Influence of Convective Adjustment on the Stability of the Thermohaline Circulation

JOCHEM MAROTZKE

Institut für Meereskunde an der Universität Kiel, Kiel, F.R.G.

4 September 1990 and 12 November 1990

ABSTRACT

Three different convective adjustment schemes are employed in the GFDL GCM to investigate if the spontaneous collapse of the thermohaline circulation under mixed boundary conditions, as observed by F. Bryan, depends on the parameterization of convective overturning. It is found that both a procedure guaranteeing complete static stability and Cox's implicit vertical diffusion scheme avoid the spontaneous collapse. Both schemes are also insensitive to the choice of time step, whereas the standard GFDL convection algorithm in conjunction with mixed boundary conditions produces results that differ qualitatively from each other when different time steps are used.

1. Introduction

In large-scale ocean models the vertical acceleration and, thus, the explicit representation of convection is eliminated by the hydrostatic approximation. The response to static instability has instead to be parameterized by a convective adjustment procedure. Precisely what parameterization is the best representation of the physics is unclear since deep convection is still a poorly understood process (see Killworth 1989 for a review). In the ocean general circulation model (GCM) developed at the Geophysical Fluid Dynamics Laboratory (GFDL), convection is handled in a conceptually simple way (Bryan 1969). At the end of a time step each water column is checked for static instability between vertically adjacent grid boxes, in case of unstable stratification the boxes involved are mixed completely.

It has been noted (Cox 1984, Killworth 1989, Smith 1989) that the standard GFDL model convection scheme does not perform mixing as stated by Bryan (1969). The unstable region is not mixed completely, moreover convection occurs in a very irregular, episodic manner. While this may have only a minor influence in most model studies, the effect seems to be more drastic if mixed boundary conditions are applied for sea surface temperature (SST) and salinity (restoring condition for SST, specified freshwater fluxes for salinity). Bryan (1986a,b) reported that a steady state, which was stable under restoring boundary conditions for surface salinity as well as for temperature, was unstable under mixed boundary conditions although the freshwater fluxes applied were the ones diagnosed from the formerly stable solution. Eighty years after shifting over to mixed boundary conditions the polar halocline spontaneously spread equatorward and the meridional circulation collapsed within 20 years ("polar halocline catastrophe"). The onset of the collapse was traced to intermittency in convection, leading to the formation of a pool of low-salinity water, which then rapidly expanded until deep-water formation was completely closed off.

The aim of this note is to investigate whether the instability observed by Bryan (1986a,b) depends on how convection is implemented. To this end three different algorithms are employed to see if the halocline catastrophe is triggered in all cases (section 2). In section 3 the results are compared and discussed.

2. Model description and numerical experiments

The model used is the GFDL primitive equation GCM, based on the method described by Bryan (1969), in the version documented and distributed by Cox (1984). Apart from the model domain, which extends from the equator to 64°N and is 60° wide, all model parameters and the specification of the thermohaline and wind forcing are exactly as in Marotzke and Willebrand (1991); for brevity the description is not repeated here.

For the surface salinity two different boundary conditions are used. In the spinup experiments restoring boundary conditions are applied both for sea surface temperature and for surface salinity, with a time constant of 30 days. In the experiments following the spinup, the surface freshwater fluxes are specified (mixed boundary conditions), the choice of E-P is described below.

Three different convective adjustment procedures are compared, each of which checks for static stability

Corresponding author address: Dr. J. Marotzke, Center for Meteorology & Physical Oceanography, Dept. of Earth and Planetary Sciences, MIT, Cambridge, MA 02139.

at the end of a time step. In the standard GFDL scheme a fixed number N of double-pass iterations is performed (Semtner 1974; Cox 1984). Starting at the top of a water column, the first pass compares and, if necessary, mixes boxes 1 and 2, 3 and 4 and so on. On the second pass, boxes 2 and 3, 4 and 5 and so on are checked. The number of these double-pass iterations, N, is a freely variable parameter. Obviously it takes at least $KM/2N$ time steps until any information about density input at the surface reaches the bottom, where KM is the number of vertical layers. The degree of stability after the mixing process depends on vertical resolution, the number of iterations in the convection scheme, and the length of the time step, as already pointed out by Killworth (1989) and Smith (1989). A shorter time step will result in a more complete vertical stability.

Generally, the standard algorithm only asymptotically approaches a truly stable state. Suppose that the column is neutrally stratified initially and a positive surface density input is imposed. Since only two vertically adjacent boxes are checked in each iteration pass, instability will never really disappear. With convective adjustment every box shares its density anomaly with its neighbor below if density decreases with depth. Any stabilization between boxes K and $K + 1$ will be removed on the next iteration, as soon as $K + 1$ is mixed with $K + 2$.

Due to the noncompleteness of vertical mixing in the standard GFDL code, another question arises. In many ocean climate models, the technique of asynchronous integration is applied, i.e., using different time steps in the heat/salt and momentum/vorticity equations, respectively. As Bryan (1984) shows, this slows down internal gravity waves and external Rossby waves. The equilibrium solution, however, is assumed to remain unchanged, which may be questioned because of the dependence of convection on the length of the time step. To see if the steady state is actually identical for different time steps, some experiments described below are performed both synchronously and asynchronously. The time step was 2 hours for the momentum/vorticity equations and 5 days for temperature and salinity during the asynchronous integration, and 2 hours for all variables when the calculation was performed synchronously. It should be noted that it is only the length of the T–S time step that is relevant here, and not the question whether the time steps for T–S and momentum/vorticity, respectively, are equal or different.

Because of the possible, and unwanted, dependence of the solution on the time step, we developed an algorithm that guarantees complete vertical stability. This scheme was used in the experiments described in Marotzke (1989) and Marotzke and Willebrand (1991). It is an iterative procedure which during each iteration first checks if static instability occurs in a water column. If this is the case, the unstable part of the column is mixed. This is done in such a way that two or more neighboring boxes, which have equal potential densities, all are included in the mixing process if the uppermost (lowest) of these boxes is found to be less dense (denser) than its upper (lower) neighbor. This means that whenever two vertically adjacent boxes have been mixed, they are treated as a unit in all further iterations during the present time step. After at most $KM - 1$ iterations all static instability is removed.

The third algorithm employed here was developed by Cox and is distributed with the option to use it as an alternative to the standard scheme. It treats convective overturn as vertical diffusion; in the case of instability the ordinary diffusion coefficient is replaced by a very high one. For numerical stability reasons vertical diffusion is calculated implicitly. In a schematic, formal way the algorithm can be written for, e.g., the temperature equation as

$$T^{n+1} = T^* + 2\Delta t \partial_z (A_v(T^{n-1}, S^{n-1}) \partial_z T^{n+1})$$

where T^* is the temperature at time step $n + 1$ after all tendencies except vertical diffusion have been added. The coefficient A_v is the vertical diffusivity which takes on a high value (here 1 m^2 s^{-1}) if at time step $n - 1$ the stratification was unstable. Note that for numerical stability reasons all diffusion terms are calculated in a forward time step, the factor 2 is included to match the length of the leap-frog step, which is applied in the advection terms. The implicit formulation in Eq. (1) means that, after the vertical discretization, a system of equations is solved simultaneously for the entire water column.

From now on the convection schemes will be denoted "standard," "complete mixing," and "implicit" (for implicit vertical diffusion), respectively, in the order in which they were introduced. Basically the same set of experiments is performed for the three model versions (see Table 1). First, a spinup is run using restoring boundary conditions both for temperature and salinity as well as an idealized wind stress distribution. The spinup is integrated asynchronously until an approximate equilibrium is reached. From the last 22 years of this run a time average of the surface freshwater fluxes is calculated. The integration is continued with mixed boundary conditions, using the diagnosed evaporation less precipitation ($E - P$) rates. If the spinup steady state were a stable equilibrium it should not change during the subsequent integration.

Figure 1 shows the meridional streamfunction at the end of the spinup of experiment 1, using the standard convection scheme with $N = 3$. There is strong downwelling at high latitudes, the thermohaline cell has a maximum strength of 9.7 Sv (Sv ≡ 10^6 m^3 s^{-1}). After shifting to mixed boundary conditions the state displayed in Fig. 1 remains stable for 80 years, then the circulation collapses as in F. Bryan's experiments. This development is visualized by displaying the time series of the meridional streamfunction at a representative interior point (52°N, 2100 m depth, Fig. 2a). Subse-

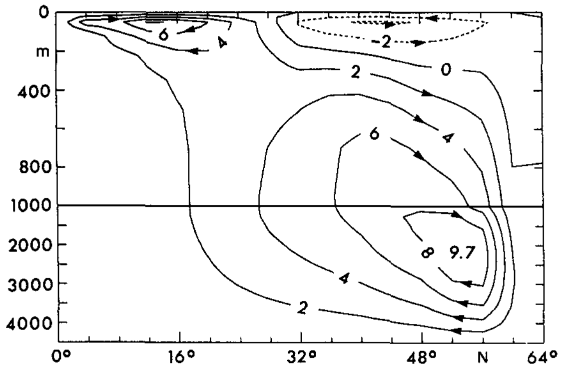

FIG. 1. Spinup steady state of experiment 1 (standard convection scheme). Meridional streamfunction in Sv.

quently the meridional overturning again increases in strength and reaches a maximum of about 30 Sv at year 200 after the shift in boundary conditions, later to collapse again.

The spinup of experiment 2 using the "complete mixing" convection scheme is very similar to the one from experiment 1 (see Fig. 1), maximum overturning strength is reduced somewhat to 9.0 Sv (not shown). Deep ocean temperature is 2.0°C, on a horizontal average, compared to 2.2°C in experiment 1 (see Table 1). The stability behavior, however, is completely different: after the switch to mixed boundary conditions the integration is continued for 2715 years, with the circulation remaining unchanged.

Overturning strength in the spinup of experiment 3 ("implicit" convection scheme, not shown) is in between the other two, with a maximum of 9.4 Sv and a very similar structure. The horizontal mean of the deep ocean temperature is 2.1°C (see Table 1). As in the "complete mixing" case the spinup steady state remains stable for at least 2715 years under mixed boundary conditions.

To test the robustness of the results three more experiments are performed for the three convection schemes in turn. They start from their respective spinup steady states and use the diagnosed freshwater fluxes, but a time step of two hours is applied in the T–S equations as well as in the momentum/vorticity equations (synchronous integration).

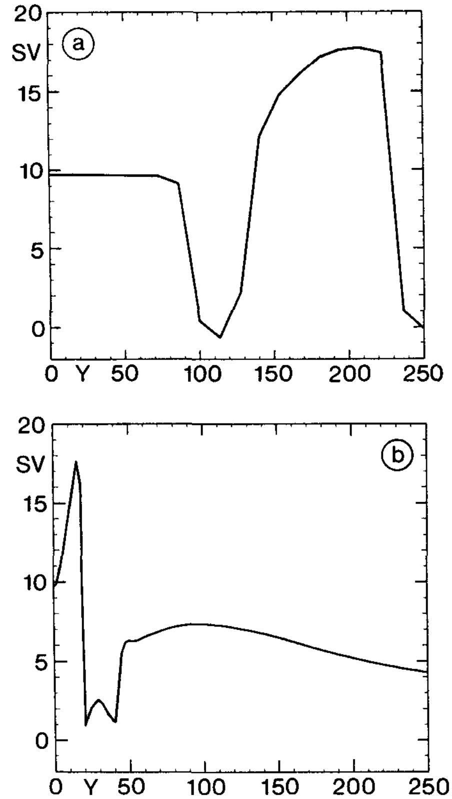

FIG. 2. Time series of the meridional streamfunction in Sv at 52°N, 2100 m depth, after the switch from restoring to mixed boundary conditions. (a) Experiment 1 (5 days time step for T–S), (b) experiment 4 (2 hours time step for T–S).

In experiment 4, which uses the standard scheme (see Table 1), this leads to an almost instantaneous drift away from the steady state (Fig. 2b). Initially the circulation increases in strength, with the maximum moving toward the northern boundary (not shown). After about 20 years a rapid and almost complete collapse follows, after 250 years the model apparently has settled to a new steady state with substantial downwelling at high latitudes.

It is not the topic here whether the time development visible in Fig. 2 will result in a stable equilibrium for any of the two cases. The important point is that with

TABLE 1. List of experiments. In column "spinup steady state" the horizontally averaged temperatures of the bottom layer and the maximum of the meridional overturning are listed.

Convection scheme	Spinup steady state (restoring boundary conditions) Time step: 5 days		Mixed boundary conditions Time step: 5 days	Mixed boundary conditions Time step: 2 hours
Standard	2.2°C	9.7 Sv	unstable (Expt. 1, Fig. 2a)	unstable (Expt. 4, Fig. 2b)
Complete mixing	2.0°C	9.0 Sv	stable (Expt. 2)	stable (Expt. 5)
Implicit diffusion	2.1°C	9.4 Sv	stable (Expt. 3)	stable (Expt. 6)

the standard convection scheme and mixed boundary conditions, the model behaves differently for the different time steps, the immediate drift in experiment 4 indicating that the spinup is not a steady state when a time step of two hours is applied. This is confirmed by the observation that if at the end of the spinup of experiment 1 (standard scheme), the integration is continued with restoring boundary conditions for salinity and a T–S time step of two hours, a slow drift sets in.

The situation is different in experiments 5 (complete mixing) and 6 (implicit) where the integration under mixed boundary conditions is performed synchronously, starting from the respective spinup steady states. In both cases the circulation remains stable, for at least 182 years in the complete mixing case and 136 years for the "implicit" scheme. It cannot be excluded that the circulation may eventually collapse if the calculation is continued. However, comparison of experiments 1 and 4 (see Figs. 2a,b) shows that if an instability occurs merely because of the switch to a shorter T–S time step, this is likely to happen shortly after the switch. In experiments 5 and 6 stability persists for considerably longer time than even in experiment 1, so for practical purposes it can be concluded that for the complete mixing and the implicit vertical diffusion schemes the circulation is stable, under mixed boundary conditions and synchronous integration.

3. Discussion

Why does the halocline catastrophe occur spontaneously only if the standard convection scheme is used and in different manners for different time steps? Consider the basin-averaged surface heat gain at the end of the spinup phase of experiment 1. Generally one would expect this quantity to approach zero when the integration is continued long enough. However, the time- and area-averaged heat loss in experiment 1 could not be lowered below 0.01 W m^{-2}. Figure 3 displaying mean surface heat uptake calculated at every time step demonstrates that it oscillates with a period of 20 time steps, a longer time series shows that the oscillation is extremely regular. 20 time steps is just the period, N_{MIX}, of how often a mixing time step is applied instead of the usual leap-frog step, in order to suppress the computational mode. During the mixing step, which here is an Euler-backward step, the time rates of change are multiplied by Δt, compared to $2\Delta t$ in the leapfrog case, before they are added to the fields at the previous time step. This has consequences if the standard convection scheme is applied.

Consider a high-latitude, vertically homogeneous water column, which is assumed to be cooled at the top through heat loss to the atmosphere and heated laterally in all layers below. In a steady state the column would be mixed completely, and lateral heat gain in all lower layers would cancel the surface heat loss. If a change in the time step occurs, however, with the standard convection scheme the information about the changed SST would be communicated down only to level $1 + 2N$, and in general the compensation of lateral heat gain and surface heat loss will not be exact, after vertical mixing has taken place. In the case considered here the topmost 7 layers ($N = 3$) are slightly warmer after a mixing time step, and consequently the heat loss to the atmosphere increases two time steps later (note that the surface heat flux is normally calculated as a forward time step with length $2\Delta t$); so the computational mode is created by the mixing time step rather than suppressed, as Fig. 3 shows.

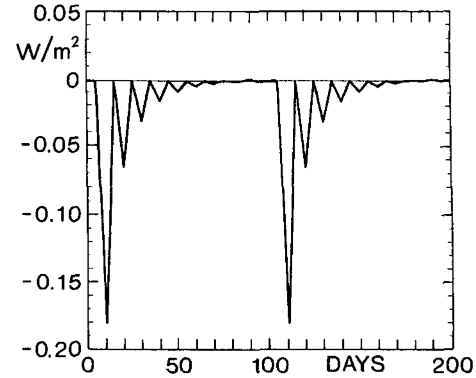

FIG. 3. Time series of the basin-averaged surface heat gain in W m^{-2} at the end of the spinup of experiment 1 (standard convection scheme).

It may appear from Fig. 3 that the model ocean is constantly losing heat on a temporal average. Actually, this is not the case: The total heat content of one of the two linearly independent solutions is reduced due to the heat loss at each other time step. However, on a mixing time step this "colder" solution is discarded again, which is equivalent to introducing artificial heat sources over the entire ocean. These artificial sources just compensate for the heat loss caused by the combination of the standard convection scheme and the shortened time step during a mixing step.

The oscillations visible in Fig. 3 do not occur with the other convection algorithms where the aforementioned compensation is exact. While this is obvious for the complete mixing scheme (experiments 2 and 5) it may be surprising for the "implicit" case (experiments 3 and 6) which does not achieve complete stability due to the finite diffusion coefficient. However, in the water column discussed above, the slight remaining static instability after calculating vertical diffusion causes high vertical diffusivities in the following time step. The information about reduced surface heat loss during a mixing time step is immediately communicated all through the water column, due to the implicit evaluation of vertical diffusion.

Convection occurs in a much more irregular fashion with the standard GFDL convection scheme, and the precise nature of this irregularity obviously depends on model parameters, which are usually not published (N, N_{MIX}). The experiments described here suggest that the spontaneous collapse of a spunup thermohaline circulation, which happens if subsequently mixed boundary conditions are applied, can be avoided if either the complete or the implicit mixing scheme is applied. This is because the freshwater fluxes diagnosed from the spinup show an analogous behavior as the heat fluxes, i.e., are much more regular in time with greatly reduced variance, in the case of complete mixing and implicit diffusion. One would also avoid the halocline catastrophe to occur merely because of the shift from asynchronous to synchronous integration, under mixed boundary conditions. The overall calculation in the model runs using the "implicit" algorithm consumed only 60% of the computer time, required for the overall calculation using the complete mixing scheme. The results presented here thus suggest that the "implicit" algorithm should be employed in all experiments that do not use a restoring condition for sea surface salinity.

It should be noted that the sensitivity of the spinup steady state under mixed boundary conditions is not created by the standard convection scheme. With the other two algorithms, the halocline catastrophe can be triggered if the spinup steady state is perturbed by an initial salinity anomaly as small as -0.01 psu, simultaneously with the switch to mixed boundary conditions. This indicates that all spinup equilibria described here are very sensitive to perturbations, if they are used as initial states for runs with mixed boundary conditions (see Marotzke 1990 for details). The spontaneous occurrence of the collapse when the standard scheme is used, merely indicates that the irregular mixing represents a large enough perturbation for inducing the instability.

Obviously the choice of the convection scheme has a much larger influence on the model results if mixed boundary conditions are used instead of restoring boundary conditions both for temperature and salinity. This is especially important since mixed boundary conditions are a much better approximation to the large-scale interaction of ocean and atmosphere. Because the surface heat flux strongly depends on the sea surface temperature (SST), SST anomalies are rapidly removed by enhanced heat gain or loss (the typical time scale of SST anomalies with spatial scales of o(1000) km is a few months). The surface salinity, however, has negligible influence on evaporation and precipitation rates, and consequently surface salinity anomalies can persist on much longer time scales.

The different coupling of temperature and salinity to the atmosphere is responsible for the existence of multiple stable equilibria in oceanic circulation models (e.g., Bryan 1986b; Marotzke and Willebrand 1991), which differ fundamentally in the location of deep water formation. The parameterization of convective sinking may thus have an impact on which steady state is actually obtained in a specific numerical model study. How important this influence actually is, compared to, e.g., the consequences of the uncertainties in the freshwater fluxes, remains to be investigated.

The arguments in favor of the complete mixing scheme and implicit vertical diffusion must be understood to be based on purely practical reasons: Dependence on various model parameters is reduced and the system becomes better predictable. From a physical point of view it would be attractive to model convective overturning as being smoothly dependent on the density difference between upper and lower layers: In a model mean potential densities over an area of (typically) 100×100 km^2 or even more are compared, whereas the convection "chimneys" have horizontal scales of 10 km or less (Killworth 1983). Taking short weather events into account, there will be spots of static instability in a grid cell, whereas the mean density still is stably stratified, though weakly. Moreover, the vertical acceleration is likely to depend on the stratification.

The standard scheme may thus represent the stochastic nature of convection more closely; additionally it is not clear that convective events should be completed within a time step of, e.g., 2 hours, as the complete mixing scheme implies. The advantage of a more regular convection algorithm is that it makes it easier to explicitly specify the appropriate noise level, instead of using model noise of unknown characteristics.

Acknowledgments. I am indebted to Michael Cox for making the GFDL model code available to me and to Jürgen Willebrand for many discussions and suggestions. This work was supported by the Sonderforschungsbereich "Warmwassersphäre des Atlantiks".

REFERENCES

Bryan, F., 1986a: Maintenance and variability of the thermohaline circulation. Ph.D. thesis. Princeton University, 254 pp.
——, 1986b: High-latitude salinity effects and interhemispheric thermohaline circulation. *Nature,* **323,** 301–304.
Bryan, K., 1969: A numerical method for the study of the circulation of the world ocean. *J. Comput. Phys.,* **3,** 347–376.
——, 1984: Accelerating the convergence to equilibrium of ocean–climate models. *J. Phys. Oceanogr.,* **14,** 666–673.
Cox, M. D., 1984: A primitive equation, three-dimensional model of the ocean. GFDL Ocean Group Tech. Rep. No 1, GFDL/Princeton University.
Killworth, P. D., 1983: Deep convection in the world ocean. *Rev. Geophys. Space Phys.,* **21,** 1–26.
——, 1989: On the parameterization of deep convection in ocean models. *"Aha Huliko a": Parameterization of Small Scale Processes,* P. Müller, Ed., Hawaii Institute of Geophysics, 59–74.
Marotzke, J., 1989: Instabilities and multiple steady states of the thermohaline circulation. *Oceanic Circulation Models: Combining Data and Dynamics,* D. L. T. Anderson and J. Willebrand, Eds., Kluwer, 501–511.
——, 1990: Instabilities and multiple equilibria of the thermohaline circulation. Ph.D. thesis. Ber. Inst. Meeresk. Kiel, 194, 126 pp.
——, and J. Willebrand, 1991: Multiple equilibria of the global thermohaline circulation. *J. Phys. Oceanogr.,* **21,** in press.
Semtner, A. J., 1974: An oceanic general circulation model with bottom topography. Numerical simulation of weather and climate. Tech. Rep. No. 9, Dept. Meteor. UCLAR, 99 pp.
Smith, N. R., 1989: The Southern Ocean thermohaline circulation: a numerical model sensitivity study. *J. Phys. Oceanogr.,* **19,** 713–726.