Identifying a Damped Oscillatory Thermohaline Mode in a General Circulation Model Using an Adjoint Model

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ABSTRACT

A damped oscillatory mode of the thermohaline circulation (THC), which may play a role in interdecadal climate variability, is identified in a global primitive equation model. This analysis is done under mixed boundary conditions using an adjoint of the primitive equation model.

The linearized versus nonlinear stability behavior of the model is studied by comparing the adjoint analysis to runs of the fully nonlinear model. It is shown that a steady-state solution obtained under larger amplitude freshwater surface forcing (and hence with a weaker North Atlantic overturning) is unstable, while a steady-state solution with stronger THC is stable. In a certain intermediate parameter regime it is found that the full nonlinear model state may be unstable, while the linearized analysis indicates that the model state is stable. It is proposed that this may be because either the instability mechanism at this intermediate regime is nonlinear or, while the model is linearly stable at this regime, it allows for temporary growth of small perturbations due to the non-normal nature of the problem.

A clear signal of variations is not found in the amplitude of the horizontal gyre circulation, possibly indicating that the gyre effect that was found in THC oscillations in some previous studies may not be essential for the existence of the THC oscillation. The long timescale of the oscillation in the present model also seems to indicate that the gyre effect may not be a main active participant in the thermohaline oscillation mechanism.

1. Introduction

The identification of the mechanism of decadal climate variability has become an important research goal in recent years due to the need to differentiate anthropogenic climate change from natural climate variability. Some of the observed decadal oscillatory climate signals seem to be related to thermohaline circulation (THC) variability (e.g., Kushnir 1994; Delworth and Mann 2000), and there have been several proposals regarding the possible mechanisms of THC variability. Some works have proposed that THC variability is due to a self-sustained oscillation that is internal to the ocean (Weaver et al. 1991; Winton and Sarachik 1993; Chen and Ghil 1995; Cai et al. 1995), others that it is due to a damped oscillatory mode of the ocean driven by stochastic atmospheric forcing (Mikolajewicz and Maier-Reimer 1990; Bryan and Hansen 1995; Griffies and Tziperman 1995; Delworth and Greatbatch 2000), and yet more that this may be a coupled ocean–atmosphere mode (Weaver and Valcke 1998; Timmermann et al. 1998). In addition, the THC variability mechanism was conjectured to be essentially 2D in the depth–meridional plane (Griffies and Tziperman 1995), or 3D, including a horizontal gyre effect (Delworth et al. 1993).

Some of the above works were based on idealized models of varying levels of sophistication, some on ocean-only general circulation models (GCMs), and others used ocean models coupled to various atmospheric representations. When analyzing the results of a coupled ocean–atmosphere GCM to identify the mechanism of THC variability, one faces complications because of the atmospheric noise in such models, which makes it more difficult to isolate the oscillation mechanism. When using an ocean-only model, on the other hand, the simplified atmospheric representation (e.g., mixed boundary conditions; Bryan 1986) may distort the physics of the oscillation and, again, make it difficult to draw conclusions regarding THC variability in the actual climate system. Therefore, complementary studies using a diverse set of models are required for approaching this issue.

The objective of this paper is to provide support to

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the hypothesis that THC variability may be due to a damped oscillatory THC mode that is continuously excited by atmospheric noise, as suggested in the analysis of a coupled GCM by Griffies and Tziperman (1995). There are three main new findings in this paper: we identify a damped oscillatory THC model in a primitive equation (PE) model that seems identical to such a mode proposed in the box model study of Griffies and Tziperman (1995); we discuss the issue of the stability of the oscillatory mode for both a linearized PE model and a fully nonlinear PE model; finally, we show that the mechanism of this damped oscillatory THC mode is essentially a 2D meridional mechanism, and that 3D gyre effects such as described in Delworth et al. (1993) are not an essential part of the oscillation mechanism in our model. The analysis in this paper is based on an adjoint of the GCM, which essentially amounts to analyzing the variability of the GCM linearized about a steady state, and under mixed boundary conditions. Based on our findings, we also point to some future extensions of the present work, in which the adjoint model may be used to calculate optimal modes.

The following sections describe the model and methodology (section 2) and present the results of the analysis of the model experiments (section 3) and conclusions (section 4).

2. Model and methodology

We use the Geophysical Fluid Dynamics Laboratory (GFDL) primitive equations model of Bryan (1969) in a coarse-resolution global configuration similar to that of Bryan and Lewis (1979), with the main difference being that the Arctic Ocean is not included in our model. The model has 12 vertical levels and a horizontal resolution of 4° latitude × 3.75° longitude. For a full model description see Sirkes et al. (1996). For the purposes of the present study, the model was first run to a steady state for 4100 years, with surface restoring to the Levitus temperature and salinity data (Levitus 1982), and using climatological winds (Hellerman and Rosenstein 1983). The restoring timescale for temperature was chosen to be 30 days, with an upper-layer thickness of 50 m. This temperature restoring timescale is also used when running the model under mixed boundary conditions. The existence of oscillatory modes, and their stability, were proposed by Tziperman et al. (1994) to be controlled by the salinity restoring time used to obtain the steady state before switching to mixed boundary conditions (see also Weaver et al. 1991). We performed several experiments using salinity restoring times of 12, 20, 30, and 120 days.

If a damped oscillatory THC mode exists, then a small perturbation to the steady-state solution of the model would result in an exponentially decaying oscillation of the North Atlantic thermohaline circulation, eventually decaying back to the unperturbed steady state. In order to identify such a damped oscillatory mode in our PE model, we would like to run the model under mixed boundary conditions (restoring for temperature and fixed $E - P$ for salinity), linearized about a steady-state solution. The steady-state solution and the $E - P$ forcing to be used are obtained under restoring boundary conditions (Bryan 1986). The signature of a damped THC oscillation would be a phase difference between temperature and salinity oscillations in the water mass formation area in the northern North Atlantic, with the expected phase lag between salinity and temperature (Delworth et al. 1993; Griffies and Tziperman 1995). Such a phase difference is an inherent and essential characteristic of the 2D (meridional depth) THC oscillation mechanism of Griffies and Tziperman (1995). In addition, we will also be looking for a possible 3D effect involving changes in the gyre circulation in the North Atlantic (Delworth et al. 1993).

As a tool for studying the linearized dynamics of the THC about a steady-state solution, we shall be using an adjoint model of the PE model, derived by Long et al. (1989, unpublished report) and used, for example, by Sirkes et al. (1996). Given a numerical general circulation model, the adjoint method can be used for data assimilation (Tziperman et al. 1992), parameter estimation (Tziperman and Thacker 1989), and sensitivity analysis (Hall and Cacuci 1983; Marotzke et al. 1999). A crucial component of the method is a numerical model composed of the adjoint equations of the GCM. The adjoint equations calculate the sensitivity of a scalar ''cost function'' composed of the model solution, to the initial conditions. The sensitivity is nothing but the derivative of the cost function with respect to the model initial conditions and/or model parameters. To calculate the sensitivity of the forward model solution at the final time of its integration, $t = T$, to its initial conditions at $t = 0$, the adjoint equations are integrated backward in time from $t = T$ to $t = 0$. For the purpose of the present study, we choose the cost function to be the meridional advective heat transport across latitude 24°N in the North Atlantic. The heat transport at this latitude is clearly of dynamical importance as far as the THC is concerned (Marotzke et al. 1999) and has been a key quantity derived from observations (Bryden and Hall 1980).

Let us briefly explain what one expects to find in the adjoint solution when the forward model is stable/unstable, or has an oscillatory mode. When the forward model is unstable at the steady-state solution used for the linearization, then a small perturbation to the initial conditions would grow exponentially in time during a forward model run. This implies that the sensitivity of the model solution at any time $t = T$ to perturbations at a previous time $t = T - \Delta t$ grows exponentially in Δt. Consequently, the adjoint solution, which is nothing but the sensitivity of the model solution to initial conditions prior to $t = T$, will also grow exponentially with Δt, that is, grow exponentially *backward* in time. Similarly, a linearly stable solution of the forward model is

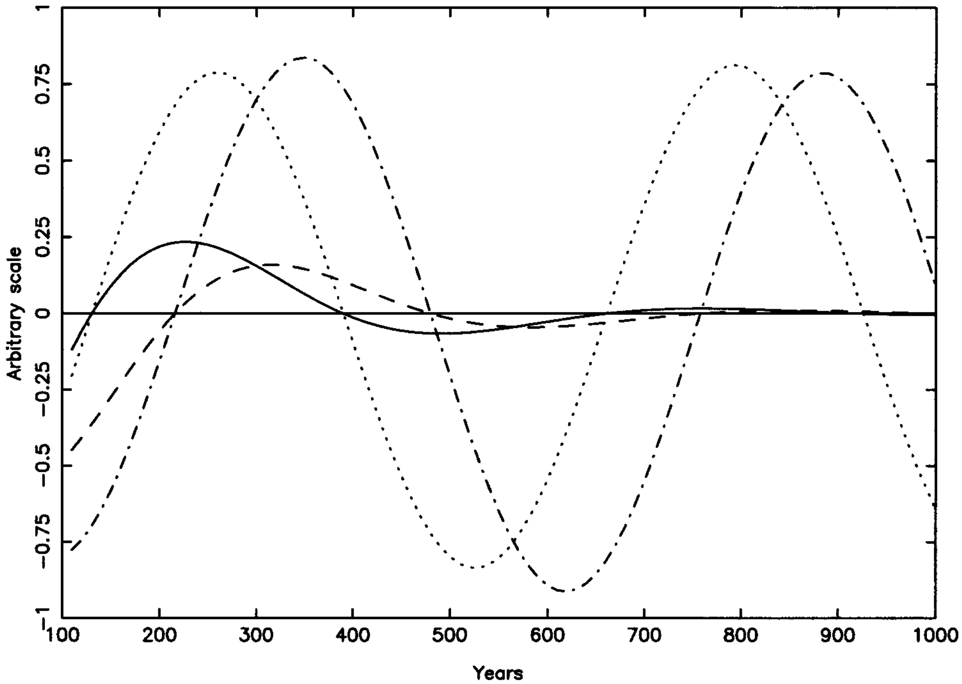

FIG. 1. Adjoint temperature (solid and dotted lines) and salinity (dash and dash-dot) for the experiment using $\gamma_s^{-1} = 30$ days, averaged over a midlatitude region of the North Atlantic (depths of 0–595 m and over 28.89°–37.79°N, 264.38°–361.88°E), as function of time. The dot and dash-dot lines are the time series multiplied by an exponential factor to exactly balance the exponential decay of the original time series.

characterized by an exponential decay of small perturbations, and the adjoint model in this case will have an exponentially decaying behavior *backward* in time. Finally, an oscillatory behavior of the forward model implies that, in a response to a small perturbation, the forward model displays oscillations (growing or damped depending on the stability), and the adjoint model solution would also display such oscillations when viewed backward in time. Our choice of a cost function serves as a forcing of the adjoint model, basically equivalent to an initial perturbation at 24°N. This initial perturbation, being located at a key area for the North Atlantic THC, is able to efficiently excite the various oscillatory modes that exist in the model.

3. Results

In the following two sections we use the results both from the adjoint model and from the forward model to discuss the existence and mechanism of a damped oscillatory THC mode (section 3a) and the mechanism of thermohaline instability (section 3b).

a. Oscillatory mode

Fig. 1 shows time series of the adjoint variables of the temperature (solid line) and salinity (dash) averaged over a midlatitude region of the North Atlantic (depths of 0–595 m; and over 28.89°–37.79°N, 264.38°–1.88°E). Note that these variables are shown backward in time, as obtained from the adjoint model integration. Clearly there is a damped oscillatory behavior and a phase lag between the temperature and salinity, as expected from a THC oscillation (Griffies and Tziperman 1995; Delworth et al. 1993). It is worthwhile at this point to briefly describe the oscillation mechanism from a 2D perspective. We shall later return to 3D effects.

The main components of the oscillation mechanism proposed in Griffies and Tziperman (1995) are the positive salinity and negative temperature feedbacks and the phase lag between the northern and southern areas of the North Atlantic Ocean. Given a small positive initial perturbation to the overturning transport (involving both stronger upwelling at midlatitudes and a stronger northward flow), this anomalous circulation advects the steady temperature and salinity gradients and creates temperature and salinity anomalies. The increased advection of warm and salty midlatitude water to the northern sinking area creates a warm and salty anomaly there. Similarly, the increased upwelling of cold and fresh deeper water to the midlatitude surface, makes the midlatitude surface North Atlantic Ocean fresher and colder. (This linearized description implies that the anomalies are small and the midlatitudes remain warmer and saltier than the deep and sinking area waters). The resulting salinity anomaly increases the north–south density gradient and therefore enhances the positive meridional overturning transport perturbation. The temperature anomaly, on the other hand, reduces the density gradient. Initially, the temperature anomalies are weaker because of the rapid atmospheric restoring of SST anomalies. Eventually the north–south density gradient anomaly and the transport anomaly are weakened by the increasing temperature anomalies and by the advection of the anomalous salinity by the mean transports. The growing circulation anomaly therefore reaches a maximum value and starts decaying. The circulation anomaly approaches zero simultaneously with the density anomaly in the sinking area but, because of the phase lag between the surface midlatitudes and surface sinking areas, there is still a cold and fresh anomaly in the midlatitude surface North Atlantic. This fresh salinity anomaly is advected poleward and causes the salinity anomaly in the sinking area to cross the zero point and become negative. The cycle described above now repeats, but with the temperature, transport, and salinity perturbations of opposite signs.

This oscillatory mechanism is clearly purely 2D meridional. Note that we assume the overturning to depend on the meridional density gradient as was shown to be the case in the GCM experiments of Hughes and Weaver (1994). It also applies to both damped and unstable exponentially growing oscillations (Rivin and Tziperman 1997). In Griffies and Tziperman (1995), the THC variability in the coupled model of Delworth et al. (1993) was explained as a random excitation of such a damped THC oscillatory mode by stochastic atmospheric forcing. The oscillations can become unstable when the meridional surface salinity gradient is stronger, which also implies a stronger $E - P$ forcing and a weaker overturning circulation (Weaver et al. 1991; Tziperman et al. 1994; Tziperman 1997, 2000).

Note that, this being a complex PE model, the linearized forward model solution is characterized by more than a single decaying oscillatory and nonoscillatory mode. After a sufficient integration time, only the dominant mode with the longest decay time survives. The adjoint variables are therefore shown in Fig. 1 from year $t_0 = 100$ to year 1000 of the adjoint integration, after the initial transients are gone and only the dominant oscillatory mode survives. Being the longest-surviving mode, this mode is the main one that would be excited by random atmospheric forcing. (It seems likely that the atmospheric variability, while having its own preferred spatial and temporal structure, will still project at some power on all THC modes; we are thus assuming that the selection of excited modes is not due to the structure of the projected atmospheric forcing, but due to the nature of the available oceanic modes; this needs of course to be verified using full coupled ocean–atmosphere models.) This oscillatory mode is parallel to that found in the simple box models of Tziperman et al. (1994); Griffies and Tziperman (1995); Rivin and Tziperman (1997). The dot and dash-dot lines of Fig. 1 show the adjoint variable time series multiplied by an exponential factor $\exp(t/\tau)$ with the exponential timescale τ chosen to balance the exponential damping time of the oscillations. This allows us to better view the phase relationships between the adjoint temperature and salinity. We shall return in the next subsection to the dependence of the exponential decay time on the basic steady state around which the linearization is performed.

Figure 2 shows zonally averaged sections of the adjoint temperature and salinity in the North Atlantic at three different stages of the oscillation corresponding to oscillation phases of 0°, 90°, and 180°. The solution is again multiplied by an exponential factor, $\exp(t/\tau)$, balancing the exponential decay. Remembering that the adjoint variables basically correspond to the temperature and salinity of the linearized forward model, one can see the salinity perturbations evolving through the interaction with the mean salinity gradients and being advected around the meridional overturning circulation (panels on the right-hand side of the figure). In particular, it is seen how a negative salinity anomaly at an oscillation phase of 0° (upper right panel) is replaced by a positive anomaly 260 years later (lower right panel). The corresponding temperature anomalies are also seen (left panels) to evolve as per the oscillation mechanism described above.

In a study of the THC variability in a coupled ocean–atmosphere model, Delworth et al. (1993) have described an interesting feedback between the THC oscillation and the horizontal gyre circulation as being an inherent part of the oscillation mechanism. The amplitude of the horizontal gyre varies due to changes in the meridional THC, and in turn also advects salinity perturbations and thus affects the meridional THC. In contrast, Griffies and Tziperman (1995) suggested that the mechanism of the THC oscillation in the same coupled model run is essentially the 2D meridional mechanism described above. In this scenario, the gyre effects are not necessarily an essential part of the oscillation, but are rather a side product of the meridional oscillation. The present analysis also examines the possible existence of a gyre effect in our model results; Fig. 3 shows three snapshots of horizontal sections of the salinity, temperature, and velocity vectors in the North Atlantic during different stages of the oscillation at depth 483 m. While there are clearly some signals in the horizontal distribution of temperature and salinity, they do not seem to be very close to those seen in the coupled model run of Delworth et al. (1993). In particular, the velocity field signal we see does not resemble a clear gyre signal as seen in the coupled model run. Note that we do see the same phase relationship between temperature and salinity as seen in the coupled model THC oscillation. This seems to indicate that the THC oscillation mechanism, at least in the present model, does not rely on the gyre effect analyzed by Delworth et al. (1993), but is essentially a 2D meridional mechanism as in Griffies and Tziperman (1995). The signal we see in our experiments in the horizontal distribution of temperature, salinity, and circulation seems to be only a by-product

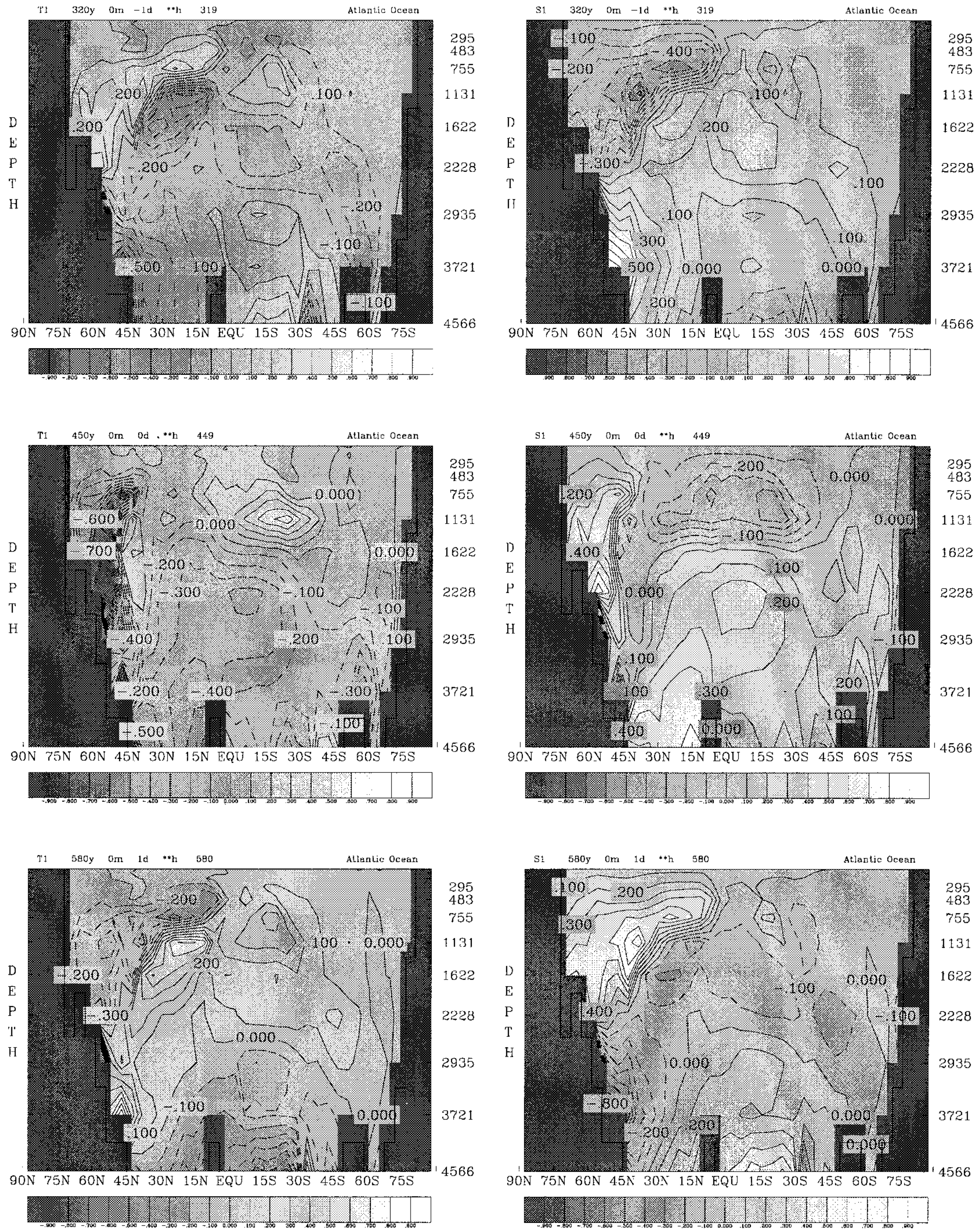

FIG. 2. Three zonally averaged snapshots of the adjoint temperature (left) and salinity (right) during different stages of the thermohaline oscillation, at years 320 (0° phase of the oscillation: upper), 450 (90° phase: middle) and 580 (180° phase: lower). Note that moving up the panels corresponds to forward movement in time of the oscillations.

of the 2D meridional THC oscillation mechanism in our runs.

We note that the timescale of the oscillation in this model (hundreds of years) is significantly longer than the timescale of advection of salinity perturbations around the horizontal gyre circulation (tens of years). The timescale of the oscillation mechanism in the box model of Griffies and Tziperman (1995) is basically set by the ventilation time of the effective North Atlantic sinking area, which could vary from tens to hundreds of years, depending on the specific water mass formation behavior in each model run. We do not understand the source of the longer timescale found here as compared to the 50-yr timescale found in the coupled model run

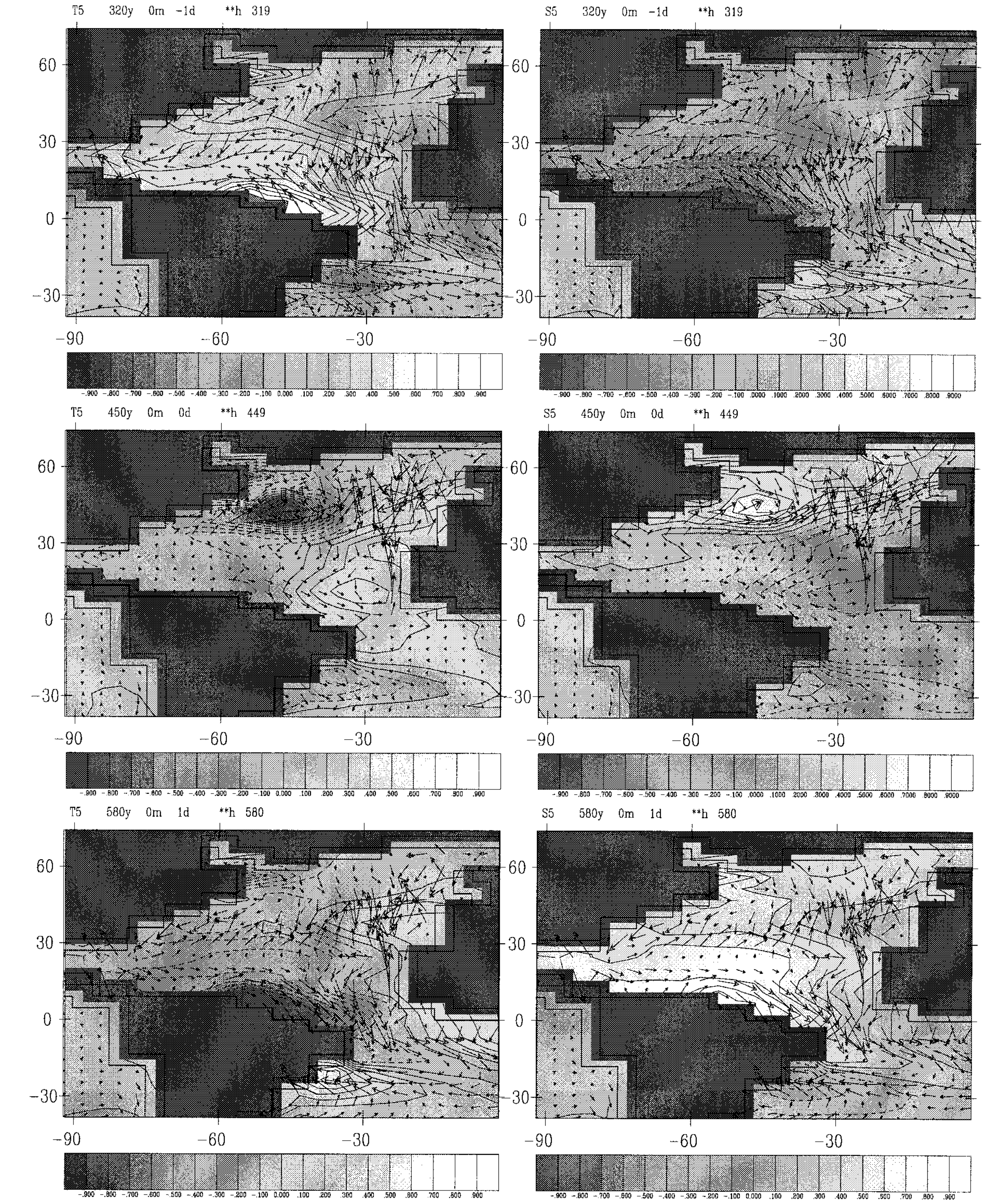

FIG. 3. The adjoint variables of the salinity (right), temperature (left) and currents (superimposed on temperature and salinity) at a depth of 483 m, during the same three stages of the THC oscillation as in Fig. 2.

of Delworth et al. (1993), which used roughly the same resolution ocean model as we do here. It is possible that the different timescale has to do with the missing gyre effect in the present study. It is also possible that our simple treatment of the atmospheric feedbacks (mixed boundary conditions) affects the timescale and that a better atmospheric feedback (energy balance or a statistical atmosphere model) would result in a shorter timescale closer to that seen in the coupled model. Admittedly, the significant difference in timescale between the oscillatory mode found here and the timescale of the oscillation in the coupled model of Delworth et al. (1993) raises some doubts regarding the direct applicability of the presently found oscillatory THC mode to the specific coupled model results of Delworth et al. (1993). We do feel, however, that the demonstrated existence of a damped oscillatory THC mode is potentially relevant to decadal climate variability in spite of these caveats.

b. Stability as function of the basic model state

Previous studies using various ocean-only and coupled ocean–atmosphere models have shown that the stability of the THC (Weaver et al. 1991) and, more specifically, the existence of THC oscillations and their exponential decay (or growth) time, is a function of the basic steady state obtained under restoring conditions (Tziperman et al. 1994; Tziperman 1997, 2000). More specifically, a stronger salinity restoring coefficient (i.e., a shorter restoring time) used when calculating the steady-state solution results in a larger meridional salinity gradient in the North Atlantic Ocean, stronger $E - P$ fluxes and weaker meridional THC circulation. These all lead to a more unstable final solution after switching from restoring to mixed boundary conditions, based on the simple linear advective instability mechanism discussed in Walin (1985) and Marotzke et al. (1988). Thus, the exponential decay time under mixed boundary conditions increases with increasing strength of the salinity restoring coefficient in the forward run under restoring conditions. For a sufficiently strong salinity restoring coefficient, the decay time may become negative, leading to unstable behavior.

Let us first clarify what we mean by stable and unstable steady states of the THC. We consider a given steady state of the North Atlantic overturning thermohaline circulation to be stable if, when used to initialize the model run under mixed boundary conditions, it does not deviate from this initial state. On the other hand, a steady state is considered *unstable* if, when used to initialize the run under mixed boundary conditions, it results in the model experiencing a sizeable drift of the THC away from the initial state. A THC instability may thus result in an increase as well as a decrease, collapse, or in strong oscillations of the THC (Toggweiler et al. 1996). In each of these cases the model may eventually settle on a new stable state, different from the initial unstable state.

We have estimated an exponential decay time for time series of the adjoint temperature variable from four sensitivity experiments of the adjoint model under mixed boundary conditions. The experiments started from steady states obtained using restoring conditions for both the temperature and salinity, with four different restoring times for the surface salinity field ($\gamma_s^{-1} = 12$, 20, 30, 120 days). As a result, the steady-state North Atlantic overturning is different in each of these experiments, so is their stability behavior. The steady-state overturning strengths for the experiments ran with salinity restoring timescales of $\gamma_s^{-1} = 12$, 20, 30, 120 days are 15.5, 16.2, 16.8, and 19.6 Sv(Sv $\equiv 10^6$ m^3 s^{-1}), respectively. The corresponding exponential decay times calculated from the adjoint runs [i.e., the times τ in the exponential factor $\exp(t/\tau)$ required to cancel the exponential decay or growth] are found to be -9, 130, 130, 60 years, respectively. That is, the salinity restoring timescale of $\gamma_s^{-1} = 12$ leads to a linearly unstable steady state based on the adjoint analysis, while $\gamma_s^{-1} = 20$, 30, 120 lead to a stable steady state, with the experiments using $\gamma_s^{-1} = 120$ days being more damped than those using $\gamma_s^{-1} = 20$, 30 days. Generally, these results are clearly consistent with our a priori expectations: strong salinity forcing (i.e., strong salinity-restoring coefficients or short salinity restoring times) results in weaker overturning and less stable behavior under mixed boundary conditions.

Our adjoint model analysis reflects the linear stability of the model under mixed boundary conditions for each steady state. But, we can go even further with the stability analysis. By comparing this linearized stability analysis to the actual stability behavior of the fully nonlinear model itself, we can obtain some insight into the thermohaline instability mechanism in the fully nonlinear model. Figure 4 shows time series of the maximum North Atlantic meridional overturning streamfunction under mixed conditions, from the fully nonlinear forward primitive equation model, starting from the same four steady states mentioned above obtained using four different salinity restoring times. The experiment starting with the long salinity restoring timescale ($\gamma_s^{-1} =$ 120 days) is clearly stable, and remains at its initial state. All of the other three runs, starting with steady states obtained with shorter salinity restoring timescales, do not remain at their initial states, indicating that these initial states are unstable.

While both the linear (adjoint) analysis and the fully nonlinear model experiments show that a steady state with a weak overturning is unstable ($\gamma_s^{-1} = 12$ day), and one with a strong overturning is stable ($\gamma_s^{-1} = 120$ day), there are differences between the two analyses for the runs with intermediate amplitudes of the salinity restoring times ($\gamma_s^{-1} = 20$, 30 day) and thus intermediate values of the steady-state North Atlantic overturning. These runs are stable according to the linearized (adjoint) analysis (positive decay times), yet unstable in the (fully nonlinear) forward model run shown in Fig. 4 (dash and dot-dash lines) where the overturning under mixed boundary conditions clearly deviates from the steady states calculated under restoring conditions. There are at least two possible explanations for this difference in stability behavior of the nonlinear model and the linearized adjoint analysis: first, that the instability mechanism of these two intermediate experiments

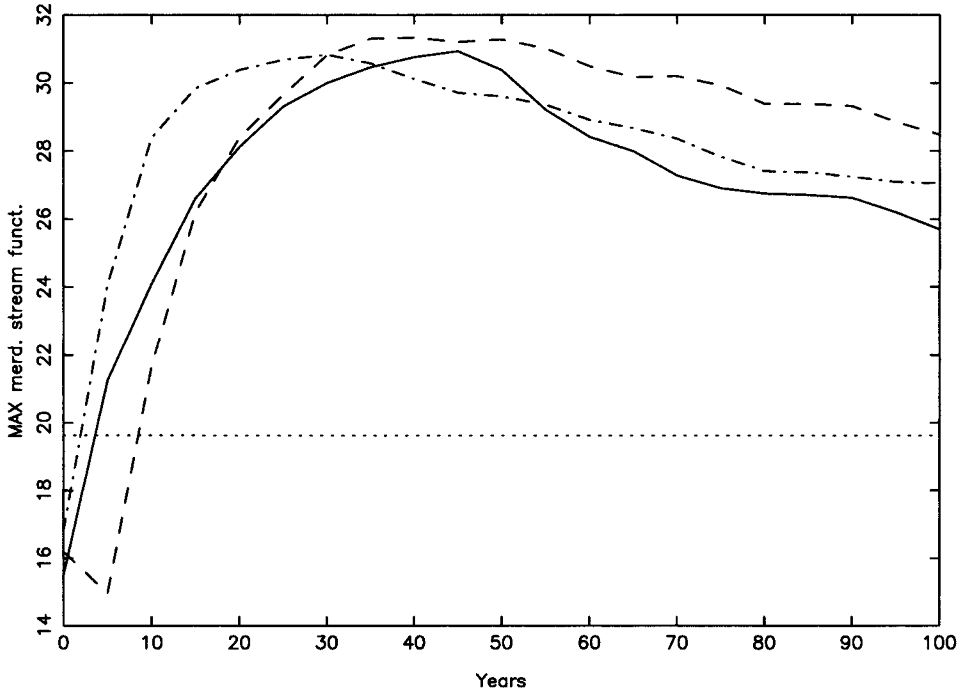

FIG. 4. Overturning index as function of time from forward model runs under mixed boundary conditions (solid, dash, dot-dash, dot for $\gamma_s^{-1} = 12, 20, 30, 120$ days, correspondingly). The runs are started from the same steady states whose analysis is examined using the adjoint runs discussed in the text.

is nonlinear [e.g., involving nonlinear convection feedbacks such as in Lenderink and Haarsma (1994)], and second, that while these two intermediate experiments are linearly stable, they allow for temporary growth of small perturbations due to the non-normal (Farrell and Ioannou 1996) nature of the problem. Such an initial temporary growth may be followed by the model nonlinearities taking over and carrying the solution further away from the initial steady state. The final outcome is a deviation from the initial state toward a different solution, even though the initial steady state is stable in the usual sense. The verification of this hypothesis calls for an optimal-mode analysis of this stability problem (Farrell and Ioannou 1996), which can, in principle (although not trivially), be carried out using adjoint-based tools.

It is worthwhile to mention at this stage that, while we have concentrated on the effects of the salinity restoring time on the model stability, it is well known now that a too short restoring time for the sea surface temperature in this formulation, as well as a restoring time that is not scale selective, also has a strong effect on the stability of models under mixed boundary conditions (Zhang et al. 1993; Mikolajewicz and Maier-Reimer 1994; Power and Kleeman 1994; Rahmstorf and Willebrand 1995). Still, it seems that in spite of these possible artifacts due to the simplified thermal feedback used in mixed boundary conditions, the qualitative dependence of the stability on the salinity forcing carries over from a simple mixed boundary condition analysis to full coupled ocean–atmosphere models (Tziperman 2000). There is therefore reason to believe that the lessons learned in this work regarding the existence of an oscillatory mode and stability will carry over to ocean models coupled to fuller atmospheric representations, and perhaps even to the climate system itself.

4. Conclusions

We have used an adjoint of a coarse-resolution global primitive equation ocean model to identify an oscillatory mode of the North Atlantic thermohaline circulation, and to analyze the linearized versus nonlinear stability behavior of the thermohaline circulation under mixed boundary conditions. As explained above, this approach is complementary to using more realistic atmospheric representations and allows us to clearly analyze the relevant physical mechanisms. We found that the ocean general circulation model has a damped oscillatory THC mode basically identical to that proposed by Tziperman et al. (1994) and Griffies and Tziperman (1995) using simple meridional box models. While we see signals in the horizontal circulation that are related to the THC oscillation, the oscillation mechanism seems to be essentially two dimensional (meridional–depth) and does not seem to require an active participation of the horizontal gyre circulation to produce the THC oscillation. This is in contradiction to the analysis of the coupled model experiments by Delworth et al. (1993), which seemed to indicate a necessary participation of the horizontal gyre circulation in the THC oscillation mechanism. The long timescale of the oscillation in the present model is such that it may also rule out a gyre-related mechanism as the source of the oscillation. We conclude from this that the variations in the amplitude of the horizontal temperature, salinity, and circulation in the present model are perhaps a side effect of the main meridional oscillation mechanism, rather than an integral part of the THC oscillation mechanism.

In addition to the discussion of the oscillatory THC mode, the adjoint analysis enabled us to study the linearized stability of the THC under mixed boundary conditions. We find, in accordance with previous studies (Weaver et al. 1991; Tziperman et al. 1994; Tziperman 1997, 2000) that a steady-state solution obtained under strong salinity forcing (and hence with a weaker North Atlantic overturning), is unstable, while a steady-state solution with stronger THC is stable. The instability mechanism is basically the linearized advective mechanism of Walin (1985) (see also Marotzke et al. 1988). While the dependence of the stability on the salinity forcing used to obtain the steady-state solution is basically that expected from simple linearized considerations, we did find that in a certain intermediate parameter regime the full model may be unstable while the linearized analysis indicated that it should be stable. We have proposed that this may be because either the instability mechanism at this intermediate regime is nonlinear or, while the model is linearly stable at this regime, it allows for temporary growth of small perturbations due to the non-normal nature of the problem (Farrell and Ioannou 1996). This calls for an optimal-mode analysis of this stability problem, which we hope to pursue in a future work.

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