A Comparison of Decadal Climate Oscillations in the North Atlantic Detected in Observations and a Coupled GCM

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ABSTRACT

Decadal climate variations in the Atlantic Ocean found in observational fields and a coupled general circulation model (CGCM) are investigated. In particular, physical processes responsible for the phase reversal are compared. Observed and modeled decadal variations have dominant periodicities around 12.3 and 9.9 yr, respectively. Both variations show similar spatial features: sea surface temperature (SST) anomalies in the western subtropical Atlantic sandwiched by those with opposite sign to the north and south, and a dipole of sea level pressure anomalies, which resemble the North Atlantic Oscillation. Their temporal evolutions are, however, different from each other, suggestive of different dynamics of the oscillation. In the observations, SST and surface-layer (0–100 m) temperature anomalies move eastward from the subtropical western Atlantic to the European coast along the Gulf Stream. Northward propagation of SST anomalies are also found along the western boundaries including the Gulf of Mexico. A budget analysis for the temperature equation shows that these features are the manifestation of the advection of SST anomalies by the mean current, which acts to switch one phase of the oscillation to another. Anomalous gyre intensity appears to have little contribution to the phase switching process of the near-surface variability, although the influence of the anomalous gyre is found in the lower subsurface up to 500 m. In contrast, SST anomalies in the CGCM are more strongly tied with subsurface temperature anomalies that propagate westward, consistent with a slow gyre adjustment by the baroclinic Rossby wave propagation. The wave-induced advection acts to change the phase of SST as well as the subsurface temperature anomalies in the model. Subduction of temperature anomalies is found to occur on decadal timescales both in the observation and in the model over the eastern basin where the winter mixed layer is deepened, although the consequence of such a process is not very clear.

In agreement with previous studies, it is suggested that the atmosphere-ocean interaction is important for the decadal variability. The anomalous heat flux originated from the wind-evaporation feedback appears to play a dominant role in the formation of the tripolar structure of oceanic thermal anomalies both in observations and in the CGCM. On the other hand, the dominant timescales of observed and simulated decadal modes are largely dominated by the mean subtropical gyre velocity, and by the propagation speed of long Rossby waves, respectively, both of which happen to have similar timescales.

1. Introduction

In recent years, numerous studies have appeared on decadal to interdecadal climate variations, using observational data and coupled general circulation model (CGCM) experiments.

Although the general aspect of decadal/interdecadal variations is complicated (see a review by Latif 1998), causal mechanisms of an interdecadal variability in the North Pacific (Trenberth and Hurrell 1994; Miller et al. 1994; Tanimoto et al. 1993; Graham 1994) and a decadal variability in the North Atlantic (Bjerknes 1964; Deser and Blackmon 1993; Kushnir 1994) are discussed in terms of the atmosphere-ocean interaction and dynamics of the wind-driven ocean circulation. With the use of a CGCM, Latif and Barnett (1994, 1996) first proposed a mechanism in which slowly varying gyre intensity induced by midlatitude stress curl anomalies can act as a delayed-negative feedback for the phase reversal of the Pacific interdecadal variation. Namely, the transient ocean response to the wind stress anomalies that have forced by the anomalous sea surface temperatures (SSTs) in the central North Pacific leads to the

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reduction of the initial SST anomaly, and consequently brings the phase reversal. The basic concept for the phase transition mechanism is supported by another model (Robertson 1996), observational (Zhang and Levitus 1997), and theoretical (Jin 1997; Münich et al. 1998) studies although the details are subject to variation. Nevertheless, they all recognize that the adjustment of the subsurface gyre explained by the spinup theory (Anderson and Gill 1975) is essential to the interdecadal oscillations.

The hypothesis of the wave adjustment described above is applied to decadal variations of the North Atlantic Ocean in a CGCM by Grötzner et al. (1998). For the Atlantic decadal variations, Weng and Neelin (1998) proposed a theory of oscillation, which consists of the transient response of the subtropical gyre to the anomalous wind stress and the atmosphere–ocean interaction. Unfortunately, however, it is not clear if the gyre adjustment is a unique mechanism for the decadal oscillations. Observational evaluation of the dynamics of decadal variability is extremely difficult and has to be complemented by simulations. Molinari et al. (1999, manuscript submitted to J. Climate) examined decadal variations in their coupled model in comparison with observations, and concluded that different mechanisms appear to work upon decadal oscillations in the two fields: wave adjustment of the subtropical gyre in the model and local Ekman pumping in observations. On the other hand, Sutton and Allen (1997) showed that SST anomalies in the western subtropical Atlantic move eastward to the European coast along the Gulf Stream, and they speculated that this traveling feature is explained by the advection of SST anomalies by the mean current.

In this paper, we compare the decadal variations in the North Atlantic derived from observed surface fields with those in a coupled general circulation model, hoping to gain insights into the mechanisms of variations. Investigations not only of similarities but also of dissimilarities between observed and simulated decadal variations may give clues to understand the dynamics of the decadal variability. The focus is placed on physical processes responsible for the phase reversal, especially, the relative importance between the advection of mean temperatures by the anomalous currents and the advection of the temperature anomalies by mean currents. Also, associated subsurface temperature variations including the anomalous subduction both in observations and the CGCM are described because, except for works by Levitus et al. (1994) and Molinari et al. (1997), there are few studies of the subsurface decadal variability in the North Atlantic in contrast to those for the North Pacific (e.g., Deser et al. 1996; Zhang and Levitus 1997; Miller et al. 1998; Tourre et al. 1999). Analyses of this study suggest physics of observed decadal variations that are different from those presented by Molinari et al. (1999, manuscript submitted to J. Climate). Our results show that the different type of advective effects are important for the observed and simulated decadal oscillations while the atmosphere–ocean interaction is responsible for the spatial patterns similar in both fields.

This article is organized as follows. In section 2, data used in this study and the design of the CGCM are described. Results of analyses are presented in section 3. The results are discussed in section 4 and section 5 gives the conclusion.

2. Data and model description

a. Data

The SST data used in this study are referred to as the Global Sea Ice and Sea Surface Temperature (GISST) data, compiled at the British Meteorological Office, with the spatial resolution of $1^\circ \times 1^\circ$ on a monthly basis (Parker et al. 1995). We use the data reduced to a $2^\circ \times 2^\circ$ grid. Although the data are available from 1903 to 1994, only those after 1945 are analyzed because of two reasons; most of the other data used here, which are described below, are available for the last few decades, and shipboard observations for SST have some systematic errors before World War II (Deser and Blackmon 1993).

The subsurface ocean temperature data are derived from the World Ocean Atlas 1994 prepared by Levitus and Boyer (1994). The data consist of temperature profiles from all the available expendable bathythermographs, mechanical bathythermographs, digital bathythermographs, conductivity–temperature–depth probes, and standard station observations. We compiled monthly $5^\circ \times 5^\circ$ gridded data during 1950–92 using the optimal interpolation technique at each 15 standard level down to 500-m level. The depth-averaged temperature for the upper ocean of 0–100 m and for the subsurface ocean of 100–500 m (denoted as $T_U$ and $T_L$, respectively) are calculated. Here $T_U$ is representative of the mixed-layer temperature while $T_L$ reflects the thermocline displacements. While the discussion of the data quality is given by Deser et al. (1996), it should be noted that the sampling of subsurface temperature is poorer than the surface observations (e.g., SST) and therefore one must be careful for the reliability of data. We also use the Levitus and Boyer (1994) climatology of temperature and salinity on monthly $1^\circ \times 1^\circ$ grid boxes in order to calculate the mean geostrophic current velocity and the mixed layer depth (MLD). The procedure is described in section 2c.

Monthly mean sea level pressure (SLP) data, compiled by the Japan Meteorological Agency, are available on a $5^\circ \times 5^\circ$ grid of the Northern Hemisphere from January 1951 to February 1995.

Monthly mean net surface heat flux ($Q$) data provided by Tohoku University are calculated using the surface meteorological data collected in the Comprehensive Ocean–Atmosphere Data Set (COADS; Woodruff et al. 1987). The data are given on a $5^\circ \times 5^\circ$ grid over the global ocean and available from January 1950 to De-
December 1990. The procedure to calculate $Q$ is described by Tanimoto et al. (1997) together with the discussion of the estimation errors. The monthly wind stresses ($\tau$) estimated from the aerodynamic bulk formula (Kondo 1975) are calculated using the SST, air temperature, and scalar wind speed data compiled at Tohoku University based on COADS. The sampling method that parameterizes intramonthly components of the bulk flux proposed by Hanawa and Toba (1987) is employed to yield more accurate values of the monthly mean stress.

Monthly sea level data for several tide gauge stations in the western Atlantic are obtained from the data archive at The Florida State University via an anonymous FTP (Sturges and Hong 1995). We use sea level time series at Bermuda (64.5°W, 32.17°N) and Charleston (79.56°W, 32.47°N), having dense observations in time from January 1944 to November 1991 and from October 1921 to October 1991, respectively.

Because the atmospheric anomalies are prevailing in boreal winter, results of analyses presented in the next section are based on anomalies averaged in winter (December, January, and February). Anomalies are defined as the deviations from the monthly climatological values.

b. Model

The coupled model used in this study consists of atmospheric and oceanic parts developed at the Center for Climate System Research (CCSR), University of Tokyo.

The atmospheric part is a global spectral GCM [atmospheric GCM (AGCM)] jointly developed by CCSR and the National Institute for Environmental Studies (Numaguti et al. 1998, manuscript submitted to J. Meteor. Soc. Japan). The AGCM has a horizontal resolution of T21 (triangular truncation at wavenumber 21) and 20 sigma levels in the vertical. The model is equipped with an up-to-date package that includes the cumulus parameterization, cloud-radiation interactions, and surface processes. The detailed description is given by Numaguti et al. (1998, manuscript submitted to J. Meteor. Soc. Japan) and Shen et al. (1998).

The ocean part of the coupled GCM is an ocean GCM (OGCM) developed by Kimoto et al. (1997) originally for the purpose of ocean data assimilation. To resolve the equatorial waves, the OGCM has the latitudinal resolution of 0.5° within 10°N–10°S, and the interval is gradually increased to 2.0° outside of 20°, while the longitudinal spacing is equally set to 2.5°. Since the model is focused on simulating the wind-driven circulation, the structure has two characteristics: concentration of depth levels in the upper ocean (15 among 20 vertical levels are put above 500 m), and employment of the level 2.5 turbulence closure scheme (Mellor and Yamada 1974, 1982). The model also includes the realistic bottom topography, horizontal diffusion, and vertical convective adjustment. The sea ice model is not implemented in the CGCM, and the sea ice distributions are fixed to climatology.

In the CGCM, no flux correction is applied except for the polar region north (south) of 55°N (55°S), where SSTs are relaxed to observed climatologies with the time constant of 30 days because of the absence of the sea ice model. Otherwise, the model domain covers the world oceans. Heat, water, and momentum fluxes at the atmosphere–ocean interface are exchanged once per day. The CGCM is integrated from an initial condition of the isothermal static state for the atmosphere and winter climatological temperature/salinity fields with no motion for the ocean. Of the 106 yr of the coupled integration, the first 16 yr are excluded from the analysis in this study. It is worth noting that the CGCM succeeds in simulating El Niño–Southern Oscillation-like interannual variability with realistic amplitude. The vertically averaged temperatures, $T_v$ and $T_s$, are calculated as in observations.

c. Climatology

Mean states of the North Atlantic surface climate in winter produced by the CGCM are compared to the observed climatology in Fig. 1. Although the model fields (Fig. 1b) show strong westerlies in midlatitudes and meridional gradients of the SST as in observations (Fig. 1a), there are several dissimilarities. The most conspicuous disagreement of the model state with the observations is a lack of the strong SST front along the Gulf Stream around 40°N due to the coarse horizontal resolution. In addition, the strongest SST gradients are found in the central basin near 30°N, which are about 10° south of the observed SST front. The difference in subsurface temperatures are similar to that shown in Figs. 1a and 1b (not shown), and they correspond to the difference in the wind-driven circulation between the observations and the model. In association with the southward shift of the meridional SST gradient, the center of the Icelandic low in the CGCM is located near 50°N, as can be inferred from the wind stress field. This is also shifted southward by 10° relative to the observation.

The observed mean subsurface currents are obtained from the geostrophic relation in which the density $\rho$ is determined by using the climatological temperature and salinity data of Levitus and Boyer (1994) on a 1° × 1° grid, and a polynomial equation of state proposed by UNESCO (1981). We assumed that mean geostrophic current $\nabla \psi$ is zero at the 1000-m depth and the current velocity is calculated by an upward integration from this level. In the shallow waters near the coastline, currents are not computed. Also, the MLD is defined by the depth where the density is equal to that calculated using the surface salinity and temperature minus 0.5 K. The computed currents at 100 m and MLDs are shown in Fig. 1c. The subtropical and subpolar gyres are reasonably reproduced with the strong Gulf Stream. The distribution of MLD is similar to that shown in Qiu and Huang (1995), except for the Labrador Sea.
The currents at 95 m and MLDs in the CGCM is also shown in Fig. 1d for comparison. The double-gyre structure is properly reproduced. However, several disagreements are conspicuous; the Gulf Stream is shifted southward and is not strong enough. Moreover, the whole gyre current is much weaker than the observation. These flaws are attributed to the fact that the CGCM fails to simulate the surface thermal structure, particularly the strong localized temperature gradient (Fig. 1b). The MLD is also different from the observation; the deepest region is found in the western basin not in higher latitudes. This suggests that the model underestimates the turbulent mixing, which deepens the mixed layer in the northern latitudes.

3. Observed and simulated decadal variabilities

a. Signal detection

We first perform the conventional empirical orthogonal function (EOF) analysis to winter-mean SST anomalies (SSTA) in the Atlantic Ocean in order to detect decadal variations as found by Deser and Blackmon (1993). The domain of the analysis is extended from 10°N to 80°N and from 100°W to 20°E, and the covariance matrix is used to calculate eigenvalues/eigenvectors.

Figure 2a shows a spatial pattern of the leading EOF for the observed SSTA during 1945–94, which accounts for 21.3% of the total variance and is separated well from the higher modes. The EOF2 will not be discussed further in this.
study although this mode is also statistically independent of the higher modes. In the leading EOF, on the other hand, the SST front lies between negative and positive maxima (Fig. 2a). The time coefficient, or the principal component (PC), of the leading EOF shows a dominant periodicity in the vicinity of 12 yr, which is significant at the 95% level and has the largest power among the spectra of the first 8 PCs (Fig. 2b), and the decadal variability is visible in the PC time series shown in Fig. 2d. This existence of distinct decadal variations in SST anomalies motivates us to apply a digital filter to the data for removing the background “noise,” such as interannual and interdecadal variations. Murakami’s (1979) bandpass filter with the half power points at 5 and 20 yr is employed for this purpose. A spatial pattern of the leading EOF of the bandpass-filtered SST anomalies is shown in Fig. 2c to be compared with Fig. 2a. It is shown that the EOF1 for the filtered SST anomalies preserves the tripole structure found in Fig. 2a, while the fractional variance increased to 28.1%. This result is insensitive to slight changes in the filter design. Associated leading PC superimposed on the original PC (Fig. 2d) represents that the filtering procedure successfully extracts oscillatory signals on decadal scales, which have nearly four cycles during recent five decades. While the “oscillations” are not so regular as one would wish during the period investigated, it is noted that the oscillatory feature is clearer if the entire GISST record of 92 yr is used for the EOF analysis (not shown). Hereafter, the leading PC of filtered SST anomalies (denoted as the observed PC1) is used as a key variable for analyses of temporal evolutions. Rectangles
drawn in Fig. 2c, which cover the two maxima in the western subtropical Atlantic and the central North Atlantic, are used for analyses in section 3c as two key regions, referred to as O1 and O2, respectively.

Similarly, the EOF analysis is applied for SST anomalies of the CGCM. The spatial pattern of the leading EOF (17.4% of the total variance, not shown) is a monopole, having the maximum on the strongest SST gradient, which is found in the more southeastward and broader area than observations, and thus the EOF1 of model SST anomalies corresponds to the observed EOF2. The pattern of the second EOF, with 11.8% of the total variance, is shown in Fig. 3a. It is reminiscent of the observed leading EOF with respect to the tripole structure, although the tripole is somewhat distorted (e.g., negative signals in the eastern basin near 30°W, 60°N, and 30°W, 25°N, which disturb the tripole). The power spectra of the PCs, as shown in Fig. 3b, are whiter than the observed PC spectra (Fig. 2b). However, there are conspicuous peaks at 5.8- and 9.9-yr periods in the first and second PCs, respectively. As in the observations, the EOF analysis to the bandpass-filtered fields of the CGCM gives the same two leading modes as the original EOFs. The second EOF for the filtered fields (Fig. 3c) shows a pattern similar to the unfiltered counterpart (Fig. 3a), except for the clearer tripole structure. The filtered EOF2 of the model corresponds well to the observed EOF1, although the former has a pattern more zonally elongated than the latter. Geographical relationship between negative/positive maxima of the EOF2 and the SST front in the model is the same as in observational results, that is, they straddle the Gulf Stream extension. Two regions in the model fields corresponding to the O1 and O2 are also defined in Fig. 3c (denoted as M1 and M2). Also, the PC for the filtered EOF2 (hereafter referred to as the model PC2) is used as the reference time series.

To show the atmospheric circulation anomalies associated with the decadal SST variations, the observed PC1 and model PC2 are regressed with bandpass-filtered SLP and wind stress anomalies of the observational and simulated fields, respectively (Fig. 4). Units of the SLP and wind stress in Fig. 4 are hPa and N m⁻², and the fields corresponding to −1 K of the SST anomalies in O1 and M1 are displayed. Both regression patterns resemble a negative phase of the North Atlantic Oscillation (NAO), a meridional dipole in SLP and anticyclonic/cyclonic anomalies in wind stress. While the amplitudes of the regression in the observation and model are similar to each other, it should be noted that the decadal-scale regression patterns are not exactly projected onto the NAO displayed by Hurrell (1995; e.g., his Fig. 1) because of the southward shift of the present pattern by 10°. If the NAO is an inherent mode of variability over the North Atlantic atmosphere on time-scales longer than a season (e.g., Kushnir and Wallace 1989), the slight difference of the pattern in Fig. 4 with the classical NAO may indicate that the SLP and wind stress anomalies in association with the decadal variation are partly maintained by the thermal forcing from the underlying sea surface.

### b. Time evolution of thermal anomalies

Using the EOFs of SSTA, decadal modes are identified both in observations and in the CGCM. To examine the temporal evolution of the decadal modes, we calculate lag regressions of filtered SST, Tₓ, and Tᵧ with reference to the PCs. While it is more appropriate to use Tₓ and Tᵧ fields in order to describe decadal cycles of upper-ocean thermal anomalies, SST fields are preferred for observations for the abundance of available observations. The lag regression maps of observed Tₓ do show similar features but are slightly noisier than those of SSTA.

Figure 5 shows lag regression maps between the observed SST (filtered) and PC1 with lags from −5 to 0 yr (approximately half a cycle of the dominant periodicity). Clear transitions in the decadal cycle of anomalous SSTs is found among the sequence of figures; at year −5, the pattern is nearly opposite to the EOF1 (Fig. 2a). SSTs over the O1 area are anomalously warm while those over the O2 area and the northern tropical Atlantic show negative anomalies of about 1 K. The tropical feature seems to be a part of the so-called decadal dipole (Mehta and Delworth 1995) in the Tropics with positive anomalies in the southern tropical Atlantic. The simultaneous regression map in Fig. 5 shows remarkable similarity to the composite SST anomalies with respect to the tropical dipole (e.g., Xie and Tanimoto 1998, their Fig. 1). The relationship between the North and tropical Atlantic decadal modes will be discussed in section 4. During years −4 and −3, positive SST anomalies in O1 move eastward to the central basin as has been pointed out by Sutton and Allen (1997), and by the year −1, they are intensified and extend to the north and south. At this time, negative anomalies are found in O1 and consequently the tripole same as in the EOF1 becomes dominant. The evolution from the year 0 to +5 shows a transition similar to Fig. 5 but with reversed signs (not shown). The regression patterns in Fig. 5 manifest that the decadal SST anomalies are propagative, although those during the transition phase (i.e., during −4 and −2 yr) have smaller amplitude and the correlations are less significant. As noted previously, lag regressions of the observed PC1 with surface-layer temperature anomalies represent similar properties as in Fig. 5, but are noisier due to the sparseness of the subsurface data. It should be noted, however, that the subsurface temperature anomalies, Tᵧ, show the temporal evolution somewhat different from that of SST and Tₓ anomalies (not shown). Namely, the Tᵧ anomalies propagate westward in the subtropics, but not eastward. The difference of the temporal evolutions in the surface- and subsurface-temperature anomalies is discussed in detail in section 3c.
Fig. 3. Same as in Fig. 2, except for the second EOFs for 90-yr SST anomalies in the CGCM. The light (dark) shading in (a) denotes a region of the meridional gradient of mean SSTs exceeding 0.6 K (0.8 K). It should be noted that the key regions in the model are not identical to those in Fig. 2, as presented in (c) (the M1 region for 50°–70°W, 20°–30°N and the M2 region for 40°–60°W, 40°–50°N).

In addition to the eastward movement of SST anomalies, a northward migration of SST anomalies is found in Fig. 5 from the northern tropical Atlantic to the O1 region along the western boundary. For instance, positive anomalies in the Tropics near 60°W, 10°N in the year -5, which are at the western edge of the subtropical anomalies, turn to the north and seem to help organize the next positive anomalies in O1 by the year -1. This northward movement is confirmed by a Hovmöller diagram of bandpass-filtered SST anomalies along the west coast, as shown in Fig. 6. Distinct northward propagations of SST anomalies are found throughout the period except for the decade between 1955 and 1965, when interannual variations dominate the decadal cycles (Fig. 2d). The amplitude of the anomaly exceeds 0.5 K (roughly half the magnitude of interannual variations) and the propagation speed is estimated as 7.2 cm s⁻¹ with the anomaly taking 5 yr to reach from point A to B. The propagation is along the North Equatorial Current (see Fig. 1c), and the phase speed is close to the mean geostrophic current speed in the region (~10 cm s⁻¹). This suggests that the SST anomalies in the O1 region are advected by the mean current, but it is not clear how the decadal anomalies could survive against various damping mechanisms of shorter timescales.

Although the decadal variability in the model fields also reveal the tripole (Fig. 3c), the time evolution is different from the observed decadal cycle found in Fig. 5. Shown in Fig. 7 is lag regressions between the model PC2 and decadal temperature anomalies for the upper 100 m, \( T_{\text{c}} \), in the CGCM. Similar spatial patterns are obtained for the SSTanomalies and \( T_{\text{c}} \) anomalies (figures not shown). Unlike the observed decadal cycle, positive anomalies in the M1 region in the year -5 do not move eastward, but rather decay moving slightly to the west and northwest until the year -1. Instead, negative
FIG. 4. (a) Linear regression patterns of observed winter SLP (contours) and wind stress (vectors) anomalies with leading PCs of Atlantic SSTs (Fig. 2d) based on the filtered data. Units are hPa and N m$^{-2}$, and values represent anomalies corresponding to $2\,^\circ$K of SST anomalies in the O1 region. Contour interval is 1 hPa and the shading denotes areas having significant correlations at the 5% level. The unit vector for wind stress anomalies is shown at the bottom of the figure. (b) Same as in (a) except for the regression of model fields with the second PC of Atlantic SSTs in the CGCM (Fig. 3d).

Anomalies in the M2 region move clockwise and propagate westward to the M1 region during years $-4$ and $-1$. Similar propagation but with opposite signs is found in the regression maps with positive lags (not shown). Also, the sequence of regression maps for SST and $T_L$ anomalies shows a character almost identical to Fig. 7. The northward propagation of SST or temperature anomalies along the western boundaries such as in Fig. 6 is not observed in the CGCM.

A comparison of Figs. 5 and 7 indicates that the time evolutions of the observed and simulated thermal anomalies in the surface layer are considerably different from each other, although both figures show the tripole structure in the mature phase presented by simultaneous regressions of temperature anomalies with the reference PCs. In the observations, the eastward migration (Fig. 5) and northward propagation of SST anomalies (Fig. 6) suggest that the advection of SST anomalies by the mean currents plays an important role in the decadal mode, because propagations are found on the main path of the subtropical gyre. In contrast, the westward movement of temperature anomalies in the model decadal mode (Fig. 7) seems to be consistent with the Rossby wave dynamics referred to in the introduction.

c. Budget analysis

A phase relationship between thermal anomalies and several components that affect the anomalies may give a valuable clue to the mechanism of the decadal oscillations. For this purpose, we calculate a linearized budget for temperatures averaged in the mixed layer. The budget analysis presented here focuses on the phase relationship between the budget terms and the corresponding temperature anomaly, but not an exact assessment of the term balance.

According to Miller et al. (1994), the thermodynamic equation for the mixed-layer in which the temperature is assumed to be uniform (equal to SST), is written as

$$ \frac{\partial T_U}{\partial t} + \mathbf{V} \cdot \nabla T_U + \frac{w_e}{H} \frac{\partial T}{\partial y} = -\frac{Q}{\rho c_p H} + \kappa \nabla^2 T_U, \quad (1) $$

where $T_U$ denotes the mixed-layer temperature defined by the depth average in the upper 100 m, and $\mathbf{V}$ and $w_e$ are the horizontal current and entrainment velocity, respectively. Here $\rho$ and $H$ represent a constant seawater density and MLD (1035 kg m$^{-3}$ and 100 m, respectively). $Q$ is the net heat flux at the ocean surface, and $\Delta T$ indicates the temperature difference between the mixed layer and the underlying layer. Equation (1) is linearized as $(\bar{T} + \bar{\phi}) = (\bar{T} + \bar{\phi})$, where the overbar denotes the mean state and the prime is the anomaly. The current, $\mathbf{V}$, is divided into two components, that is, geostrophic ($\mathbf{V}_g$) and Ekman parts. The perturbation temperature equation,

$$ \frac{\partial T_U}{\partial t} = -\mathbf{V}_g \cdot \nabla T_U - \nabla_x \cdot \nabla T_U + \frac{\tau_x}{\rho f H} \frac{\partial T_U}{\partial y} + \frac{\tau_y}{\rho f H} \frac{\partial T_U}{\partial y} $$

$$ - \frac{w_e}{H} \frac{\Delta T}{\partial y} - \frac{\tau_y}{\rho c_p H} \frac{\partial T_U}{\partial y} + \kappa \nabla^2 T_U, \quad (2) $$

is regarded as a linear combination of several temperature tendencies shown in the right-hand side of Eq. (2). Here $\tau_x$ and $\tau_y$ are the mean and anomalous zonal wind stresses, respectively. For convenience, the first term, advection of the mean temperature by the anomalous geostrophic current, is referred to as the gyre adjustment term. This term is responsible for the scenario by Latif and Barnett (1994, 1996) as described previously. The notation of the advection term in this study refers to the
Fig. 5. Lag regressions of observed SST anomalies with the leading PC of SSTs, (a) displaying from lag at -5 yr to (f) 0, where negative lags denote lagging of the PC. Areas surrounded by white lines denote the region where the correlation coefficient is significant at the 5% level. Regression and correlation are based on filtered winter fields.
Fig. 6. (b) Hovmöller diagram of observed SST anomalies along the western boundaries and (a) grids used to depict the figure denoted as black squares. The points A and B drawn in (a) coincide with edges of (b) as shown at the bottom. In (b) the contour interval is 0.1 K and negative anomalies are shaded. Climatological SST isotherms are superimposed on (a) with every 2 K.
Fig. 7. Same as in Fig. 5 except for regressions between surface-layer temperature ($T_u$) anomalies and the SST PC2 in the CGCM.
second term, called interchangeably the advection of the temperature anomaly by the mean geostrophic current. A sum of the third and fourth terms is denoted as the Ekman term while the zonal Ekman transport is ignored. The fifth and sixth terms together are called the entrainment term. The last two terms are the anomalous heat flux and diffusion effects, respectively. Actual calculation of each tendency is based on the spherical coordinate not Cartesian in Eq. (2), and the bandpass filter is applied after the calculation as in the previous section. Due to the difficulty in calculating the entrainment velocities in the observation and the CGCM, the entrainment term is not calculated. The advection term in the observation is computed using SSTA instead of the mixed layer temperature anomalies because of the the assumption of the temperature uniformity in the mixed layer and the coarse resolution of subsurface temperature fields.

Figures 8a and 8b show lag correlations between observed winter SSTA and three budget terms in Eq. (2), and the advection, Ekman transport, and anomalous heat flux terms, in O1 and O2 regions. First of all, it is reconfirmed that the observed decadal oscillation has the dominant periodicity of about 12 yr from the autocorrelation functions of SST anomalies, which cross the zero axis at years -3 and +3. Both the heat flux and Ekman transport terms are roughly in phase with SST anomalies, although the heat flux anomaly is leading the SST anomaly by 1 yr in O1. On the other hand, correlations between SSTA and the advection term reveal a strong asymmetry with respect to the lag 0. They correspond to the in-phase relation of the advection term with the SST tendency, rather than SST anomaly itself. Thus Figs. 8a and 8b suggest that, in the observations, anomalous Ekman transports and heat fluxes in association with circulation anomalies in the atmosphere (Fig. 3a) to amplify SST anomalies, whereas the advection contribute to the phase transition of the decadal oscillation through the delayed-negative feedback.

The budget calculation also enables a quantitative estimation of the relative contribution of each term to the SST tendency. Amplitudes of budget terms in O1 and O2 regions are inferred from the standard deviation (std dev) of the time series as shown in Table 1 together with std dev of decadal SST anomalies. Standard deviations of SST anomalies have similar values in the two regions (0.243 and 0.318 K) and the heat flux forcing dominates in both regions (0.794 and 0.784 K yr⁻¹). The Ekman components, in which the first term of the budget is dominant, are smaller than the heat flux and suggestive of their secondary importance in the atmosphere-ocean interaction. The advection terms have the smallest amplitudes of the three terms. However, this partly comes from the strong seasonality in the surface atmospheric (i.e., wind and heat flux) anomalies, which show the largest variance in winter. The decadal components of wind and heat flux anomalies extracted from the monthly fields that exclude the seasonality have amplitudes roughly one-fifth of those based on the winter-mean data used in this study.

The observed gyre adjustment term cannot be calculated accurately because salinity observations are not sufficient to reconstruct density fields for the evaluation of geostrophic current anomalies. However, it is possible to calculate temperature-induced current anomalies (see section 2c) using the monthly temperatures during 1950–92 above the 500-m level. The current anomaly calculated in this way may differ from an actual anomaly over regions where salinity has a significant influence to the density anomaly; therefore the anomalous current computed here is distinctively denoted as \( \vec{V}_a \). Because the contribution of salinity to the density anomalies is not dominant in the subtropics as a whole, \( \vec{V}_a \) is used to check the phase relationship between SST anomalies and the anomalous gyre intensity. In addition, another observable variable is employed: the sea level difference, \( \Delta \eta = \eta_t - \eta_s \), where \( \eta_t \) and \( \eta_s \) denote winter sea level anomalies at Bermuda and Charleston. Based on the assumption that the anomalous gyre velocity is mainly generated by the first baroclinic mode (Sturges and Hong 1995; Killworth et al. 1997), \( \Delta \eta \) is proportional to \( \vec{V}_a \), northward current velocity, in the O1 region. Figure 8c shows lag correlations of SST anomalies in O1 with \( \vec{V}_a \), integrated over the upper 300 m in the same region, \( \Delta \eta \) and the stress curl anomalies in 20°–50°W, 30°–50°N. It is revealed that the basin-scale curl anomalies reach negative (anticyclonic) maxima leading the positive maxima of the O1 SST by 1–2 yr. A negative correlation between the SST and stress curl anomalies at lag 0 represents the relationship between the anomalous SST pattern and the NAO-like anomalies such as in Figs. 2c and 3a. It should be noted that \( \Delta \eta \) and \( \vec{V}_a \) are in phase and reach their maxima at the lag +3 yr. This implies that the meridional geostrophic current in O1 is anomalously weak before the peak of subsurface temperature anomaly in the region, therefore not contributing to the warming.

To summarize, Fig. 8c is interpreted as follows: negative curl anomalies in the midlatitude leading the positive SST anomalies in the O1 region by 2 yr spin up the subtropical gyre. The interval between the negative

<table>
<thead>
<tr>
<th>( T_o )</th>
<th>Adjustment</th>
<th>Advection</th>
<th>Ekman</th>
<th>Heat</th>
</tr>
</thead>
<tbody>
<tr>
<td>O1</td>
<td>0.243</td>
<td>0.199</td>
<td>0.449</td>
<td>0.794</td>
</tr>
<tr>
<td>O2</td>
<td>0.318</td>
<td>0.107</td>
<td>0.639</td>
<td>0.784</td>
</tr>
</tbody>
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1 Since our ocean model is a level model unlike Miller et al. (1994), the entrainment velocity is not a diagnostic variable.
peak for curl anomalies and positive peaks for $\Delta \eta$ and $\bar{v}'$ (4–5 yr) is consistent with the transit time of long Rossby waves around 30°N shown by Killworth et al. (1997). Thus Fig. 8c strongly suggests that the gyre adjustment does not work to switch the phase of the observed decadal SST anomalies in the O1 region. The results shown in Fig. 8 are insensitive to the slight change in the location of reference regions, since O1 and O2 have large extents relative to the data resolution.

The budget for the CGCM fields is also calculated, but the horizontal velocity is not divided into the geostrophic and Ekman component due to the difficulty in such a separation from the simulated velocity. Lag correlations of $T_u$ with the gyre adjustment, advection, and anomalous heat flux terms in the M1 and M2 regions are shown in Figs. 9a and 9b. The correlations in the M1 region (Fig. 9a) show that the gyre adjustment (advection) term reaches the positive (negative) maximum 1 or 2 yr before the positive temperature maximum. As in the observation (Fig. 8a), the anomalous heat flux leads the temperature anomaly by 1 yr. In the CGCM, roles of the gyre adjustment and advection effects are reversed in comparison with the observations. The gyre adjustment term is acting to force the phase change of the decadal SST anomaly in M1 while the advection term works to suppress it. The phase relationship of the adjustment and heat flux terms with the temperature anomalies in M2 is similar to that in M1 (Fig. 9b) except for the higher correlations. The advection in the M2 region shows negligible correlations with the temperature anomalies.

Standard deviations of the model budget terms and temperature anomalies shown in Table 2 indicate that the heat flux is dominant as in observations. The gyre adjustment in the CGCM has a magnitude similar to the observed advection term while the model advection is less dominant. The smaller advective effect is probably due to the weaker mean current (Fig. 1d) and smaller amplitude of decadal temperature anomalies.

Shown in Fig. 10 are the time-longitude sections of observed and modeled temperature anomalies at centers of the subtropical gyre (32.5°N and 26°N) for the surface and subsurface layer ($T_s$ and $T_L$). As expected from Figs. 5 and 7, the $T_u$ anomalies propagate slightly eastward in observations (Fig. 10a) while those in the CGCM propagate westward (Fig. 10c). It should be noted that the $T_s$ anomalies propagate westward not only in the model but also in observed fields (Figs. 10b and 10d). The meridional components of the anomalous current at the 100-m level, $\bar{v}'$ and $v'$, also represent the westward propagation, which are in quadrature (i.e., $\bar{v}'$ and $v'$ lead the $T_s$ anomalies by 2–3 yr), which satisfies the thermal wind relationship (not shown). The

![Figure 8](http://journals.ametsoc.org/doi/abs/10.1175/1520-0442%281999%29012%3C2920%3AACODCO%3E2.0.CO;2?journalCode=clim)

**Fig. 8.** (a) Lag correlations of observed SST anomalies with the advection (thick solid), horizontal Ekman (thick dotted), and anomalous heat flux (thick dashed) terms in the O1 region, superimposed on the SST autocorrelations (thin solid). All the series retain only decadal components. (b) As in (a) except for correlations in the O2 region. (c) As in (a) except for correlations of SST anomalies in O1 with $\eta'$ integrated over the upper 300 m in the region, stress curl anomalies in 20°–50°W and 30°–50°N, and $\Delta \eta$. 

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Inconsistency of propagating features between the anomalous $T_L$ and $T_U$ in observations, which is also discussed in section 3b, suggests that the thermal anomalies below the mixed layer have temporal characters different from those in the mixed layer. This implies that the surface-layer thermal anomalies are coupled with the gyre dynamics in the CGCM while they appear to be decoupled in the observations.

In the CGCM, thermal anomalies take about 10 yr to cross the Atlantic Ocean, although the period of oscillation is also about 10 yr. Using a simplified model that contains only the Rossby wave dynamics, Münich et al. (1998) simulated a decadal oscillation in which the period is twice the transit time of Rossby waves. In their results, the zonal wavenumber of pressure anomalies is $\frac{1}{2}$, while Figs. 10c and 10d indicate the anomalous $T_L$ and $v^\prime$ having the zonal wavenumber 1 in our model.

Since the frequency of oscillation becomes twice for the doubled wavenumber, the period and transit time of Rossby waves have similar values.

d. Vertical structure of temperature anomalies

In the previous section, inhomogeneity of temperature in the upper layer has not been considered assuming the uniformity in the mixed layer. In this section, the vertical structure of decadal temperature anomalies is investigated more carefully. As mentioned in the introduction, descending motions of the anomalous temperature, referred to as subduction, on decadal timescales are observed in the North Pacific (Deser et al. 1996; Tourre et al. 1999). Gu and Philander (1997) consider that the subduction and consequent exchange of heat between midlatitudes and Tropics are crucial for the Pacific interdecadal oscillation. In the North Atlantic, the anomalous temperature subduction on decadal timescales has not been reported, although the seasonal subduction referred to as the “Stommel’s demon” is well studied (e.g., Williams et al. 1995; Qiu and Huang 1995).

In the temporal evolution of the decadal modes (Figs. 5 and 7), SST anomalies decay in the eastern basin near 20°W, 50°N in the observations, while $T_U$ anomalies turn to the southwest near 20°W, 30°N in the model. They correspond to the eastern edges of stationary signals in the O2 and M2 regions (Figs. 2c and 3c). Figure 11 represents the filtered subsurface temperature anomalies at 27.5°W regressed with the anomalous SST time series averaged over the regions 20°–30°W, 40°–50°N for the observed and 20°–30°W, 25°–35°N for the simulated fields with a lag from 0 to +3 yr. The simultaneous regressions indicate that the temperature anomalies are confined within the mixed layer, whereas the regressed anomalies lagging the SST anomaly sink as the lags increase. As shown in the top panels, differences in MLDS between the late winter (March) and spring (May) conspicuous in these regions (more than 150 m in observations and 75 m in the CGCM) coincide with the detrainment of the surface water into the seasonal thermocline. A part of the detrained water parcels move into the main thermocline due to the Ekman pumping and the lateral induction. The sinking rates of anomalous temperatures estimated by tracing the maximum of

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2 The demon refers to a process in which the water parcel is detrained from the mixed layer into the main thermocline only at the end of winter when the seasonal thermocline rapidly raises. The name is derived from the Maxwell demon.
anomalies on the regression maps are roughly 75 m yr$^{-1}$ for observations and 50 m yr$^{-1}$ for the CGCM. The former is consistent with the observed subduction rate of 50–100 m in the North Atlantic (Marshall et al. 1993). These results suggest that the subduction occurs even on decadal scales in the eastern basin of the North Atlantic. Depth–time sections of temperature anomalies also reveal the descending motion in the regions (Fig. 12). Differences in the subduction rate between observations and the CGCM are also apparent from different vertical gradient of temperature anomalies. This difference is partly explained by the weaker Ekman pumping velocity in the model (std dev of $1.2 \times 10^{-2}$ m day$^{-1}$ in the region of Fig. 12), which is roughly one-third of the observation (std dev of $3.5 \times 10^{-2}$ m day$^{-1}$).

It is not clear if the subduction process plays an active role in the decadal oscillation. The subduction of temperature anomalies related to the entrainment term in Eq. (2) has more to do with the advection by the mean current rather than to the gyre adjustment. If the temperature anomalies are not dissipated, the water parcels subducted in those regions are advected southwestward along the pycnocline by the subtropical gyre as shown by Williams et al. (1995), and will reappear at the surface near the western boundary. Their results, however, indicate that the timescale of the movement is much longer than that considered here (it takes 7 yr for the parcel to be transported to the central basin near 40°W, 20°N from the subduction region, 30°W, 40°N). Therefore, the advection of subducted signal does not appear to play an essential role in the observed oscillation.

e. Role of the anomalous heat flux

As emphasized in section 3c, the dynamics that drive the decadal oscillations appear substantially different in observations and in the CGCM. Nevertheless, spatial structures of SST and/or temperature anomalies in the surface layer are similar to each other. The resemblance of anomalous patterns may be explained by the atmo-
FIG. 11. (a) Observed temperature anomalies along 27.5°W regressed upon the SST anomalies averaged in 20°–30°W, 40°–50°N with lags from 0 (top panel) to +3 (bottom panel) years, where positive lags denote that the reference SST time series lags the temperature fields. All the regression patterns are based on the decadal components. The values are the temperature anomalies (K) corresponding to the reference SST anomaly of 1 K. The contour intervals are 0.1 K, while contours below ±0.2 K are suppressed. The mean MLDs in Mar and May are shown in the top panel by the solid and dashed lines, respectively. (b) As in (a), except for the lag regressions in the CGCM 90-yr fields, and the reference SST anomalies are averaged over the region 20°–30°W, 25°–35°N.
Figure 12 shows the regression patterns as in Fig. 4, but for the heat flux and zonal wind stress (which is nearly proportional to the zonal wind) anomalies superimposed on the zero contours of climatological wind stresses. It is found that, in both fields, negative wind anomalies in midlatitudes between 40°N and 60°N corresponding to the reduction of westerlies result in less heat fluxes to the atmosphere that can cause positive SST anomalies beneath. On the other hand, positive wind anomalies in the subtropics extend not only to the mean westerly latitudes (i.e., north of 30°N) but also to lower latitudes where the easterly trade winds dominate. Those wind anomalies lead to the anomalous heat losses in the former region due to the intensified westerlies, and to anomalous heat gains in the latter region due to the weakening of easterlies. The dipole anomalies in the near-surface atmospheric circulation shown in Fig. 4 in association with the mean state could therefore largely determine the pattern of underlying SST anomalies. The basin-scale relationship between (latent and sensible) heat flux anomalies and the atmospheric circulation is studied by Cayan (1992) in detail. The features in Fig. 13 are in good agreement with his Fig. 3.

In summary, the dipole atmospheric anomalies similar to each other in the two fields yield the tripole anomaly patterns of heat flux through the wind–evaporation feedback, and thereby form the tripole of decadal SST anomalies. If the NAO-like wind anomalies on decadal scales (Fig. 4) are induced by the decadal SST anomaly pattern (Figs. 2c and 3c), this process is a positive feedback and can determine the amplitude of decadal oscillations.
FIG. 13. Same as in Fig. 4 except for regressions upon (a) observed and (b) model zonal wind stress (contours) and net heat flux anomalies (shadings). The contour interval is 0.01 N m$^{-2}$ and negative contours are dashed. The light (dark) shading represents the heat flux anomalies of more (less) than 10 (210) W m$^{-2}$, where the positive anomaly denotes anomalous heat loss for the ocean. The zero contours of mean (climatological) zonal wind stresses are plotted by thick lines.

4. Discussion

Analyses in section 3 suggest possible mechanisms of decadal cycles in observed and simulated fields. In the observations, the cycle consists of two major processes: the advection of temperature anomalies by the mean current and the atmosphere–ocean interaction. For example, positive and negative SST anomalies in O1 and O2 change to the opposite sign due to the western boundary advection (cf. Fig. 6), then the wind anomalies coupled to those SST anomalies (Fig. 4a) excite the tripole SST pattern including the positive SST anomalies in the subtropics (Fig. 13a). These positive anomalies will be advected to the north and eventually switch the sign of SST anomalies in O1 again. In this cycle, the anomalous heat flux exchange can determine the pattern and amplitude of the mode while the period of oscillation is largely affected by the mean current speed of the subtropical gyre, especially the western boundary current. In the CGCM, it is found that the atmosphere–ocean interaction acts to amplify the decadal mode as in the observations. On the other hand, unlike the observed decadal mode, the advection of mean temperature gradient by the anomalous geostrophic current appears to be responsible for the phase transition. The anomalous current velocity is possibly induced by the wind stress anomalies through the Rossby wave propagation. The period of oscillation in the model is determined by a combination between the zonal wave number and transit time of the Rossby waves depending on the phase speed and the basin width of the Atlantic Ocean.

Why the dominant processes driving the oscillations on decadal scales are different in observations and the CGCM? In other words, what determines which of the two components, $-\nabla \cdot \nabla T_{o}$ or $-\nabla' \cdot \nabla T_{o}$, dominates over the other? First, it is found that the near-surface baroclinicity in the CGCM is considerably weaker than that in observations, although the Sverdrup transport must be similar in two fields due to similar wind stress fields (Fig. 1). This deficiency results from the weak meridional temperature gradient particularly around the Gulf Stream. Consequently, local $\nabla$ in the model is smaller than the observational estimate of $\nabla$ in the upper layer (Fig. 1d). On the other hand, the anomalous gyre velocities associated with the (first baroclinic) Rossby wave have similar magnitudes above the main thermocline in observational and simulated fields. The relative magnitude of $V'$ to $\nabla$ in the model becomes greater than that observed. Since the horizontal gradients of the mean and anomalous temperatures, $\nabla T$ and $\nabla T'$, are on the same order between observations and the CGCM as inferred from the similarity between Figs. 2c,d and Figs. 3c,d, the gyre adjustment term is prevailing in the CGCM while the advection term can dominate in observations. Therefore, it is speculated that the preferred mechanism depends on the mean state of the subsurface ocean, particularly the vertical shear of the Gulf Stream. The similarity in the periods of two decadal modes are probably due to the Rossby wave speed in the CGCM having a value similar to the mean subtropical gyre speed in observations. The above discussion is probably applicable to decadal cycles in other CGCMs. For example, Grötzner et al. (1998) have shown an oscillation with a 17-yr period with spatial SST and SLP patterns similar to the decadal mode in our coupled model. Their decadal mode is primarily driven by the wave adjustment as in our CGCM, and this seems to be derived from the loose meridional SST gradient in their model, which results in mean currents

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*This similarity has actually been confirmed by solving the vertical structure equation for the two fields (not shown).*
weaker than observations. On the other hand, the period of simulated decadal modes in Grötzner et al.'s (1998) model is somewhat longer than ours. In their model, the decadal SST mode reveals an elongated pattern in the longitudinal direction. This implies that the Rossby waves in their model have a zonal structure larger than that in our CGCM, and agrees with the longer period of their oscillation. Theoretically, the zonal structure of oceanic Rossby waves depends on the zonal scale of the anomalous wind stress curl (Weng and Neelin 1998). The period of the dynamically coupled oscillation found in the model is determined not only by the transit time of long Rossby waves but also by the spatial pattern of the wind stress anomalies, which may be generated by the midlatitude atmosphere–ocean interaction. As noted in section 3a, the SLP dipole related to the decadal SST modes both in observations and in the model is somewhat different from the classic NAO pattern. The low-frequency variability of the NAO is associated not only with the Atlantic SST fluctuations but also with other climate variability such as the Southern Oscillation (Rogers 1984). An observed index of the NAO (Hurrell 1995) does not show the spectral peak around 12 yr. These differences suggest that the anomalous atmospheric circulation patterns on the decadal scale are not exactly projected onto the internal atmospheric variability (i.e., the NAO). If so, the decadal atmospheric anomalies are sustained, at least partly, by forcings such as the anomalous SSTs in the extratropics, as suggested by Rodwell et al. (1999). Although the influence of midlatitude SST anomalies on the atmosphere has been examined with great detail of interest (e.g., Palmer and Sun 1985), there remain many uncertainties, for example, about the role of the anomalous eddy feedback and about the effective regions of the SST anomalies. The dynamical processes of the midlatitude atmosphere–ocean interaction on the decadal timescales should be investigated further. On the other hand, it is an important viewpoint that the variability of the stochastic atmosphere without forcing of the SST anomalies can lead to low-frequency variabilities longer than a decade in the ocean (Frankignoul 1985, and references therein). The simple model experiments that introduce the stochastic atmospheric interactions with the atmosphere–ocean interaction into the models (Münnich et al. 1998; Weng and Neelin 1998) showed that the stochastic atmosphere contributes to maintain the decadal-scale variances.

It is known that there is a decadal variability in the tropical Atlantic surface climate (e.g., in SSTs, winds, and heat fluxes), which shows opposite polarity between the north and south of the equator and has temporal characteristics similar to the North Atlantic decadal variations as shown in Fig. 1d (Carton et al. 1996). The dipole of tropical SST is also observed in our results (Fig. 5), in which the northern anomalies are joined with the tripole of the midlatitude SST decadal mode in the subtropics between the equator and 20°N. Chang et al. (1997) reproduced the decadal dipole of tropical SST anomalies using the coupled model with an empirical atmosphere. Their results imply the importance of local atmosphere–ocean interaction for simulating the dipole of tropical SST anomalies. On the other hand, Xie (1999) showed that the preferred decadal timescale of the variation is not determined in the Tropics, but rather affected by the forcing from midlatitudes. The advection mechanism for the North Atlantic decadal variation shown here for the observed fields matches this Xie’s (1999) result. Namely, forming the sandwich pattern of SST anomalies is forced by the atmosphere–ocean positive feedback in the North Atlantic (see section 4e). The SST anomalies south of 20°N, which is a part of the sandwich, induce the southwesterly wind anomaly in the Tropics (Chang et al. 1997) and at the same time, move northward along the western boundary (Fig. 6), presumably advected by the ocean boundary currents. Thus it is suggested that the decadal variations in the northern North and tropical Atlantic Oceans are coupled, with the latter dependent on the former. This view for the Atlantic decadal variability is similar to the “pan-Atlantic decadal oscillation” presented by Xie and Tanimoto (1998). In the CGCM, the wave adjustment process works only within the midlatitudes (Fig. 7), although the concomitant wind anomalies could generate SST anomalies in the Tropics in the mean easterly wind regime (Fig. 13). In addition, SLP anomalies in the subtropics shown in Fig. 4 are weak relative to those observed. As a consequence of these differences, the decadal dipole in the CGCM is more tenuous in the Tropics as shown in Fig. 7.

5. Conclusions

In this article, a comparison has been made between decadal variations of the North Atlantic surface climate found in observations and the CGCM. Decadal modes detected in 50- and 90-yr long fields in observations and the model, respectively, show irregular oscillations with dominant periodicities of 12.3 and 9.9 yr. Their spatial characters are similar to each other: tripoles in SST and surface-layer temperature, which consist of positive (negative) anomalies in the western subtropical Atlantic and negative (positive) anomalies surrounding the north and south, and dipoles in SLP and wind stress that resemble the NAO pattern.

We have investigated the temporal evolutions of the decadal variations in observed and simulated fields paying attention to the dynamics that produce the oscillatory features. The results strongly suggest that the North Atlantic decadal modes in the observations and the CGCM are driven by different mechanisms. The advection of temperature anomalies by mean currents of the subtropical gyre appears to be responsible for the phase transition process in the observed decadal oscillation, while the advection of mean temperatures by the anomalous gyre is crucial for the decadal oscillation in
the CGCM. In both decadal cycles, the anomalous heat flux forcing and horizontal Ekman transport contribute to maintain the anomalies, especially the heat flux anomalies associated with the wind–evaporation feedback have a dominant role in forming the characteristic decadal SST patterns. In the model decadal mode, SST anomalies together with temperature anomalies in the subsurface down to 500 m propagate westward in the midlatitude, indicative of the transient response of the subtropical gyre to the stress curl anomaly through the baroclinic Rossby wave propagation. In contrast, observed decadal SST and temperature anomalies for the upper 100 m in the western subtropical Atlantic move to the east along the Gulf Stream and those in lower latitudes propagate northward along the western boundaries from the Brazilian coast to the western subtropics. The observed temperature anomalies in the subsurface layer propagate westward as in the CGCM, suggestive of the anomalous gyre effect decoupled with the near-surface temperature variability. The different mechanisms between the observed and modeled decadal oscillations are conceivably derived from differences in the mean subsurface states, especially the vertical shear of the subtropical gyre. However, it is not yet proven that other mechanisms (the advection for the model decadal mode and the wave adjustment for the observed decadal cycles) do not contribute at all to the oscillating process and the maintenance of the modes.

The subduction processes on the decadal timescales are found both in observations and the coupled model. They occur in the eastern basins of the North Atlantic where winter (spring) mixed layer is deepened (shallowed), and are consistent with the selective subduction called the Stommel’s demon. It is not clear whether the subducted temperature anomalies are actively involved in the decadal cycles. Although we did not find strong evidence that the subduction plays an important role in the North Atlantic decadal oscillation, further analyses and numerical experiments are needed. Also the potential importance of the extratropical atmosphere–ocean interaction should be investigated in more details.

The results presented in this study caution that the apparent similarity in periodicity and in spatial patterns of extreme phases does not readily imply the success of CGCM simulations since different types of the phase switching mechanisms can operate in the North Atlantic. Clearly, a more detailed observational analysis and refinement of coupled model physics should go hand in hand for a better understanding of the North Atlantic decadal variability.

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