Climate Drift in a Global Ocean General Circulation Model

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ABSTRACT

A global version of the GFDL modular ocean model is forced using conventional restoring boundary conditions (BCs), mixed BCs (i.e., restoring the upper-level temperature but specifying a fixed salt flux), and stochastic fluxes of both heat and freshwater.

The climatology of the model is found to drift if stochastic freshwater fluxes are applied at high latitudes under mixed BCs. The drift is global in extent: the ocean is generally warmer in the North Pacific and Weddell Sea but cooler and fresher at depths elsewhere in the Southern Ocean and in the North Atlantic. There is a slight reduction (by about 5%) in the meridional overturning of the Southern Ocean and the North Atlantic. The drift of the barotropic flow is most pronounced in the Southern Ocean and is associated with a permanent meandering of the Antarctic Circumpolar Current.

The drift occurs within a few decades, suggesting that it may be important in enhanced greenhouse scenarios for early next century that have been obtained using coupled atmosphere–ocean GCMs. It is also possible that some of the intrinsic variability identified in the same models is actually a residual drift.

The drift depends upon convective adjustment to occur but can be amplified by the surface heat flux parameterization, both locally and by an additional feedback associated with large-scale flow changes. In an extreme case, the latter leads to a total collapse of the thermohaline circulation associated with North Atlantic Deep Water Formation. A similar mechanism underlies the drift that can occur when the switch from restoring to mixed BCs is made.

The heat flux feedback represents the atmosphere–ocean coupling in the model, so this aspect of the drift can be regarded as a coupled mode that actually contributes to the mean state of the coupled system. The existence of such modes makes some climatic drift in coupled models inevitable, if the individual components are equilibrated separately prior to coupling.

The applicability of these results to more sophisticated coupled models depends, in part, upon how well the restoring BC on temperature captures the heat flux feedback they exhibit.

1. Introduction

A number of climate models consisting of coupled global atmosphere and ocean general circulation models (GCMs), with sea ice components of varying sophistication, have been used to investigate climatic variability on interannual and decadal timescales. (See Meehl 1990, 1992 for recent reviews). If the surface fluxes are not adjusted to compensate for systematic model errors, then existing models tend toward a state in which each component exhibits a climatology quite different to the one it exhibited prior to coupling.

To avoid this systematic change or “*climate drift*” most groups employ some form of *flux adjustment* (e.g., Sausen et al. 1988; Manabe and Stouffer 1988; Manabe et al. 1991, 1992), in which information passing from one component of the coupled model to another is modified. This adjustment ensures that the climatologies of the fluxes each component receives is

not very different from those necessary to maintain the decoupled climatologies.

The adjustment methods are not entirely successful as some drift is still evident particularly at high southern latitudes (e.g., Cubasch et al. 1991; Mitchell 1992; Lunkeit, F. 1992 personal communication; Santer et al. 1993). One possible reason for this drift might be due to the fact that while the seasonal fluxes of heat and freshwater are (usually) adjusted, the deviations exhibited by the fluxes about these cycles are not (Sausen et al. 1988). As a result, the presence of nonlinearities in the various model components (and in the coupling terms) can lead to an asymmetric response to these deviations, that is, a net drift.

This is especially important if the corrections are based on separate integrations of the individual components. For example, usually monthly climatological atmospheric data (or its proxy) is employed to force the decoupled ocean model. As a result, the impact of atmospheric variability on most timescales is neglected. When the models are coupled, however, there is tremendous variability on daily, day-to-day, and longer timescales.

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One way of helping to alleviate this problem might be to calculate the flux adjustments during a preliminary coupled phase, but even then there is no guarantee that the statistics of the deviations under the artificial restoration required during this phase will remain the same once the restoration is abandoned and the models are fully coupled (and flux adjusted). This is because the more extreme fluxes will tend to be moderated by the restoration employed. As a result, once the OGCM (ocean GCM) is coupled it will find itself in a new environment. Similarly, the coupled atmosphere will begin to “see” sea surface temperatures more extreme than it had previously.

Here we will employ various surface boundary conditions and apply both heat and freshwater flux noise to a global version of the Geophysical Fluid Dynamics Laboratory at Princeton OGCM, to see if it has any effect upon the climatology exhibited by the model.

A number of studies have already applied freshwater flux noise to ocean models, but all were primarily interested in variability rather than any mean drift. Mikolajewicz and Maier-Reimer (1990), for example, found that the Hamburg Large-Scale Geostrophic OGCM exhibited pronounced variability in a frequency band centered around 320 years. Weaver et al. (1993) found that the decadal variability exhibited by their flat 60° single basin version of the GFDL model could persist in the presence of stochastic forcing under certain circumstances. Mysak et al. (1993) employed a flat, latitude–depth, single basin version of the model developed by Wright and Stocker (1991) and also found variability under stochastic forcing, provided that the vertical diffusion was small enough.

Of more interest here, however, is the mean drift or shift in the climatologies exhibited by these models under these conditions. Mikolajewicz and Maier-Reimer (1990) found that a mean drift in the magnitude of the transport of the Antarctic Circumpolar Current through Drake Passage (from 126 to 118 Sv; $\text{Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) occurred, while Mysak et al. (1993) concluded that the model they employed was generally stable to noise but that quite different equilibrium solutions could be obtained for certain combinations of the horizontal and vertical diffusivities. In a more recent study, Stommel and Young (1993) showed that multiple equilibria could exist for a simple box model under stochastic forcing.

In this study we will apply the stochastic fluxes to different parts of the globe under various accompanying tracer flux conditions. Additionally, different versions of a simple 1D forced diffusion model with convective adjustment will be used in conjunction with the OGCM results, to help determine some of the dominant underlying mechanisms of the drift.

The remainder of the paper is divided into five sections. In the following section we will briefly describe the OGCM and the various kinds of tracer boundary conditions employed. The results from the OGCM and

the 1D model are presented in sections 3 and 4, respectively. Here, we take the view that the OGCM, when combined with simple surface flux feedback parameterizations, constitutes a very crude coupled model. This is done to highlight the uncertainties associated with applying the results to more sophisticated coupled models. These results, together with the uncertainties, are discussed in section 5. The paper concludes with a summary of the main findings in section 6.

2. The OGCM and tracer boundary conditions (BCs)

a. The OGCM

The OGCM used here is a recent version of the GFDL code (Pacanowski et al. 1991) based on the work of Bryan (1969) and Cox (1984) and given the acronym MOM (modular ocean model).

The horizontal tracer grid was initially designed for coupling to an atmospheric model (Moore and Gordon 1993). It is compatible with an R21 Gaussian grid, which has a longitudinal spacing of 5.625° and a latitudinal spacing of approximately 3.2° .

There are 12 vertical levels ranging in thickness from 25 m at the surface to 900 m in the deep ocean. The horizontal eddy viscosity is artificially large ($9 \times 10^5 \text{ m}^2 \text{ s}^{-1}$) to ensure that the western boundary currents have horizontal scales that are resolved by the coarse grid. The horizontal eddy diffusivity is $2.5 \times 10^3 \text{ m}^2 \text{ s}^{-1}$. The vertical eddy viscosity and diffusivity take the same numerical values of 20×10^{-4} , 1.5×10^{-4} , and $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in the first, second, and subsequent levels, respectively, in order to crudely simulate a surface mixed layer. Further details regarding the configuration of this model have been given by Moore and Reason (1993) and Power and Kleeman (1993).

Most of the results that will be presented here represent quasi-equilibrium solutions, which have been obtained using the acceleration techniques discussed by Bryan and Lewis (1979) and Bryan (1984): the time step is increased with depth and a much larger time step is used in the tracer equations than in the barotropic and baroclinic equations. A time step of 2 days is used for the tracer equation and 1200 s for the baroclinic and barotropic equations. Acceleration factors ranging from 1 at the surface to 8 at the bottom are also employed. All of the experiments are integrated for at least 200 000 surface time steps to ensure that they are close to equilibrium (e.g., the globally averaged salinity has a trend less than or equal to approximately 1.8×10^{-3} ppt per century in all cases).

b. The tracer BCs

The preliminary experiments conducted are represented in Fig. 1. First, the OGCM is integrated while restoring both the upper level temperature (T) and

FIG. 1. A schematic representation of the experimental method employed. Here **R** is the solution obtained under restoring boundary conditions (BCs), **M** the solution obtained under mixed BCs, and **N** the solution obtained under mixed BCs with an additional stochastic (or noisy) tracer flux. The legs were integrated for either 200 000 (**R** and **M**) or 300 000 (**N**) 2-day (surface) time steps with an acceleration factor varying from unity in the top level to 8 in the deepest level.

salinity (S), with a time lag of 20 days, to annually averaged climatological values based on the data compiled by Levitus (1982). This leads to the restored solution represented as **R** in Fig. 1. A number of ocean modelers (Stommel 1961; Welander 1986; Bryan 1986; Weaver and Sarachik 1991) have argued that while the restoration of T is justifiable on the basis of observational work of Haney (1971) and Oberhuber (1988), there is no compelling reason to suppose that sea surface S anomalies are damped by atmospheric processes. To model this a fixed freshwater flux condition is imposed. The term “mixed BCs” is sometimes used to refer to this condition in combination with the conventional restorative condition on T .

In order to apply mixed BCs we need to have some estimate of the freshwater flux field. One way of obtaining this estimate is to diagnose it from the restored solution **R**. This is done over a period of about 50 surface years [after Weaver and Sarachik (1991) and Moore and Reason (1993)]. Averaging is necessary because the surface fluxes fluctuate, especially at high latitudes where intermittent deep convection occurs. Then **R** is integrated further under mixed BCs to obtain the second solution **M**.

c. The noise

In order to obtain the third solution **N** (also depicted in Fig. 1), a stochastic flux is also applied in conjunction with the fluxes described above. The noise is white, normally distributed with a zero mean and has a standard deviation of 1.8 mm day^{-1} for the freshwater flux and 50 W m^{-2} for the heat flux. The noise has no spatial (or temporal) coherence. The first value lies between the extremes considered by Mysak et al. (1993) and Weaver et al. (1993). Recent results from a 10-yr integration of the BMRC atmospheric GCM using realistic SSTs rather than climatological values (R. Coleman 1993, personal communication) indicate that this choice is rather modest, especially at low latitudes [see also Latif et al. (1990) and Mikolajewicz et al. (1993)]. During January 1980, for example, there are regions where the standard deviation of the precipitation ex-

ceeds 25 mm day^{-1} . As a result the magnitude of the drift exhibited here can probably be regarded as a lower bound. The value chosen for the heat flux is rather arbitrary, but we shall see that heat flux noise plays only a minor role.

A relatively large number of experiments are conducted in order to help determine the dynamics of the drift. The noise is applied (i) globally, or is restricted to either (ii) an equatorial band between about 25° N and S , or (iii) high latitudes north (south) of 55° N (S). Heat and freshwater flux noise is applied together in some experiments and separately in others.

3. Results

a. The drift

To begin with, consider the equilibrium solution obtained after integration under mixed BCs with a noisy freshwater flux applied at high latitudes only. The model climatology has drifted away from the control. To illustrate this, first consider the difference in the barotropic transport streamfunction before and after the noise is added (i.e., $N - M$) in Fig. 2a. The drift is most pronounced in the Southern Ocean and is consistent with an anomalous meandering of the Antarctic Circumpolar Current. The drift is modest, amounting to about a 2.5 Sv change in the zonally averaged horizontal transport (Fig. 2b).

The ocean is generally warmer and saltier at depth in the North Pacific and Weddell Sea but cooler and fresher at depth elsewhere in the Southern Ocean and in the North Atlantic (Figs. 2c,d). Other differences include a surface freshening of the polar oceans (except along the section of Antarctic coast south of Africa and the Indian Ocean), a surface cooling over the Southern Ocean (except in the southwest Pacific), and a weakening of the (maximum) magnitude of the overturning in the Southern Ocean and in the North Atlantic (associated with North Atlantic Deep Water Formation) by about 5%. The evolution of the latter is depicted in Fig. 3a (along with the upper-level temperature of the Southern Ocean in Fig. 3b).

FIG. 2. The difference $(N - M)$ in (a) the barotropic transport streamfunction and (b) its zonal average; (c) the temperature and (d) salinity at a depth of 2350 m, when a white noise freshwater flux is applied at high latitudes only. Contour intervals of (a) 2 Sv, (c) 0.1°C , and (d) 0.01 ppt are used. Dashed lines indicate negative values.

There has also been a slight increase in the rate at which convective events occur at most depths and at most locations. There are exceptions to this. First, the rate has been *reduced* in the northern North Pacific and North Atlantic, off the West Australian coast, around northern New Zealand, and over a significant fraction of the Southern Ocean. Second, there is a *substantially* increased rate in a few regions near the Antarctic coast. These results are consistent with those of Lenderink and Haarsma (1994), who showed that there can be regions which have the potential to convect, while in other regions convection is easily suppressed.

Similar experiments were conducted under the various BCs described in the previous section and the results are presented in Fig. 4. The drift is again measured here as the maximum change in the barotropic transport streamfunction in the Antarctic Circumpolar Current (ACC). [The connection between the thermohaline forcing and the ACC is no coincidence:

Power et al. (1993) have shown that a barotropic ACC of about 80 Sv can be generated in this model under restoring BCs by thermohaline forcing with no wind forcing whatsoever. This is consistent with findings in other ocean models (W. Cai, D. Olbers 1993 personal communication).] Recall that restoring, mixed, and fixed BCs are considered. The symbols $\mathcal{T}$ and $\mathcal{S}$ indicate that noise has been applied to the temperature or salinity equation. (The symbol " $\mathcal{T} \& \mathcal{S}$ " indicates that noise was applied to both.) The 3D plot depicts the results when the noise was applied in the equatorial band (bars closest to the front), globally (middle bars), or at high latitudes only (bars at the back).

The first thing to notice is that there is very little drift if the noise is applied at equatorial latitudes alone. Second, the drift is small under restoring BCs on *both* T and S . Third, the greatest drift occurs under mixed BCs and then only if a noisy freshwater flux is applied. The last point to note is that under the fixed flux conditions the drift due to the imposed noisy freshwater flux is substantially reduced. Thus, significant drift de-

FIG. 3. (a) Evolution of the maximum magnitude of the overturning streamfunction in the North Atlantic for *M* (control experiment, no noise added), *N* (freshwater noise added at high latitudes under mixed BCs), and BIGN, an additional experiment similar to *N* except that the standard deviation of the noise is increased ten-fold. (b) Evolution of the average of the Southern Ocean. Time is measured in units of 1000 surface days.

depends upon the restorative condition on *T* as much as it does on having a noisy freshwater flux at high latitudes.

The impact of the noise is not immediate, however, nor is the OGCM totally insensitive to the noise in the absence of the restoring BC. In fact the heat flux feedback plays only a minor role in the very early response to the freshwater noise. To illustrate this, freshwater is exchanged impulsively in the upper level of the OGCM. It is added in one case (*p*) and subtracted in another (*m*). The linear combination $p + m - 2c$, where *c* is a control experiment in which no freshwater anomaly is applied, is depicted in Fig. 5, first with the heat flux feedback operating (Fig. 5a) and then when it is not (Fig. 5b). The temperature in a vertical section along 65°S is presented. If the response was perfectly symmetric, then the field would be zero. In both cases the response to the imposed anomalies is asymmetric. We know, however, that the eventual drift is small without the heat flux feedback. So despite the early asymmetry in the response, the water column eventually settles

down into something very similar to the control, unless the surface feedback operates. When the acceleration methods are employed, the bulk of the adjustment in *N* occurs within the first 2000 surface days.

b. Why no vacillation?

Mikolajewicz and Maier-Reimer (1990) also applied a noisy freshwater flux (globally) to their large-scale geostrophic model and they found that in addition to a modest drift there was a significant vacillation with a period of about 320 years. In contrast, the response evident does not exhibit any significant variability. This is illustrated in the evolution of the maximum meridional overturning associated with North Atlantic Deep Water Formation (Fig. 3a) and the average temperature of the entire Southern Ocean (Fig. 3b). This difference in behavior will be discussed at length in section 5.

c. Timescale of the drift

We have seen that the climatology of the OGCM under mixed BCs drifts. In transient, enhanced CO₂ experiments, scenarios are often given for early next century. So if the drift occurs in the first 40 years or so, then it will be of some significance in this context.

The rates of change quoted above applied to the accelerated case in which timescales are distorted. In order to obtain a more faithful estimate the acceleration techniques were abandoned, and the model integrated for 4500 days. By this time the drift had reached 4 Sv ($\text{Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) in the barotropic transport and was

FIG. 4. The stochastic drift (Sv) as measured by the maximum change in the barotropic transport for the various experiments conducted. Noisy fluxes are applied to the temperature (*T*) and salinity (*S*) equations either globally, at high latitudes or in an equatorial band, under either restoring, mixed, or fixed BCs.

FIG. 5. Linear combination ($p + m - 2c$) of the temperature in a zonal section along 65°S after freshwater has been impulsively exchanged with the top level. Freshwater was extracted in p and added in m . The symbol c stands for the control experiment in which no anomalous freshwater was exchanged. Two cases are considered one in which the heat flux feedback operates (a) and the other in which it does not (b).

again restricted to the Southern Ocean. Thus, the drift does occur quickly enough to be of significance in CO_2 experiments, and certainly in experiments aimed at diagnosing variability on the same (and longer) time-scales.

An additional control experiment was also conducted in which the integration was performed synchronously but no noise was added. In this case the drift amounted to no more than 0.9 Sv after the same period. Integration of both the anomaly (N) and control (C) runs for a further 13 500 days produced drifts of 7 and 1 Sv for N and C, respectively, indicating that the bulk of the drift is stimulated by the addition of noise rather than by the abandonment of the acceleration techniques.

d. Noise amplitude and heat flux feedback

One might expect that the drift is a strong function of the noise amplitude. In order to determine the dependence of the drift upon the amplitude, a number of further experiments under mixed BCs were conducted, with a noisy freshwater flux again applied at high latitudes only. In each additional experiment the standard deviation of the noise was altered, and the acceleration techniques specified previously were again employed to obtain new equilibrium solutions. The results are depicted in Fig. 6, where we see that the magnitude of the drift does indeed increase with the standard deviation of the noise but that the relationship is not a near-linear one [as found by Mysak et al. (1993) in their simplified model].

FIG. 6. The net drift in the barotropic transport (Sv) when the freshwater flux noise (applied at high latitudes only) has an increasing standard deviation, from 0 to 1.8 mm day⁻¹.

Given the importance of the heat flux in determining 1) the magnitude of the drift (section 3a), 2) OGCM climatology in general (Maier-Reimer et al. 1993; Moore and Reason 1993) and 3) the response to surface freshening (Power et al. 1994; Power and Kleeman 1994; Zhang et al. 1993), it seems reasonable to suppose that it also plays a role in determining the response to the freshwater noise. To test this the relaxation time constant τ in the heat flux formula $Q = Q^* + \rho C_p h(1) T' / \tau$ (where Q^* = the heat flux diagnosed from M , ρ the density, C_p the specific heat, $h(1)$ the upper-level depth, and T' the deviation of the upper-level temperature away from its mean value in M) was varied. Given the scale dependence of the atmosphere's ability to dampen SST anomalies (Bretherton 1982; Schopf 1985; Kleeman and Power 1995) weaker restoring conditions were considered: $\tau = 60$ and 600 days, in addition to the value of $\tau = 20$ days already employed. The drift was reduced to only 5 Sv (using the same criterion as in Fig. 4) if $\tau = 60$ days and only 0.3 Sv if $\tau = 600$ days. It is interesting to note, however, that significantly larger drifts can occur even with the weaker restoring if the amplitude of the noise is large enough: A drift of 4 Sv occurred when noise with a standard deviation of 18 mm day⁻¹ was applied with $\tau = 600$ days. When the same noise was applied with $\tau = 20$ days, the overturning in the North Atlantic completely collapsed (Fig. 3a)!

4. Forced 1D diffusion model with convective adjustment

We have seen that under mixed BCs the climate of the OGCM drifts if a noisy freshwater flux is applied at high latitudes. If the ocean model were completely linear then there could be no mean drift. An imposed positive anomaly would yield a response with the same magnitude (but opposite sign) to a negative anomaly. Evidently there is a nonlinear response to freshwater anomalies at high latitudes.

Deep convection is generally restricted to high latitudes and is known to be sensitive to freshwater anomalies under mixed BCs (Bryan 1986; Maier-Reimer and Mikolajewicz 1989; Marotzke 1989; Wright and Stocker 1991; Stocker and Wright 1991; Power et al. 1994; Power and Kleeman 1994; Zhang et al. 1993). It is also modeled as a switch—convection only occurs if the water column becomes statically unstable, which we have already mentioned is only intermittently the case. Furthermore, the suppression of deep convection has been suggested as the mechanism by which a nonlinear response to surface heat flux forcing can occur in a global OGCM (Bryan and Manabe 1985; Bryan et al. 1984). Deep convection is therefore a likely candidate for the nonlinearity underlying the drift exhibited here.

In order to understand how this nonlinear response can occur consider the 1D diffusion equation; that is,

$$T_t = (\kappa T_z)_z, \quad (1)$$

where the subscripts denote partial differentiation; t is the time and z the vertical ordinate. Convective adjustment is presumed to operate whereby the vertical diffusion coefficient $\kappa = \kappa(z)$ is set to some small background value (1 cm² s⁻¹) if the density profile is statically stable but increased substantially (to 10⁵ cm² s⁻¹) if the profile becomes unstable.

The climatology of this simple model has been shown to drift under the action of sinusoidal surface heat flux forcing by Wetherald and Manabe (1972)—a surface mixed layer is formed. In fact, drift in this model can only be avoided if $(\overline{\kappa T_z})_z = 0$ at every level, where the overbar indicates temporal averaging. This equation holds if (i) $\kappa T_z = 0$ or if (ii) $\kappa T_z = f(t)$, a simple function of time alone. Equation (i) does not necessarily hold if convection is excited because a fluctuation in κ is a non-even function of a fluctuation in T_z and so they may be correlated in time. At $z = 0$ (the surface) $\kappa T_z = F(t) \neq 0$ (the surface flux), but $\kappa T_z = 0$ at $z = L$ (the bottom) because of the insulating BC. Thus, κT_z varies with z , and so (ii) does not hold either. Consequently, drift is possible if convection occurs.

Now consider two versions of a more general 1D model: one with a linear equation of state ($\Delta\rho = \alpha\Delta T + \beta\Delta S$, with expansion coefficients $\alpha = -0.2$ and $\beta = 0.8$) and one with the same nonlinear equation of state used in the OGCM [a third-order polynomial approximation to the Knudsen equation devised by Bryan and Cox (1972)]. Both models are integrated numerically (with T and S fixed in the bottom level) and are restored to values of $S(z)$ and $T(z)$, which approximate values taken from the OGCM (from solution M) at 65°S, near where deep convection occurs in the model. The noise is then applied under mixed BCs with the interior salinity and heat source terms (due to the processes not incorporated in the 1D model—that is, ad-

FIG. 7. The various (a) temperature and (b) salinity profiles exhibited by the 1D model in which a *linear* equation of state is used to test for vertical stability in the convective adjustment procedure. Here *M*: control, no noise added; *N*: noise added; *nq*: no heat flux feedback; *nc*: no convective adjustment.

vection and horizontal diffusion) fixed to values diagnosed from the restoring leg.

The results obtained when noise with an amplitude of 1.8 mm day^{-1} was applied are depicted in Fig. 7. It is evident that convective adjustment is essential for drift to occur in the model with a linear equation of state (Fig. 7a). This is because it is the only source of nonlinearity in the model. The drift is modest. It is characterized by a general cooling at most depths with *S* slightly increased by about 0.01 ppt in the near surface layers. Without the heat flux feedback the drift is substantially reduced, involving only a minor adjustment of the top layer.

The fact that the “restoring” BC on temperature actually helps to further destabilize the system at first seems counterintuitive. It should be kept in mind that it is the temperature in the top level that is being restored and not necessarily the density profile of the water column. To illustrate this, suppose the response to some given perturbation can be minimized by warming the surface level to increase the stability of the near surface layers. The restoring BC will work against this by tending to cool the upper level if its temperature increases, so a further readjustment of the water column will result. In short, if minimizing the

impact of a perturbation requires adjustment of the upper-level temperature, then the restoring BC can act as a destabilizing influence.

The results obtained in the model with a more complete equation of state are presented in Fig. 8. Again we see that no appreciable drift occurs without convective adjustment and that the heat flux feedback, while not essential for drift, can amplify the drift [in $T(z)$]. The nonlinear equation of state in the 1D model modifies the conditions under which convection will occur: A given T - S structure may be stable when a linear equation of state is used but might not be if the nonlinear equation were employed. The same nonlinearity also helps to boost the importance of freshwater forcing at high latitudes relative to lower latitudes because the buoyancy fluctuation corresponding to a given salinity fluctuation is larger at the lower temperatures found there.

The relative importance of the heat flux feedback in driving the drift is greater in the OGCM than it is in the 1D model. There are two reasons for this. First, we have only considered one temperature-salinity profile. Greater suppression of drift in the 1D model in the absence of the feedback might have occurred had other

FIG. 8. The various (a) temperature and (b) salinity profiles exhibited by the 1D model in which a *nonlinear* equation of state [devised by Bryan and Cox (1972)] is used to test for vertical stability in the convective adjustment procedure. Here, *M*: control, no noise added. In (a) *N*: noise added; *nq*: no heat flux feedback; *nc*: no convective adjustment. In (b) “Other” refers to all experiments *other* than *N*.

profiles been considered. The second reason lies with the fact that the 1D model provides only a subset of the possible sources of drift, since any alteration to the source terms via changes to the accumulation of heat and/or salt by anomalous advection or horizontal diffusion are neglected. Such changes can provide important additional feedbacks. The best known of these was first described by Stommel (1961). This particular feedback is essentially an expression of the way in which the enhanced poleward advection of salty water produces changes to the density structure of the ocean, which promotes a further enhancement of the advection. Conversely, reduced advection produces changes that promote further reduction. These mechanisms have been subsequently used to help account for the behavior of various OGCMs under mixed BCs (Bryan 1986; Marotzke and Willebrand 1991; Power and Kleeman 1994) and generalized mixed BCs (Rahmstorf and Willebrand 1995). Feedbacks associated with advective changes and the large scale could be expected to operate on much longer timescales than the 1D mechanisms. This suggests that it is this second kind of feedback that gives rise to the collapse in the overturning evident in Fig. 3, as it occurs on a very long timescale even if the acceleration methods are employed. It should be added that this feedback only operates if SST anomalies are damped at a sufficient rate (e.g., Power and Kleeman 1994; Rahmstorf and Willebrand 1995). Again, this is consistent with the findings here because the collapse was prevented when much weaker restoring was applied.

5. Discussion

We have seen that the drift exhibited by the OGCM is global in nature but causes the greatest change to the barotropic transport in the Southern Ocean. The greatest residual drift in a number of flux adjusted coupled models also occurs in this region (Cubasch et al. 1991; Mitchell 1992; Lunkeit 1992 personal communication). In such models the greatest source of freshwater flux noise at high latitudes presumably comes from fluctuations in ice-ocean coupling rather than in $E - P$. So while we have merely imposed the noise, in the coupled models brine rejection and ice melt might well be the major source of the noise.

The drift exhibited by the OGCM under mixed BCs is only significant under the restorative condition on T . In ice covered regions this simple parameterization might not satisfactorily model the heat flux, and so we should not assume that the drift here is governed by the same dynamics as it is in the more sophisticated coupled model. It does, nevertheless, provide a stop-gap explanation for at least part of the drift exhibited in some more sophisticated coupled models and certainly provides an excellent hypothesis for further research.

We again note that Mikolajewicz and Maier-Reimer (1990, hereafter MM) obtained a vacillation with a

period of about 320 years and variability on decadal timescales under stochastic white-noise freshwater forcing (Weisse et al. 1993). We did not. Perhaps the models occupy slightly different parts of parameter space and that both vacillation and drift are possible depending upon the choices made for the various model parameters (e.g., vertical diffusion—as in the model used by Mysak et al. 1993).

There are a number of other possible reasons for this difference in behavior, such as (i) convective parameterization, (ii) the noise applied, (iii) the background freshwater flux field, or other differences in the mean states of the models, (iv) the absence of nonlinear terms in the momentum equations of the MM geostrophic model, (v) differences in the heat flux formulation, (vi) their incorporation of a simple thermodynamic sea ice model, (vii) seasonal forcing, and (viii) our use of acceleration techniques.

Weaver et al. (1993) have investigated the impact of the first four of these differences upon the variability exhibited by their single basin OGCM, concluding that (i) behavior is little changed if convection is treated as either implicit vertical diffusion or by an iterative procedure provided that the latter completely stabilizes the water column after it is activated, (ii) the magnitude of the variability diminishes as the magnitude of the noise is increased, (iii) details of the background freshwater flux field can play a critical role in determining the behaviour, and (iv) the omission of the nonlinear terms is probably not significant in this context.

More recently, Barnett et al. (1993) showed that there is a threshold value (of 1 mm day^{-1} in their model) below which no variability will occur. Most of the experiments conducted here had noise imposed with a standard deviation above this level. Furthermore, the 320-yr vacillation modeled by MM was not present when noisy fluxes of heat and momentum accompanied the noisy freshwater flux in the presence of a weakened heat flux restoring BC (Santer et al. 1993). On the other hand, variability on multidecadal timescales was evident in both runs (Weisse et al. 1993; Santer et al. 1993) but not here.

We treat convection as enhanced, implicit vertical diffusion, whereas MM employed an iterative procedure (Maier-Reimer et al. 1993). Our noise has no spatial coherence—unlike that deployed by MM. While we used a larger standard deviation for our freshwater noise (1.8 compared with 0.53 mm day^{-1}) in most of our experiments, we also considered smaller values and found that no variability was evident even if 0.36 mm day^{-1} was employed. Thus, the absence of variability in our case does not lie with finding (ii). The background freshwater flux applied here and by MM can be expected to differ substantially since both reflect differing model parameter choices and model errors.

In this study we have simply restored towards annually averaged SSTs to obtain our heat flux whereas MM restore toward a seasonally varying apparent air

temperature that is calculated taking atmospheric heat advection into account. At high latitudes the presence of sea ice in their model partially weakens the heat exchange. Both of these differences are significant given the importance of the heat flux formulation at high latitudes in determining the model climatology (Maier-Reimer et al. 1993) and in determining the model response to high-latitude freshening (Power et al. 1993; Power and Kleeman 1994; Zhang et al. 1993).

Additionally, other differences in the mean state of the model might preclude the possibility of variability. For example, Toggweiler and Samuels (1992) have highlighted the importance of the position of Drake Passage relative to the maximum wind stress over the Southern Ocean on outflow and variability in the North Atlantic. Weisse et al. (1993) concluded that the relative isolation of the Labrador Sea in the Hamburg LSG OGCM was essential for the multidecadal variability to occur. So perhaps the models differ in these or other important but as yet unidentified regards.

In summary, either the different positions occupied by the two models in parameter space, differences in the mean state (perhaps due to orographic or bathymetric differences), differences in the noise applied, the convective schemes, or in the background freshwater flux forcing or the heat flux response, or our use of acceleration techniques remain as possible reasons for the difference in behavior. We hope to determine the reasons in a future study.

In addition to the climate drift exhibited by the model due to the noise, the climate drifts substantially when the restoring BCs are replaced by mixed BCs, as has been noted previously by a number of authors (Bryan 1986; Weaver and Sarachik 1991; Moore and Reason 1993; Power and Kleeman 1993). This second kind of drift has also been discussed as a source of drift in fully coupled models (Weaver and Sarachik 1991). This drift has been attributed to the fact that systematic model errors are no longer kept in check when the restoring BC on S is replaced by a fixed flux condition (Moore and Reason 1993). However, Power and Kleeman (1993) showed that if fixed fluxes of *both* freshwater and heat (diagnosed from the restored solution) are applied to the mixed BC solution, the ocean evolves into a state that is barely distinguishable from the restored solution. So the drift away from the restored solution requires more than just an accumulation of systematic errors.

We can apply the line of reasoning given for the stochastic drift to account for this drift, except that the noise in this case is generated internally by both systematic model errors and intermittent convective events. Thus, the 1D and 3D positive feedback mechanisms involving the heat flux feedback discussed earlier are probably responsible for this drift as well. It is important to note this drift is like the stochastic drift in the sense that it will only carry over to fully coupled models if the restoring condition adequately represents

the heat flux, particularly at high latitudes, which is not necessarily the case (Anderson and Willebrand 1992; Power and Kleeman 1993, 1994; Power et al. 1994; Zhang et al. 1993; Rahmstorf and Willebrand 1995). Nor will it necessarily carry over if dynamically related feedbacks involving either the momentum or freshwater fluxes become important (e.g., Power et al. 1994; Nakamura et al. 1995).

As the restoring BC represents the atmosphere-ocean coupling in the model, the stochastic drift seen here is, in part, a coupled phenomenon. So while we have investigated this problem from the point of view of exploring some of the limitations of flux correction, this drift highlights a simple yet fundamental truth about coupled modeling: If there are coupled modes in the system that actually contribute towards the *mean* climatic state, then climate drift is inevitable in a coupled model if the individual components are equilibrated separately. This remains true even if the components are ideal (in the sense that they include all relevant physics and contain no errors associated with discretization or parameterization). The coupled model needs to be integrated as a single unit to obtain an equilibrium solution, prior to imposing any perturbation.

At present this is of course prohibitively expensive in the context of climate models. If the residual drift (i.e., drift which occurs despite flux adjustment) exhibited by the coupled model is large enough to be of concern then, in the interim, we could devise schemes whereby statistics of the deviations are also corrected. It should be noted, however, that it is still possible that coupled, positive feedback mechanisms, which have been kept in check during the preliminary experiments (including any coupled leg during which flux corrections were calculated), are unleashed as soon as the "free-wheeling" coupled integration begins. As a result the problem of unambiguously identifying natural variability in fully coupled models will be with us for some time to come. Indeed Santer et al. (1993) showed that while certain features of the variability in the coupled control run performed by Cubasch et al. (1994) could be interpreted as bona fide natural variability, other aspects were consistent with a residual drift. Significantly, it was not always possible to distinguish one from the other.

6. Summary

The main conclusions are as follows:

- 1) The climate of a global version of the GFDL Modular Ocean Model (Pacanowski et al. 1991) and a simple 1D diffusion-convection model drift if forced by a noisy freshwater flux, especially under mixed BCs. Heat flux noise has very little impact on the OGCM under the same conditions.

- 2) The drift relies upon the asymmetric response of high-latitude deep convective regions to the noise. The 1D model was used to show that convective adjustment

is essential for drift to occur and that nonlinearities in the equation of state can play an important secondary role by modifying the conditions under which convection arises.

3) The drift can be amplified by the restoring BC on the upper-level temperature. This amplification occurs in two ways—a 3D mechanism associated with changes in the large-scale flow field and a 1D mechanism involving covariances between the heat flux and the convective adjustment.

4) Similar mechanisms underlie the drift exhibited by some ocean models when the switch is made from restoring to mixed BCs.

5) The drift is global in nature but causes the greatest change of the barotropic transport in the Southern Ocean, in the form of a permanent meandering of the Antarctic Circumpolar Current.

6) Significant drift occurs within a few decades. At no stage does this drift include any significant vacillatory behavior, although important caveats were attached to this particular result (e.g., the use of acceleration techniques).

7) The amplifying role played by the heat flux feedback can be regarded as a coupled phenomenon that actually contributes to the mean state of the coupled system. Coupled modes with this characteristic make climate drift inevitable if the individual components are equilibrated separately prior to coupling, even if those components are ideal.

8) The magnitude of the drift depends upon the amplitude of the noise and the rate at which the upper-level temperature is restored. Responses ranged from very modest to a complete collapse of the meridional overturning associated with North Atlantic Deep Water formation.

9) Most, if not all of the mechanisms underlying the drift described here exist in more sophisticated coupled models, and so the same mechanisms might therefore be responsible for some of the residual drift they exhibit.

It is hoped that a more faithful representation of the surface heat flux response as well as the magnitude and the spatial, temporal, and statistical character of the noise will be employed in a future study.

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