceans. The mechanism causing this oscillation seems to be consistent with the early ideas originally proposed by Namias (1959, 1969) for the Pacific and Bjerknes (1964) for the Atlantic Oceans. Temperature anomalies in the North Atlantic cause a specific response of the atmospheric circulation in the midlatitudes. The westerlies and hence the air–sea interactions are modified in such a way that the original anomalous meridional temperature gradient is reinforced, thereby forming a positive feedback loop. Two processes contribute to the phase-switching mechanism. First, temperature anomalies generated by anomalous local air–sea interactions in the eastern subtropics are advected westward by the mean currents and replace the initial temperature anomalies in the west after some years. Second, the atmospheric response is associated with a characteristic anomaly in the wind stress curl. This provides also a delayed negative feedback by modifying the strength of the subtropical gyre, which is accomplished by baroclinic Rossby waves. Whether advection or Rossby wave adjustment is more important for the phase switching mechanism is not clear and will be matter of further investigations.

If one neglects the very long-term interdecadal to centennial variations in the observed SSTs of the North Atlantic, which might originate from variations in the thermohaline circulation, considerable similarities between the observed variability patterns and those simulated by the ECHO model exist. Hence, the results of the ECHO model may provide a contribution to the understanding of the mechanisms causing decadal variability in the North Atlantic. However, the observations indicate enhanced variability at a period of about 12 yr, while the mode simulated by the model has a period of about 17 yr. Both the phase speeds of the baroclinic Rossby waves and the intensity of the mean currents are underestimated by the model due to deficiencies in the representation of the relevant physical processes. Therefore, the difference in the oscillation periods of the model and the observations can easily be explained within the limitations of the model.

The decadal fluctuations of the model's North Atlantic might be also influenced by processes outside that region. Our model results indicate coherent fluctuations of the North Atlantic and North Pacific subtropical temperatures, and the decadal modes in the Pacific and Atlantic seem to be linked to each other. Uncoupled SST sensitivity experiments with the atmospheric component ECHAM3 reveal a strong anticorrelation between the PNA and NAO patterns, a feature that is not found in observations, at least on the timescales considered here. The North Atlantic decadal cycle described in this paper could be viewed as the result of a damped coupled ocean–atmosphere oscillator under the influence of stochastic forcing. The *e*-folding time of the T_{100m}-POP (Fig. 6) of approximately 21 yr supports this picture. If the stochastic, weather-induced noise is substantial to overcome the inherent damping of the system or if it induces only additional irregularities cannot be decided from the model results. However, a link to the Pacific mode by an atmospheric teleconnection could lead to a kind of resonance that might explain the similarities in the timescales of the Atlantic and Pacific modes.

The model's North Atlantic mode may also affect the tropical Atlantic Ocean. The tropical Atlantic exhibits strong variations at decadal timescales. Several authors (e.g., Servain 1991; Metha and Delworth 1995; Chang et al. 1996) have claimed the existence of a dipole-like decadal oscillation, characterized by opposite changes on both sides of the thermal equator. Since there are also strong arguments speaking against the existence of a dipole-like variability mode (Houghton and Tourre 1992), this issue is still very controversial. Our coupled model does not support the existence of an Atlantic tropical dipole. However, the northern Tropics are strongly influenced by the southern branch of the North Atlantic decadal mode discussed in this paper, while no influence could be found south of the equator. The southern tropical Atlantic exhibits only irregular variations independent from the northern regions.

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