Interdecadal Variability Driven by Mismatch between Surface Flux Forcing and Oceanic Freshwater/Heat Transport

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ABSTRACT

The author describes a series of mechanistic experiments showing the generation of interdecadal variability in the Bryan–Cox ocean general circulation model driven by a constant two-dimensional freshwater or heat flux field alone. The model ocean has a flat bottom and idealized model geometry of size comparable to the North and South Atlantic.

A set of experiments examines the variability of saline circulation. Four spins are carried out: (a) under a restoring boundary condition on salinity alone (run R3); (b) under a restoring boundary condition on salinity and a wind forcing (run R3W); (c) under restoring boundary conditions on both temperature and salinity without wind forcing (run R3T); and (d) under the same forcings as (c) but in the presence of the wind forcing (run R3TW). Four fields of surface freshwater flux are diagnosed from the steady state of each spinup. Four cases are then run, each under a diagnosed freshwater flux field. Except in the case under the field diagnosed from run R3, internal variability takes place. The internal variability is induced by a mismatch between the freshwater transport implied by the surface freshwater flux forcing and the oceanic freshwater transport. Parallel experiments are carried out to study the internal variability driven by a constant two-dimensional heat flux field alone. Internal oscillations again develop as a result of a mismatch between the atmospheric and oceanic heat transport. In a coupled atmosphere–ocean system the atmospheric freshwater (or heat) transport needs not always match the oceanic freshwater (or heat) transport. This may play a role in the generation of the variability in the coupled system.

The mismatch mechanism can operate in a system forced by a Haney restoration for surface temperature and a flux condition of salinity (mixed boundary conditions). A positive feedback mechanism associated with mixed boundary conditions misrepresents the role of the thermal and saline forcings. This can lead to destruction of the thermally driven circulation feature and yields solutions similar to those without thermal forcing, that is, with a persistent oscillation.

1. Introduction

Recently, there has been a surge of research interest in the internal variability of the ocean thermohaline circulation (e.g., Welander 1986; Broecker et al. 1990; Marotzke 1990; Mikolajewicz and Maier-Reimer 1990; Wright and Stocker 1991; Weaver and Sarachik 1991a,b; Weaver et al. 1993; Lenderink and Haarsma 1993; Winton and Sarachik 1993). The research is especially important for the detection of greenhouse signals. Over the past few years, studies of the role played by freshwater flux in thermohaline circulation have been carried out in a number of ocean models under the so-called mixed boundary conditions [i.e., the use of a restoring boundary condition on surface temperature and a flux boundary condition on surface salinity, Bryan (1986)] and in the presence of wind stress forcing (e.g., Marotzke 1990; Weaver and Sarachik 1991a,b; Weaver et al. 1993; Winton and Sarachik 1993). Decadal variabilities forced by freshwater flux have been found in these models, and the role of freshwater flux in the thermohaline circulation is examined in the presence of wind and thermal forcings. However, the effect of freshwater flux in isolation remains largely unexplored.

Huang and Chou (1994) studied the unique dynamic character of the freshwater flux in an ocean model forced by specified zonally uniform freshwater flux alone. They found solutions ranging from a steady state with no oscillation, a state with regular oscillation, to one with chaotic variabilities, depending on the model parameters and the profile of the zonally uniform freshwater flux. Although the model sensitivities are examined by Huang and Chou (1994) in a substantial number of experiments that span a broad range of parameter space and different freshwater profiles, many aspects still need to be further investigated. For example, it is not clear to what extent their result depends upon the zonal uniformity of the forcing and whether the oscillatory behavior of their solutions exists in a
model forced by a two-dimensional freshwater flux. Note that in the sensitivity studies of Huang and Chou (1994), freshwater flux is provided as a natural boundary condition for salinity balance. The natural boundary condition modifies the rigid-lid approximation (that is, the vertical velocity is set to zero at the ocean surface) by setting the vertical velocity at the surface to be the rate of the evaporation minus precipitation ($E - P$). However, in the majority of published work, forcing for salinity is generally provided through the virtual salt flux diagnosed from a restorative spinup. While there are advantages in using the natural boundary condition, our study uses the virtual salt flux as forcing for surface salinity. This enables an examination of the role of the freshwater flux component of mixed boundary conditions.

Kushnir (1994) presented evidence showing oceanic variabilities in the North Atlantic with a 30–40-year timescale. In a similar study, Deser and Blackmon (1993) proposed that the general warming trend in the North Atlantic sea surface temperature (SST) during the 1920s–1930s and the cooling in the 1960s are associated with changes in the Gulf Stream system. These new results emerge at a time when results from the Geophysical Fluid Dynamics Laboratory (GFDL) coupled ocean–atmosphere model show the existence of similar oscillations in the thermohaline circulation in the North Atlantic with a comparable timescale of 40–50 years (Delworth et al. 1993). Deser and Blackmon (1993) also showed a variability of about 10 years. At this stage, it is not clear what role the freshwater and heat flux forcings play in generating these variabilities.

The central issue to address in this paper is "what is the response of ocean general circulation models (OGCMs) to mismatches between surface freshwater flux forcing and oceanic freshwater transport?" We propose that the mismatch will generate oscillations. When a freshwater flux is diagnosed from a restoring spinup, there is an implied atmospheric freshwater transport. This transport is determined entirely by the ocean, and the imposed, time-independent restoring surface thermohaline field used in the spinup, in particular, the influence of the model dynamics and the mixing parameterizations in driving the model surface field away from the imposed restoring field. Consider a freshwater flux field diagnosed from the steady state of an ocean driven by restoring the model sea surface salinity (SSS) to an imposed salinity field alone. This diagnosed flux field is fully compatible with the model circulation dynamics, in particular, with the modeled meridional overturning circulation. Using this diagnosed field alone to drive the same model will produce a steady solution the same as the steady solution of the spinup with no oscillation. In these two steady solutions, the saline circulation is driven by the pressure gradient created by vertical salt mixing, and the freshwater transport by the ocean is fully in balance with the implied atmospheric freshwater transport. Now consider an ocean driven by another freshwater flux field alone that is diagnosed from other spinups, for example, in the presence of a wind forcing, or in the presence of a restorative forcing for SST, or in the presence of both. In this situation the atmospheric and oceanic freshwater transports can no longer be in balance. This is because the oceanic circulation that takes part in maintaining the freshwater balance with the atmosphere in the spinups is no longer dynamically supported in the system driven by the freshwater flux alone. In this situation internal variability may take place. A comparison of the above two situations makes the generation mechanism clear. There is no dynamic reason why in the coupled atmosphere–ocean system, the oceanic freshwater transport is always in balance with the atmospheric freshwater transport. For example, the $E - P$ field is not fully determined by the ocean. Despite the role by SST in the evaporation, the $E - P$ field is also influenced by other nonlinear atmospheric processes. Indeed, it is not clear whether the balance between the oceanic and atmospheric freshwater transport can be maintained in the face of other factors. Hence, the mechanism identified here may be partially responsible for the oscillatory nature of the coupled system.

Note that the mismatch between surface forcing and oceanic salt (or freshwater) transport has several inferences. First, given an ocean system, which is defined by its prescribed internal physics and external forcings, the salt transport that it can maintain is not entirely determined by the forcing alone. Instead it is also influenced by the internal physics. However, if a flux forcing is used, the ocean is actually constrained by a constant imposed salt (or freshwater) transport, which may not be able to be maintained by the internal physics. Hence, the mismatch can be interpreted as one between the imposed oceanic salt (or freshwater) transport and the internal physics. Second, since the maintenance of salt transport at a given latitude by the ocean is carried out locally by advections of salt, locally the mismatch is between the imposed divergence/convergence of atmospheric freshwater flux and the oceanic salt divergence/convergence rate.

It will be shown that the mismatch mechanism is present in ocean models under mixed boundary conditions. Under mixed boundary conditions, the restoring condition on temperature promotes a positive feedback, which misrepresents the role of a freshwater flux forcing. This leads to a change in thermal circulation, which then leads to a mismatch between the implied atmospheric freshwater transport and the ocean circulation, and hence to internal variability.

Similar consideration can be given to a system driven by heat flux alone. Forcing the ocean with a diagnosed surface heat flux from a spinup implies a fixed atmospheric transport needed to balance the atmospheric radiation budget. Diagnosed from a spinup, this implied atmospheric transport is entirely deter-
mined by the ocean. In reality, the atmospheric heat transport required to balance the radiation budget is not determined by the ocean alone; there are other factors that influence the atmospheric heat transport, such as the temperature contrast between the land and the ocean, the mountain ranges that run north—south across the zonal flow, and so on. Hence, it is not clear whether the balance between the oceanic and atmospheric heat transport can be maintained in the face of these other factors.

The structure for this paper is as follows. In section 2 we briefly describe the ocean model, the surface forcing, and the model runs. In section 3 we examine oscillations under constant freshwater flux forcing alone. The role of freshwater flux in mixed boundary conditions is examined in section 4. In section 5 we study the oscillations under constant heat flux forcing. Section 6 provides the conclusions.

2. The ocean model and model runs

a. The model

This study employs the Pacanowski et al. (1991) version of the Bryan–Cox OGC that is based on Bryan (1969). The model domain is a flat-bottomed, two hemispheric basin, 60° in width, and extending from 68°S to 68°N. The Southern Hemisphere includes a modeled Antarctic Circumpolar Current (ACC) passage from 48° to 68°S. An ACC with a strength of 200 Sv ($S = 10^6$ m$^3$ s$^{-1}$) is imposed throughout all the experiments, following Marotzke and Willebrand (1991). Note that the imposed ACC is not necessary for generating the oscillation in the paper. We include the ACC for the purpose of obtaining a freshwater flux field from a more realistic spinup.

The horizontal grid spacing is 4° latitude by 4° longitude. The model has 12 levels in the vertical, at depths listed in Table 1. Values assigned to the various model parameters are listed in Table 2. No-slip and insulating boundary conditions are applied at the lateral boundaries. The model uses the Cox (1987) parameterization to compute vertical diffusion and convection implicitly. The enhanced vertical diffusivity in regions of static instability is set at $10^5$ cm$^2$ s$^{-1}$, which is the convective adjustment in the model. For simplicity, bottom friction is ignored.

b. The model runs

Three sets of model runs are carried out. The first set of experiments is devoted to studying the saline circulation and its variability driven by a freshwater flux alone. Instead of forcing the model with a specified freshwater flux, the forcing fields are diagnosed from spinups using Haney (1971) restoring on salinity—temperature. The diagnosed flux field is two-dimensional and is somewhat more realistic than the specified zonally uniform field in Huang and Chou (1994).

The second set of experiments examines the role of freshwater flux under mixed boundary conditions. The thermal forcing in the mixed boundary conditions uses the Haney restoring scheme in these experiments.

The third set of experiments examines the circulation driven by a two-dimensional heat flux forcing alone. The heat flux field is again diagnosed from spinups under Haney restoring forcing for temperature/salinity.

3. Saline circulation under constant freshwater flux

a. Spinups for diagnosing the freshwater (salinity) flux

The most common way to force OGC models is to apply Haney restoring surface boundary conditions to both SST and SSS. A relaxation condition on salinity in effect provides a salinity flux, or a freshwater flux, which is strongly dependent on the SSS. In reality, the freshwater flux from evaporation and precipitation is almost independent of the underlying salinity. Thus, in reality the upper boundary condition on salinity is flux forcing rather than relaxation (e.g., Wclander 1986; Weaver and Sarachik 1991a,b; Moore and Reason 1993; Rahmstorf and Willebrand 1993). Four spinups are performed to provide four fields of diagnosed freshwater (salt) flux, as listed in Table 3. All the spinups in this section use the Haney restoring scheme on SST/SSS. In run $R_S$ model topmost level salinity is relaxed to a zonally uniform salinity $S$ alone given by the profile shown in Fig. 1a. The e-folding restoring timescale is 20 days. We chose this timescale since our upper-

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### Table 1. Distribution of vertical levels.

<table>
<thead>
<tr>
<th>Level</th>
<th>Thickness (m)</th>
<th>Depth of ($T$, $S$) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>25.0</td>
<td>12.5</td>
</tr>
<tr>
<td>2</td>
<td>25.0</td>
<td>37.5</td>
</tr>
<tr>
<td>3</td>
<td>40.0</td>
<td>70.0</td>
</tr>
<tr>
<td>4</td>
<td>70.0</td>
<td>125.0</td>
</tr>
<tr>
<td>5</td>
<td>110.0</td>
<td>215.0</td>
</tr>
<tr>
<td>6</td>
<td>200.0</td>
<td>370.0</td>
</tr>
<tr>
<td>7</td>
<td>330.0</td>
<td>635.0</td>
</tr>
<tr>
<td>8</td>
<td>450.0</td>
<td>1025.0</td>
</tr>
<tr>
<td>9</td>
<td>650.0</td>
<td>1575.0</td>
</tr>
<tr>
<td>10</td>
<td>900.0</td>
<td>2350.0</td>
</tr>
<tr>
<td>11</td>
<td>900.0</td>
<td>3250.0</td>
</tr>
<tr>
<td>12</td>
<td>900.0</td>
<td>4150.0</td>
</tr>
</tbody>
</table>

### Table 2. Standard values of model coefficients.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value (m$^2$ s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal diffusivity</td>
<td>$\alpha_{D}$</td>
<td>$1 \times 10^3$</td>
</tr>
<tr>
<td>Horizontal viscosity</td>
<td>$\alpha_{m}$</td>
<td>$3 \times 10^3$</td>
</tr>
<tr>
<td>Vertical diffusivity</td>
<td>$\alpha_{V}$</td>
<td>$1 \times 10^{-4}$</td>
</tr>
<tr>
<td>Vertical viscosity</td>
<td>$\alpha_{w}$</td>
<td>$1 \times 10^{-4}$</td>
</tr>
</tbody>
</table>
TABLE 3. Model runs for studying the salt oscillation and the role of freshwater flux forcing under mixed boundary conditions. See text for details of each experiment.

<table>
<thead>
<tr>
<th>Run</th>
<th>Initial condition</th>
<th>Restoring on</th>
<th>Flux on SSS</th>
<th>Wind</th>
<th>Period (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_{w}$</td>
<td>rest</td>
<td>SSS</td>
<td>no</td>
<td>steady</td>
<td></td>
</tr>
<tr>
<td>$R_{w}$</td>
<td>rest</td>
<td>SSS</td>
<td>yes</td>
<td>steady</td>
<td></td>
</tr>
<tr>
<td>$R_{w}$</td>
<td>rest</td>
<td>SSS and SST</td>
<td>no</td>
<td>steady</td>
<td></td>
</tr>
<tr>
<td>$R_{w}$</td>
<td>rest</td>
<td>SSS and SST</td>
<td>yes</td>
<td>steady</td>
<td></td>
</tr>
<tr>
<td>$F_{w}$</td>
<td>rest</td>
<td>yes</td>
<td>no</td>
<td>steady</td>
<td></td>
</tr>
<tr>
<td>$F_{w}$</td>
<td>rest</td>
<td>yes</td>
<td>no</td>
<td>106.6</td>
<td></td>
</tr>
<tr>
<td>$F_{w}$</td>
<td>rest</td>
<td>yes</td>
<td>no</td>
<td>48.1</td>
<td></td>
</tr>
<tr>
<td>$F_{w}$</td>
<td>rest</td>
<td>yes</td>
<td>no</td>
<td>43.8</td>
<td></td>
</tr>
<tr>
<td>$F_{w}$</td>
<td>rest</td>
<td>yes</td>
<td>yes</td>
<td>19.9</td>
<td></td>
</tr>
<tr>
<td>$M_{w}$</td>
<td>$R_{w}$</td>
<td>SST</td>
<td>yes</td>
<td>no</td>
<td>71.8</td>
</tr>
<tr>
<td>$M_{w}$</td>
<td>$R_{w}$</td>
<td>SST</td>
<td>yes</td>
<td>yes</td>
<td>steady</td>
</tr>
</tbody>
</table>

Most level has a thickness of 25 m. (A thickness of 50 m would then correspond to a restoring timescale of 40 days.) In this run the temperature is kept fixed at a uniform value of 7°C. In run $R_{w}$, in addition to the restorative forcing on salinity the model is subject to a zonally uniform wind stress of Bryan (1987), symmetric about the equator. In run $R_{w}$, in addition to the salinity forcing in run $R_{w}$, the topmost level temperature is restored to a zonally uniform temperature $T_{w}$, also shown in Fig. 1b, with the same relaxation timescale as for salinity. In run $R_{w}$, the model is subject to the wind forcing and restoring forcing on both temperature and salinity as in run $R_{w}$.

The spinup model results themselves are interesting and are summarized in Figs. 2–4. Figures 2 and 3 show the meridional overturning streamfunction and the diagnosed implied freshwater flux from the steady state of the four spinups. Among these four fields of freshwater flux, the one from run $R_{w}$ is more realistic, since it is derived under all forcing components. The zonal means of the freshwater flux fields are shown in Fig. 4. The freshwater flux is diagnosed by averaging the monthly output of the last 50 years of the integration so that the global integration of the field is essentially zero. Looking at these figures, we see that the meridional overturning in an ocean driven by salinity alone is characterized by a two-cell structure at each hemisphere: deep water forming at midlatitudes and returning to the surface at tropical and high-latitude regions. The hemispheric asymmetry in these fields is apparently due to the asymmetry in model geometry, that is, the inclusion of the ACC passage in the Southern Hemisphere, as found in previous studies (e.g., Marotzke and Willebrand 1991; Hughes and Weaver 1993).

A comparison between runs $R_{w}$ and $R_{w}$ shows that the wind does not change the overall pattern of the overturning much, suggesting the predominance by the salinity forcing and the small effect by wind. The insignificant role by wind will be seen later in other experiments. However, the freshwater input to the ocean in the tropical and high-latitude regions increases slightly.

A comparison between run $R_{w}$ and run $R_{w}$ (or runs $R_{w}$ and $R_{w}$) indicates the predominance of thermal circulation in the modeled thermohaline circulation. In the presence of temperature forcing (runs $R_{w}$ and $R_{w}$) the characteristics of the saline circulation completely disappear, as can be seen from Figs. 2c and 2d.

Fig. 1. (a) Surface relaxation temperature and (b) surface relaxation salinity.
in both runs a predominant northern sinking cell appears and is accompanied by a southern sinking cell and a northern intrusion cell from the south. In the presence of thermal forcing, the hemispheric asymmetries in the freshwater flux are greatly enhanced, with net precipitation in the Northern Hemisphere and net evaporation in the Southern Hemisphere. The modeled circulation is dominated by a Gulf Stream-like “conveyor belt” (Broecker 1987; Cai and Greatbatch 1995). Deep water is mainly formed in the northeast corner, and its outflow moves along the northern and western boundaries. The outflow passes the equator to the Southern Hemisphere, and this appears to be compensated by the modeled Benguela Current-like flow at subsurface depth, consistent with the result presented by Gordon et al. (1992) and Cai and Greatbatch (1995). This cross-equatorial flow is dynamically supported by a cross-hemispheric pressure gradient.

A comparison between run $R_{SW}$ and run $R_{STW}$ again shows the small role of the wind forcing: in the overturning circulation, the wind forcing leads to several wind-driven cells confined to the upper 500 m (Figs. 2c and 2d); wind-induced equatorial upwellings bring up cold water leading to an increased heat intake by the ocean there, which is then transported poleward by the wind-induced currents (see section 5); this small wind-induced increase in heat transport is accompanied by a small wind-induced enhancement of the implied hydrological cycle, as can be seen in Figs. 3c, 3d, and 4. The implied freshwater fluxes from both runs $R_{ST}$ and $R_{STW}$ show similar two-dimensional patterns with net precipitation in the high latitudes and net evaporation in low latitudes. The insignificance of wind forcing seen here is also consistent with the recent result in a global OGCM reported by Cai (1994).

The above four spinups are carried out to provide the freshwater flux fields that will be used as the only forcing to drive the model.

b. The generation of internal variability under constant freshwater flux

Before we proceed further, we report that upon switching from restoring to the diagnosed freshwater flux forcing in runs $R_S$ and $R_{SW}$, no appreciable change from the end state of the spinup takes place. In either case, the implied atmospheric freshwater transport is in balance with the oceanic freshwater transport. Note that the switch here from a restoring to a flux boundary condition is different from the switch from restoring boundary conditions on both temperature and salinity to mixed boundary conditions; the latter switch generates a mismatch through a positive feedback due to the restoring boundary condition on temperature. This will be discussed in section 4. Similarly, in runs $R_{ST}$ and $R_{STW}$, switching from restoring forcings to flux forcings for both salinity and temperature produces virtually no change, in particular, no oscillations in the model solution. This is not surprising given that the same result has been found previously in studies of thermohaline circulation (e.g., Zhang et al. 1993; Powers and Kleeman 1993; Cai and Godfrey 1995). In runs $R_{ST}$ and $R_{STW}$, the ocean is in balance with the atmosphere in terms of the implied atmospheric heat and freshwater transports. As will be clear in section 4, a flux boundary condition on temperature promotes a negative feedback between SST and overturning, which prevents the overturning circulation from the idiosyncratic behavior under mixed boundary conditions.
Fig. 3. Freshwater flux fields diagnosed from spinup. Panels (a), (b), (c), and (d) are for runs $R_S$, $R_{SW}$, $R_{ST}$, and $R_{STW}$, respectively.
Note that since the ocean model used here is formulated using the rigid-lid approximation, namely the vertical velocity at the ocean surface is set to zero [as opposed to the natural boundary conditions in Huang and Chou (1994)], it is not possible to force the ocean model directly with a freshwater flux. We actually use a salt flux to force the model. In the course of diagnosing the implied freshwater flux, we diagnose the salt flux, although Fig. 4 shows the implied freshwater fluxes. Four runs are carried out, each forced by one of the four diagnosed implied freshwater flux fields (that is, the equivalent salt flux fields) alone. These four experiments are termed as runs $F_S$, $F_{SW}$, $F_{ST}$, and $F_{STW}$ and are forced by the implied freshwater flux diagnosed from runs $R_S$, $R_{SW}$, $R_{ST}$, and $R_{STW}$, respectively (see Table 3). These runs start from an ocean at rest with the identical ocean model as in the spinups, and with a uniform salinity of 34 ppt. No lower-level acceleration is used for these runs. Temperature is kept constant and uniform at 7°C. Each experiment is run for 1500 years. The same size of the ACC is imposed as in the spinup experiments.

Run $F_S$ settles to a solution with no oscillation, similar to that in run $R_S$, as expected. This is because the surface forcing matches the model dynamics. However, run $F_{SW}$, upon initial adjustment, settles to a steady circulation with regular oscillations for a period of 106.6 years. Figure 5 shows the time series of the overturning over the period of the last 500 years. Figure 6 shows the mean meridional circulation for each experiment over five steady cycles. Figure 7 shows the zonally averaged salinity. Note that the location (32°N, 2350 m) where the overturning time series is sampled does not coincide with the place where the overturning under the flux forcing is at a maximum, nor with the place where the amplitude of oscillation is at a maximum. The generation of the oscillation is apparently due to the incompatibility between the surface forcing and the modeled circulation. Recall that at the steady stage of run $R_{SW}$, to maintain the atmospheric freshwater transport implied in the diagnosed freshwater flux, the ocean circulation that supports the freshwater balance between the atmosphere and the ocean is also dynamically driven by the wind forcing. In run $F_{SW}$, the elimination of wind forcing leads to a mismatch between the implied atmospheric freshwater transport and the ocean circulation.
Fig. 5. Time series of zonally integrated meridional overturning from runs forced by diagnosed freshwater flux field alone. Shown are for the period from year 1000 to year 1500. The time series are sampled at (32°S, 2350 m), (32°N, 2350 m), (32°S, 1575 m), and (32°S, 1575 m) for runs F_s, F_sw, F_st, and F_sw, respectively.

We have also carried out another experiment, FW_sw, forced by the same freshwater flux field as in run F_w but with the wind forcing as well. The steady state is very much the same as that in run R_sw and there is no oscillation. That the oscillation is generated by the mismatch due to the elimination of the wind forcing is further confirmed in the following experiments, in which run FW_sw is rerun with different decreasing values of the wind stress. We find that when the wind stress is reduced to 50% there is no oscillation; when the wind stress is reduced to 40%, oscillation with a period of 68.1 years takes place; and when the wind is reduced to 25%, oscillation with a period of 87.2 years occurs. Apparently, at a critical value of the wind stress, a bifurcation occurs.

The oscillation in run F_sw and the no-oscillation solution in run F_s can be seen as due to the difference in the freshwater flux forcing. Through a comparison between runs R_s and R_sw, we have shown that the wind only slightly affects the overturning circulation and the
freshwater flux field. Yet it is this small difference due to the wind effect during spinup that is responsible for the oscillation in run $F_{SW}$. This high sensitivity is confirmed by the study of Huang and Chou (1994) in a model forced by a zonally uniform surface freshwater flux alone and in models under mixed boundary conditions by Weaver and Sarachik (1991a,b) and Weaver et al. (1993).

Runs $F_{ST}$ and $F_{STW}$ also produce oscillations with periods of 48.1 and 43.8 years, as can be seen in Figs. 5c and 5d. In run $F_{ST}$ the mismatch is provided by the elimination of the thermal forcing. In run $F_{STW}$ the mismatch is provided by the elimination of both the thermal forcing and the wind forcing. In these two runs, the time evolution of each regular cycle is much more complex: each major cycle contains superimposed subcycles. The complexity appears to be associated with the nature of the mismatch between the forcing and the ocean circulation. In the sensitivity study of Huang and Chou (1994), the feature of smaller-amplitude subcycles within a major cycle was also produced. They
found the generation of this feature is sensitive to profiles of freshwater flux.

In runs $F_{ST}$ and $F_{STW}$, the mean meridional circulation (Fig. 6) shows a predominant southern sinking cell, with a maximum located north of the ACC. This is a cross-hemispheric circulation cell. Deep water is formed at about 35°S. Part of the deep water flows across the equator to the Northern Hemisphere and returns to the Southern Hemisphere through surface flows. This feature is not unlike the cross-hemispheric circulation found by Bryan (1986) in a model under mixed boundary conditions. The meridional circulation is also like the hemispheric reversal of that from runs $R_{ST}$ and $R_{STW}$, although the maximum overturning takes place at about 35°S. The salinity in the northern high latitudes freshens as can be seen in the zonally averaged salinity shown in Fig. 7. This freshening is the ocean’s response to the strong net precipitation in the freshwater flux forcing, and to the weak salt transport associated with the collapse of the northern overturning. The strong hemispheric asymmetry in the surface forcing with large net precipitation in the Northern Hemisphere also creates a synoptic cross-hemispheric pressure gradient, necessary for maintaining the overturning cell. Note that this cross-equatorial pressure gradient in runs $F_{ST}$ and $F_{STW}$ is also a qualitative reversal of that in runs $R_{ST}$ and $R_{STW}$. In those two cases the thermal circulation dominates, and there is a cross-equatorial flow from the Southern to the Northern Hemispheres.

c. The oscillation in run $F_{SW}$

We describe the oscillation in run $F_{SW}$ in detail. A comparison between the forcing field of runs $F_{S}$ and $F_{SW}$ is made to further illustrate the oscillation mechanism. The oscillatory features in other runs are similar to those in this run.

Before we proceed, it is important to note that when an OGCM is forced by a constant freshwater flux forcing, the OGCM is constrained at a given latitude by a constant meridional salt transport implied in the flux field. Since salt transport is the product of meridional velocity and salinity, the system maintained this fixed salt transport through distributions of salinity and velocity fields, which in turn are dynamically connected through the pressure gradient. Given an OGCM and its prescribed internal physics, when the constant salt transport represents too strong a constraint for the internal physics to maintain, that is, when a mismatch exists, the maintenance of a constant salt transport will produce anomalies in either salinity or meridional flow.

The implied salt transport in the forcing field for run $F_{SW}$ and the difference between runs $F_{SW}$ and $F_{S}$ ($F_{SW}$ minus $F_{S}$) are shown in Fig. 8. The implied salt transport in the forcing for run $F_{SW}$ is poleward (equatorward) from 27° to the pole (equator). Since we know that the OGCM and the associated internal physics can maintain the salt transport implied in the forcing field for run $F_{S}$, the difference field gives us some indications about the mismatches. We see that under the forcing field for run $F_{SW}$, the implied poleward salt transport from 27° to 46° is smaller than that required by the oceanic internal physics in order to have a nonoscillatory state. The horizontal pattern of the mismatch is shown in Fig. 9a in terms of a difference field between the freshwater flux forcing (runs $F_{SW}$ minus $F_{S}$). Several features emerge. First, the mismatches that coincide with deep-water formation regions are located around 35°N and 35°S, and the zonal integration of the difference field (i.e., Fig. 9a) along these latitudes has a net evaporation (see also Fig. 4). Second, the maximum mismatch near 35°N is larger than that near 35°S. Third, at a given latitude, there is a strong zonal dependency. It will be clear that a mismatch that has a net evaporation over a deep-water formation region (e.g., near 35°N) is more conducive to oscillations than a mismatch with a net evaporation but in a region without deep-water formation (e.g., in the polar regions).

Before we examine the dynamical process involved, it is useful to examine the mean horizontal circulation. The mean state of SSS, and currents at surface and at 1450 m, are shown in Figs. 9b and 9c. From Figs. 3b, 6b, 9b, and 9c, we see that surface water with relatively low salinity flows to the deep-water formation region, and becomes salinized due to the net evaporation. It then sinks and flows out of the sinking region at depth in opposite directions to the surface flows. The mean salt transport at any latitude calculated from the mean state is identical to that implied in the surface forcing as expected.

Since the surface forcing is constant, the implied salt transport and the mismatch are both constant. This means that these fields do not respond to fluctuations in ocean circulation, and the ocean has to constantly adjust its flow and salinity to maintain the imposed salt transport. This can be seen in Fig. 10a, which shows the time evolution over an oscillation cycle of the salt transport at 32°N. In the course of attempting to balance the imposed constant salt transport, the ocean produces overshoots, and salt transports oscillate around the imposed. This process can be seen at all latitudes in Fig. 10b, which shows the salt transport anomalies covering a cycle. We see that the maximum salt transport anomaly occurs at the place where the mismatch in terms of implied salt transport is a maximum, that is, near 35°N (see Fig. 8). Figures 10c and 10d show the time series of overturning circulation and zonal mean salinity at 32°N. The anomalous net evaporation leads to the generation of anomalous salty surface water and subsequently to the strengthening of the meridional circulation (Fig. 10c, from the time when the overturning is at a minimum, near year 1410). The strengthening of the overturning then drives away the accumulated anomalously salty surface water through vertical convection, and in the mean time subducts the ambient rela-
tively freshwater water to the sinking region, leading to the decreasing of surface salinity (Fig. 10d, from year 1418). Hereafter, the surface salinity is determined by the competition between the effect of horizontal advection of the relative freshwater and the effect of evaporation. The decrease in salinity due to the former effect reaches a maximum soon after the overturning reaches its strongest phase (Fig. 10d, year 1355). Hereafter, the surface salinity increases as the anomalous evaporation continues, getting ready to enhance the overturning circulation from its minimum, and then the process repeats itself.

Figures 11 and 12 show the overturning streamfunction and zonal-mean salinity anomalies at different time of a cycle. We see that the zonally averaged salinity anomalies extend to over 1000-m depth. Figures 13 and 14 show the corresponding anomalies for SSS and surface velocities. Apparently the oscillation is associated with a large-scale anomaly of horizontal circulation. Consider the northern overturning cell 28°N northward. Comparing the zonal-mean salinity anomalies with the meridional overturning anomalies, we see that when the overturning strength is stronger than the mean (negative anomaly, Figs. 8a and 8b), the zonal-mean salinity anomaly in deep-water formation region is negative (Figs. 9a and 9b). This is associated with the southward movement of the positive salinity anomaly in Figs. 9a and 9b. Then, when the overturning strength is weaker than the mean (positive anomaly, Fig. 8c), the zonal-mean salinity anomaly at midlatitudes is positive (Fig. 12c). This is associated with a northward shifting negative salinity anomaly, as shown in Figs. 9b and 9c. This relationship further demonstrates the mechanism of the oscillation. Since the surface freshwater flux is constant, the salinity anomaly in the deep-water formation region is controlled by the strength of advection from other latitudes, where the salinity is smaller. This means that when the overturning is stronger than the mean, salinity in the deep-water formation region decreases, as we have seen above. On the other hand, the strength of the overturning is itself affected by the salinity anomaly and can be expected to weaken in response to the decreases in salinity in the deep-water formation region (through the pressure gradient). In this way, the enhanced overturning circulation is the cause for its own decline, and the overturning

\[ \text{Implied Salt Transport Varying With Latitudes} \]

\[ F_{SW} \quad F_{SW-F_S} \]

\[ (1 \times 10^4 \text{ Cubic Meter/Sec}) \]

\[ \text{Latitude (in degrees)} \]

Fig. 8. Implied zonally integrated salt transport for run \( F_{SW} \) (solid line) and for the difference between runs \( F_{SW} \) and \( F_S \) (\( F_{SW} \) – \( F_S \)).
Fig. 9. Panel (a) shows the difference of implied freshwater flux between runs $F_{SW}$ and $F_{S}$ ($F_{SW} - F_{S}$). This indicates the horizontal mismatch pattern. Panels (b), (c), and (d) show the mean solution from run $F_{SW}$ of surface salinity, surface flow, and the flow at level 9, respectively. The flows along the model Drake Passage are not plotted because of extremely large vectors.
subsequently weakens and a reversed process takes place. The oscillation cycle can also be described by the residence time of a water parcel. Since there is a constant rate of evaporation in the deep-water formation region, the surface residence time of a water parcel determines the salinity anomaly there. When the overturning circulation is stronger than the mean, the surface residence time is short, the time for salinization is shorter, and negative salinity anomalies result. Such anomalies can be seen in Fig. 13a along the pathway of the southwestward flow (see Fig. 14a) at a time when the surface southwestward flow anomalies is enhancing. This relatively buoyant water then acts to reduce the strength of the overturning and, subsequently, to increase the surface residence time of water parcels. Water parcels are then cooled more than before, leading to the elimination of the buoyancy anomaly and its replacement by relative dense water (Fig. 13c) that sinks and enhances the circulation. In this way the oscillation repeats itself.

Fig. 10. Time series of an oscillation cycle of (a) salt transport at 32°N (dashed line) in comparison with the mean at this latitude (solid line), (b) salt transport anomalies at all latitudes, (c) overturning circulation at (32°N, 2350 m) (that is, a blow-up of an oscillation cycle shown in Fig. 5b), and (d) zonal-mean surface salinity at 32°N.
to the zonal phase difference of the salinity and meridional flow anomalies. Some of the anomalies are actually out of phase (Figs. 13 and 14). This means that the oscillation cycle described above in terms of zonal integral/mean quantities are the net upon cancellations of anomalies of different signs. We like to stress that locally the anomalies are generated by the mismatch between the divergence/convergence of the atmospheric freshwater flux and the oceanic salt divergence/convergence rate, and the generation mechanism of the local anomalies is similar to what we have described above, which involves a negative feedback process and

Fig. 11. Anomalies of meridional overturning streamfunction from the mean (a) at years 1340 (at the time with a negative maximum overturning), (b) at year 1370, (c) at year 1400 (at the time with a negative minimum overturning), and (d) at year 1420 for run $P_{SW}$ (see Fig. 5b).

Fig. 12. The same as in Fig. 11 but for zonally averaged salinity anomalies. Contour interval is 0.01.
Fig. 13. Horizontal pattern of SSS anomalies (a) at years 1340 (at the time with a negative maximum overturning), (b) at year 1370, (c) at year 1400 (at the time with a negative minimum overturning), and (d) at year 1420 for run F_{sw} (see Fig. 5b).
Fig. 14. The same as in Fig. 13 but for surface flow anomalies.
Fig. 15. Results from runs $M_{ST}$ and $M_{STW}$. Panels (a) and (b) are the zonally integrated meridional overturning, while panels (c) and (d) are the zonally averaged salinity.

overshoot. The negative feedback and overshooting processes take place simultaneously at locations where a local mismatch is significant, and this forms the Howard–Malkus loop seen in the horizontal anomaly fields at a given instant. Since at a given latitude the overturning at a depth is zonally integrated, the zonal variation is not reflected in this field; however, an examination reveals that the double-peaked feature in the overturning circulation (Fig. 10c) is in fact due to the dominance by SSS anomalies at different locations at different times within a cycle.

The characteristics of the oscillation in runs $F_{ST}$ and $F_{STW}$ are similar to what is described above. Plots (not shown) of horizontal SSS anomalies from both runs show similar anomaly patterns circulating in the Northern Hemisphere. Similar patterns will be seen in oscillations under mixed boundary conditions.

d. The oscillation period

Since the salinity anomalies are locally driven, vertical advection and deep current advection play a negligible role in the regeneration of the salinity anomalies. In fact, this feature has been reflected in Fig. 12, which shows that below 1000 m, salinity anomalies are confined to the deep convection region. Deep flow advection and vertical advection do play a role in transporting the anomalous water mass associated with the oscillation from the deep-water formation region. However, since the deep flow and vertical advections are both weak, the associated timescales are much longer. It follows that the oscillation period is determined by the decay and the regeneration of the salinity anomalies of the first 1000 m or so. Huang and Chou (1994) confirmed that the period of the oscillation depends on the timescale for water to move across the so-called Malkus–Howard loop; that is, $L/u$, where $L$ is the loop length scale and $u$ is the mean circulation velocity. The essential feature of the loop oscillation has been proposed by Welander (1986). In a series of sensitivity studies, Huang and Chou (1994) also found that the frequency of oscillations is very sensitive to a number of parameters, such as the amplitude and profile of the freshwater flux, the values of the vertical and horizontal diffusivities, and the horizontal grid resolution. They produced oscillations with periods ranging from 20 to several hundred years. The high sensitivity of the oscillation is confirmed here and in the study by Winton and Sarachik (1993). Since the loop is determined by several parameters, one has no control over the looping and hence no control over the oscillation period.

In the present study we have seen that the oscillation period in our experiments ranges from 43.8 to 106.6 years. In the model studies of Weaver and Sarachik (1991a,b) and Weaver et al. (1993), oscillations of decadal scale are produced. In an experiment, hereafter called run $F_{STW}^{FW}$, an oscillation with a period of 19.9 years is produced. This run is forced by both the surface wind stress and the freshwater flux diagnosed from run $R_{STW}$ (see Table 3). However, the oscillation amplitude is much smaller than that in run $F_{STW}$. Apparently, the wind forcing has reduced the incompatibility. Note that the mean overturning circulation and the mean zonally averaged SSS are similar to those of run $F_{STW}$, although with wind-driven cells.

4. The mismatch mechanism in models under mixed boundary conditions

To examine whether the mismatch mechanism operates in models forced by mixed boundary conditions,
two more runs (runs $M_{STW}$ and $M_{ST}$) are carried out. Runs $M_{STW}$ and $M_{ST}$ are integrated from the end states of run $R_{STW}$ and run $R_{ST}$, with the surface forcing for SSS switched to the freshwater flux diagnosed from run $R_{STW}$ and run $R_{ST}$, respectively. Each experiment is run for another 1500 years. Again no lower-level acceleration is used. In either run, the thermohaline circulation in the Northern Hemisphere collapses, and a southern sinking cell similar to that observed in runs $F_{ST}$, $F_{STW}$, and $FW_{STW}$ settles in, as can be seen in Fig. 15. We see from Figs. 15a–b and 6c–d the similarity in overturning between runs $M_{STW}$ and $F_{STW}$, and between runs $M_{ST}$ and $F_{ST}$. The difference lies in the fact that the maximum overturning is now shifted slightly southward, apparently due to the restored temperature forcing. Figure 16 shows the time evolution upon switching to the mixed boundary conditions. We see that both runs $M_{STW}$ and $M_{ST}$ develop a Northern Hemispheric “polar halocline catastrophe” and that interdecadal variability takes place in run $M_{ST}$ but not in run $M_{STW}$.
a. The positive salt feedback under mixed boundary conditions

To understand the process that leads to the solution under mixed boundary conditions, it is useful to identify the oceanic processes involved. First, the net precipitation in the high latitudes will lead to freshening there. The freshening has an effect of weakening the strength of surface sinking, the overturning circulation, and the poleward heat transport. Second, if the temperature is dynamically determined, the temperature at high latitudes will cool due to the weakened poleward heat transport. This cooling will in turn intensify the sinking, offsetting the effect of freshening. The latter process is the oceanic negative feedback, which stabilizes the overturning. In runs $M_{STW}$ and $M_{ST}$, SST is restored. This disables the negative feedback mechanism. As the net precipitation continues, poleward salt transport weakens, and the surface water in the sinking region becomes increasingly lighter. Eventually convective activities terminate. This is the now well-known positive feedback process (Stommel 1961). The positive feedback process leads to the destruction of the thermally driven circulation in the Northern Hemisphere. At the same time, the cross-hemispheric pressure gradient associated with the freshwater flux becomes predominant. In this system with the positive feedback, it is this pressure gradient that dominates the final state of the solution. The above process is partially reflected in Fig. 16. We see that both the northern and southern sinking cells overshoot but return to a mean state.

In run $M_{ST}$ an oscillation with a period of 71.8 years ensues. In this case, to maintain the implied freshwater transport, the system requires SSTs to evolve and to play a role. But this is prevented by the Haney restoration condition; that is, the effect of SSTs is effectively eliminated. In this regard, this oscillation is quite similar to that in run $F_{ST}$, where the mismatch is provided by the elimination of thermal forcing. The striking similarity in the solutions under freshwater flux alone and under mixed boundary conditions can also be seen in a comparison of the zonally averaged salinity (Figs. 15c–d, and 7c–d). The freshening in northern high latitudes in runs $M_{ST}$ and $M_{STW}$ is the same as in runs $F_{ST}$ and $F_{STW}$. These similarities suggest that models under mixed boundary conditions may have problems properly representing the role of freshwater flux forcing.

The point that the positive feedback mechanism associated with mixed boundary conditions disables the thermal circulation and generates the mismatch is further tested in an experiment using Schopf’s (1983) boundary condition to provide the thermal forcing condition. Under Schopf’s boundary condition the model SST is restored to the atmospheric equilibrium temperature [diagnosed from the spinup of run $R_{ST}$, see Cai and Greatbatch (1995) and Zhang et al. (1993) for details] in a timescale of several hundred days (here 300 days). The forcing for salinity is the same as in run $M_{ST}$, that is, under the diagnosed salt flux. As in Rahmstorf and Willebrand (1995), this thermal forcing condition allows the negative temperature feedback between temperature and overturning. In this run, the northern overturning cell is stabilized by the negative feedback mechanism, and no mismatch between the implied atmospheric freshwater transport and the circulation is generated; therefore no oscillation is induced.

In run $M_{STW}$, some initial variabilities occur, but a steady solution with no oscillation settles. This no-oscillation solution indicates that any mismatch between the ocean circulation and surface forcing is not large enough to generate persistent oscillations. In an ocean driven by both thermohaline and wind forcings, there are a number of factors that can act to reduce the mismatch. First, temperature anomalies associated with the initial variability act to weaken the oscillation. But in run $M_{ST}$, the weakening effect from temperature anomalies is not enough to damp the oscillation. Second, we have already noticed that the effect of wind forcing in run $FW_{STW}$ is to reduce the mismatch. This effect of wind is also seen in the study by Winton and Sarachik (1993). In one of their runs forced by mixed boundary conditions, in the presence of a wind forcing, no oscillation is induced; however, upon switching off the wind the same run produces oscillations.

b. Features of the oscillation driven by mismatch mechanism in models under mixed boundary conditions

It should be noted that although in our model the positive feedback destroys the thermally driven regime, the positive feedback can also act to enhance it. For example, in the flat-bottom model of Moore and Reason (1993) upon switching to mixed boundary conditions, the North Atlantic overturning intensifies to more than double the size of that under restoring boundary conditions; that is, the system becomes more thermally dominant. After the enhancement, the system settles to a state with regular oscillations. In such a case a mismatch is created in an opposite sense to our case. In their case the intensified overturning and the associated circulation require an enhancement of the hydrological cycle, but the hydrological cycle is fixed by the implied constant freshwater flux forcing. It is likely that it is this mismatch that drives the oscillation in their models.

Huang and Chou (1994) have demonstrated that the property of the oscillation is determined by many model parameters, such as the structure and magnitude of the freshwater flux forcing and mixing coefficients. Although overturning oscillation in run $M_{ST}$ is weak, the salinity anomalies are strong. This can be seen from Fig. 17. Comparing Figs. 13 and 17, we see that the anomaly patterns are strikingly alike and that the anoma-
Fig. 17. Horizontal pattern of SSS anomalies at years (a) 1047 (at the time with negative maximum overturning), (b) 1065, (c) 1083 (at the time with negative minimum overturning), and (d) 1101 for run M_{ST} (see Fig. 16).
aly circulates mainly around the Northern Hemisphere. Two factors determine these similarities. First, the mismatch between the modeled circulation and the imposed freshwater transport is larger in the Northern Hemisphere than in the Southern Hemisphere. Second, in a circulation system dominated, or purely driven, by saline forcing, the SSS anomaly, and hence the looping, are generated in a similar process, with the same interrelationship between SSS anomaly and overturning.

An examination of the interrelationship between the salinity and temperature anomalies reveals that positive salinity anomalies are associated with positive temperature anomalies. That is, buoyancy anomalies due to salinity are partially offset by the effect of temperature anomalies. Under mixed boundary conditions, the use of the restoring boundary condition on the surface temperature produces a relationship in which enhanced heat loss is associated with anomalously high SST, while increased heat gain is associated with anomalously low SST. Such a relationship is the feature of the oscillation forced by the freshwater flux of the mixed boundary conditions. Recently, Deser and Blackmon (1993) have shown the variability with a timescale of 10 years (their EOF 2 analysis) and the variability with a timescale of 30–40 years. Although no clear relationship between temperature anomaly and heat flux anomaly is found in their EOF 1 analysis, the EOF 2 analysis shows that warm SST anomalies are associated with anomalous heat gain by the ocean, and cold SST anomalies with anomalous heat loss, which is in sharp contrast to what is described above. The generation mechanism of this observed variability in the EOF 2 analysis appears not to be the one presented in run $M_{ST}$ or in other models forced by mixed boundary conditions. Clearly, other mechanisms are in operation and remain to be explored.

### 5. Oscillations driven by constant heat flux

In pure thermally driven, or thermally dominant, circulation, the oceanic negative feedback between SST and overturning operates. Negative SST anomalies at high latitudes where convective activities take place enhance the meridional overturning, which in turn drives the high SST water from low latitudes, and reduces the anomalies. The inertia system may produce overshoots, driving oscillations. Forcing a model with heat flux imposes an implied atmospheric heat transport that needs to be balanced by the forced circulation. When an imbalance arises, the mismatch mechanism for internal variability discussed above operates.

To verify this point, we have carried out a similar set of experiments driven by diagnosed heat flux from four spinups. This set of experiments is detailed in Table 4. Run $R_T$ is forced by restoring the topmost level temperature to the specified temperature profile shown in Fig. 1a, with a timescale of 20 days. Run $R_{TW}$ is the same as run $R_T$ but is also subject to the wind forcing. Runs $R_{TS}$ and $R_{TWS}$ are identical to runs $R_{ST}$ and $R_{STW}$. The overturnings from the four spinups are similar to those in runs $R_{ST}$ and $R_{STW}$ and are not shown here. The similarity again indicates the predominance of the thermal forcing in the thermohaline circulation. The heat flux fields from four spinups are all similar: all feature substantial heat losses in the western boundary current and high-latitude regions, and heat gains in the low latitudes. In particular, there is a strong similarity between runs $R_T$ and $R_{TS}$ and between runs $R_{TW}$ and $R_{TWS}$, the latter again confirming the small wind effect. Another four runs under the diagnosed heat flux alone are carried out. These runs are termed as runs $H_T$, $H_{TS}$, $H_{TWS}$, and $H_{TWS}$ and are forced by the heat flux field from runs $R_T$, $R_{TW}$, $R_{TS}$, and $R_{TWS}$, respectively. All four experiments start from an ocean at rest. Salinity is kept uniform and constant at 34 psu.

The mean overturnings of the four cases are very similar to run $R_{ST}$. Solutions from runs $H_T$ and $H_{TS}$ are steady with no oscillation. The no-oscillation solution in run $H_{TS}$ indicates that no significant mismatch is created upon the removal of salinity forcing. Runs $H_{TW}$ and $H_{TWS}$, however, both produce similar solutions with a regular oscillation of period of about 25 years. The oscillations in the two cases are very similar due to the similarity in the forcing.

Figure 18a shows the time evolution of the overturning over the last 500 years of integration for run $H_{TW}$. This is taken at a location where during the spinup the overturning reaches a maximum. Figure 18b shows the zonally averaged SST anomaly for the same run covering 25 years. The interrelation between temperature anomaly and overturning anomaly is similar to that in the salt oscillation but the negative SST anomaly takes the role played by the positive SSS anomaly. At a fixed location, as the overturning strength weakens, the temperature decreases. Then, as the overturning strengthens, the temperature increases. The oscillation in this case is mainly confined to the Northern Hemisphere, although weak variability in the Southern Hemisphere is also present. This relationship between SST anomalies and overturning anomalies is very similar to that of the oscillation in the GFDL coupled

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**Table 4. Model runs for studying the thermally induced oscillation. See text for details of each experiment.**

<table>
<thead>
<tr>
<th>Run</th>
<th>Initial condition</th>
<th>Restoring on</th>
<th>Flux on SST</th>
<th>Wind</th>
<th>Period (years)</th>
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<tr>
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<td>SST and SSS</td>
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<tr>
<td>$R_{STW}$</td>
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<td>steady</td>
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ocean and the atmosphere. Cai et al. (1995) have shown that in the presence of a realistic feedback strength, oscillation can be generated, although the atmospheric heat change associated with this feedback weakens the oscillation.

6. Discussion and summary

We have described a series of numerical experiments using the Bryan–Cox model. These show the generation of interdecadal variability in an ocean model forced by constant two-dimensional freshwater or heat flux alone. The variability is driven by a mismatch between the implied atmospheric and oceanic freshwater (or heat) transport.

The variabilities of saline circulation in this study are similar to those produced by Huang and Chou (1994) under a natural boundary condition and with specified zonally uniform freshwater flux forcings. As in Huang and Chou (1994), the oscillations produced here have many similarities to those produced in models under mixed boundary conditions (Weaver and Sarno 1991; Weaver et al. 1993). Huang and Chou (1994) speculated that oscillations should also occur in a system forced by a two-dimensional surface freshwater flux field. However, the necessity, or the lack thereof, of a mismatch between the circulation and the forcing in generating oscillations was not addressed by their work. Our result indicates that oscillation will be generated once a significant mismatch exists between the surface forcing and the circulation of the forced system, regardless of whether the forcing is one- or two-dimensional. Given the high sensitivity to freshwater flux, forcing a model with a specified surface freshwater flux alone as in Huang and Chou (1994) will almost certainly provide a mismatch and generate oscillations.

Similarly, thermally driven oscillations can be generated under a constant surface heat flux forcing in which a mismatch exists between the local heat storage rate and the implied heat transport.

In a system driven by both the thermohaline forcings, the temperature anomaly acts to weaken the salt oscillation, while the salinity anomaly acts to weaken the thermally induced oscillation. For one type of oscillation to be predominant, the damping effect must be weak. Observed interdecadal variabilities (Deser and Blackmon 1993; Kushnir 1993) confirmed that such situations exist. Indeed, there is no dynamic reason why in the coupled atmosphere–ocean system the oceanic freshwater (or heat) transport is always in match with the atmospheric freshwater (or heat) transport. For example, the $E - P$ field is not fully determined by the ocean. Despite the role of SST, the $E - P$ field is also influenced by other nonlinear atmospheric processes. Hence, oscillation may be a fundamental feature of the system. Examination of the interrelationship between SST and ocean–atmosphere heat exchange is important
for understanding the nature of the variability. The observational study by Deser and Blackmon (1993) showed that the variability at a timescale near 10 years (their EOF 2 analysis) does not fit into the variability in models driven by mixed boundary conditions with Haney thermal restoring. Hence, variation of this timescale may be thermally driven. On the other hand, as is shown in Cai et al. (1995), thermally driven oscillation relies heavily upon the extent to which the atmosphere and the ocean are thermally coupled. For example, they have shown that thermally driven oscillation can only be induced under an atmosphere of small heat capacity (weak coupling), not under an atmosphere of infinite heat capacity (strong coupling, Haney restoring). In section 5 we forced the model with a constant heat flux field. Constant heat flux forcing allows the model ocean temperature to evolve freely, while maintaining a constant atmospheric heat transport. However, constant heat flux forcing means that there is no feedback between the ocean and the atmosphere. In reality some degree of thermal feedback between the ocean and the atmosphere exists. Cai et al. (1995) have produced oscillations in the presence of such a feedback.

We have also examined the role of freshwater flux in mixed boundary conditions. In the present study, the positive feedback mechanism associated with this type of forcing conditions is seen to have magnified the role of the freshwater flux forcing. This leads to a destruction in thermal circulation, which then yields solutions similar to those without thermal forcing. This suggests that models under mixed boundary conditions may have problems properly representing the role of freshwater flux forcing. Furthermore, it is shown that the generation of oscillation under mixed boundary conditions is due to the mismatch created as a result of the destruction of the thermal circulation.

It should be noted that the present study is a highly simplified mechanistic attempt to identify the generation mechanisms for internal variability. The original question we aim to address is "what is the response of OGCMs to mismatches between surface flux forcing and oceanic freshwater and heat transports?" The answer is that the system oscillates. A natural progression of the present study is to address what generalized extensions of the present conclusions would yield. For example, consider the entire spectrum of possible freshwater forcings for an OGCM. One might, for instance, have heavy precipitation in the polar regions, as is observed; or one might have precipitation only along the eastern boundary; or only in the western boundary current region; or in a checkerboard pattern; or in any other configuration which one can imagine. The question is "what class of forcings yields steady solutions and what class yields oscillating solutions?" Our result suggests that forcings that are consistent with the circulation they drive produce steady solutions, while those that are inconsistent generate oscillations. However, a quantifying answer to this question depends upon many factors such as the magnitude of these forcings, the damping effect by other forcings, model resolutions, and the model physics, particularly the vertical mixing strength. The high dependency upon these parameters has been shown in the study by Huang and Chou (1994). Further work attempting to address this question is being carried out in a global OGCM with realistic topography and driven by observed surface forcings. This will be reported in a future paper.

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