An Efficient Convective Adjustment Scheme for Ocean General Circulation Models*

F. L. YIN AND E. S. SARACHIK

The Joint Institute for the Study of the Atmosphere and Ocean, University of Washington, Seattle, Washington

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ABSTRACT

The implicit vertical diffusion (IVD) convective adjustment scheme in common use in ocean general circulation models (OGCMs) could have large residual static gravitational instability at each time step. An iterative and explicit scheme is devised, based on similar physical considerations as the ones for the IVD scheme. It guarantees a complete removal of static instability in a vertical water column and is more efficient than the IVD scheme in overall spinup of the model.

The two convective schemes are compared in an ocean model that is in a state of interdecadal limit cycles. While the model solution with either of these two schemes is characterized by interdecadal oscillations, the variability is different in each scheme. The primary oscillation has a period of about 11 years, but the basin mean kinetic energy shows large differences. The 11-year cycle is modulated by a 33-year oscillation with the IVD scheme, while it is modulated by a 22-year cycle with the complete scheme. The amplitude of the variation of kinetic energy with the IVD scheme is also about twice as large as that with a complete adjustment scheme. It is therefore suggested that complete and incomplete convective schemes can lead to different model variability when convective changes in temperature and salinity have large variations over a short period of time.

1. Introduction

One of the ocean's important roles in the earth's climate is transporting heat from low to high latitudes. The overturning thermohaline circulation is the dominant process in modeled meridional heat transports. Since convective sinking is an essential driving mechanism of the thermohaline circulation, the treatment of gravitational convective adjustments is a crucial component of ocean general circulation models (OGCMs) (Bryan 1969).

There are mainly two types of deep convection in the ocean: convection near an ocean boundary and open ocean deep convection (Killworth 1983). Their basic causes are buoyancy losses at the ocean surface that gravitationally destabilize the water column so that dense water reaches the deep ocean through the subsequent overturning. Because deep water formation involves small-scale processes that are not resolved in large-scale OGCMs, it has to be parameterized (Bryan 1969) through convective adjustments. There have been a number of papers on convective adjustment schemes in large-scale ocean models (e.g., Bryan 1969; Semtner 1974; Cox 1984; Adamec 1986; Killworth 1989; Smith 1989; Marotzke 1991). However, because of the poor understanding of deep convective processes in the ocean, exactly how to represent the physics of convections in large-scale OGCMs remains unclear. In many present OGCMs, convective adjustments still follow the same original idea (Bryan 1969): whenever a water column is statically unstable, temperature and salinity are vertically adjusted to make the water column neutrally stable, with heat and salt conserved in the processes. Different convective schemes can be classified mainly in terms of the efficiency of removing gravitational instability, and if enough iterations are carried out, they are all physically equivalent in the sense that no static instability exists in a water column.

Marotzke (1991) investigated the influence of convective schemes on the stability of the thermohaline circulation upon switching from surface restoring (both temperature and salinity are restored to specified values) to mixed (temperature is restored, and salt flux is diagnosed from the equilibrium with restoring boundary conditions) boundary conditions. He compared three convective schemes commonly used in OGCMs: original standard scheme from the Geophysical Fluid Dynamics Laboratory (Cox 1984), the implicit vertical diffusion scheme (Cox 1984; Killworth 1989), and the complete adjustment scheme (Marotzke 1991) (hereafter we will refer to these three as the S, IVD, and M schemes, respectively). The spunup states with an asynchronous integration (longer time step for temperature and salinity than for velocities) under restoring surface boundary conditions were found to have some differences with these three schemes in that the solution

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Corresponding author address: F. L. Yin, JISAO, GJ-40, University of Washington, Seattle, WA 98195.

with the S scheme has a warmer deep ocean and larger meridional overturning than that of the M scheme, while the solution with the IVD scheme is between those of S and M. When switched to mixed surface boundary conditions, the state with a three-iteration S scheme suffered spontaneous drift with a synchronous integration or collapsed after about 80 years of an asynchronous integration, while the ones with the IVD and M schemes were still stable after more than 130 years of a synchronous integration or after more than 2700 years of an asynchronous integration. Marotzke also pointed out that the overall calculation for the model using the M scheme consumes about 40% more computer time than that using the IVD scheme.

In this paper, we will present a newly devised convective scheme (hereafter referred to as YS scheme), which is physically equivalent to the IVD scheme but guarantees a complete removal of static instability in a water column at each time step and is computationally more efficient than the IVD scheme in the overall spinup of the model. We will show that the IVD scheme could have large residual static instability due to the nature of the vertical finite difference treatment of diffusion regardless of choice of infinite or finite "convective diffusivity." Since there have been many discussions on the performance of the S scheme in the literature (Killworth 1989; Smith 1989; Marotzke 1991), we only present a comparison of the YS and IVD schemes. Our focus is on how much difference these two schemes can induce in an ocean model that is in a state of interdecadal oscillations. The rest of the paper is organized as follows: in section 2, we will first give a brief discussion of the S and IVD schemes, then present a detailed description of the YS scheme; section 3 presents a comparison of the IVD and YS schemes in a coarse resolution sector OGCM; section 4 summarizes and concludes the paper.

2. An efficient convective scheme

Gill (1982) gave a comprehensive discussion of static stability in the ocean. In general, the full effects of temperature, salinity, and pressure need to be considered to determine the stability of a water column. Using potential density to measure gravitational stability can be a good approximation for the upper ocean, but can induce a large error in the deep ocean. In the GFDL model, a local pressure is used to determine the static stability of two adjacent points for the purpose of minimizing the error introduced through pressure effects. Before going into a detailed description of the YS scheme, we first briefly discuss the S and IVD schemes since the YS scheme is based on the IVD scheme.

The S scheme takes an iterative approach (Bryan 1969; Semtner 1974; Cox 1984). For each iteration, every two vertically adjacent points are compared, and if unstable, their temperature and salinity are mixed so that they are neutrally stable with respect to each other. The iteration continues between all adjacent levels until the gravitational instability is removed in the whole water column. Because the adjustment acts on only neighboring points, the number of iterations required to reach the final stable state is infinite for a given unstable profile (Smith 1989). In practice, however, the number of iterations is always set to a finite number, and this leads to some residual instabilities in many cases (Killworth 1989; Marotzke 1991).

The IVD scheme was developed to remove these residual gravitational instabilities in the S scheme (Cox 1984; Killworth 1989). In this scheme, the stability between vertically adjacent grid points is tested and, if unstable, the vertical diffusivity is set to a large value (convective diffusivity) in order to smooth out the instability. In order not to reduce the time step caused by the large diffusivity, the transport equations of temperature and salinity are solved with an implicit time step vertically. This scheme is efficient in removing instability in a vertically unstable region, but less efficient in determining the final boundary of the adjusted vertical region. It is also computationally expensive as it requires an inversion of a tridiagonal matrix at each step. To illustrate this point, we consider a static ocean with vertically fixed uniform salinity. Imagine a neutrally stable water column with vertically uniform potential temperature. Now impose a negative temperature anomaly in the surface grid point. It is clear that the water column is gravitationally unstable near the surface and the final adjusted state is one with uniform potential temperature of reduced value. However, it will take at least the number of steps equal to the vertical grid points for the IVD scheme to adjust to such a profile since the negative anomaly can only move down one grid point every time step. The instability cannot be smoothed out in one step because zero temperature gradient below cannot transport heat in spite of the large "convective diffusivity."

Note that in this particular example it takes a number of steps equal to the number of vertical grid points for the IVD scheme to neutralize the profile to the bottom. In adjusted regions, the profile is always marginally stable due to the smoothing effect of the large convective diffusivity. Consider a water column with multiple unstable regions. Following the nature of a diffusive equation, the IVD scheme adjusts all unstable regions simultaneously by convective diffusive fluxes. After one time step, within each region the water column is marginally stable, but their upper and lower boundaries have expanded relative to the initial positions. Often there exist instabilities near the new upper and lower boundaries, and further adjustments are necessary in order to remove these residual instabilities. The IVD scheme is very efficient in an instantaneous adjustment within all unstable regions, but it is less efficient in determining the final upper and lower boundaries of the adjusted region since for every vertical grid point that the upper and lower boundaries of the region move to, the vertical convective diffusive equation has to be solved once.

Using a locally iterative method to determine the upper and lower boundaries of each adjusted region while keeping the instantaneous adjustment within each unstable region, a new convective scheme can be devised as the follows:

1) Find the total number of unstable vertical regions N, the beginning and ending indices b_n and e_n of region n ($n = 1, N$) in each vertical column (region $n = 1$ is the one closest to the surface). This is accomplished by comparing the density of adjacent points as in the S scheme. Each unstable region includes adjacent marginally stable grid points at the top and at the bottom since potential temperature θ and salinity S will be homogenized after adjustments.

2) First calculate the adjusted θ and S values (the mean) within each unstable region, then, if necessary, extend the upper and lower boundaries of each unstable region by recalculating the mean θ and S values after each new adjacent point is included in the unstable region until each region is marginally stable relative to its adjacent points at the top and at the bottom.

3) After step (2), there might be overlapping between region n and $n + 1$. There are only two types of overlapping. In the first case, one region contains the other, and the final result is to discard the region that covers the smaller vertical extent. In the second case, one is not contained by the other, and there will be a complete mixing between regions n and $n + 1$. The lower and upper boundaries of the new region needs to be determined iteratively as in step (2) after regions n and $n + 1$ are combined. This step repeats itself until there are no overlapping regions.

4) After steps (1)–(3), the adjustment of a water column is complete. The θ and S values are the mean within each convective region, and they are unchanged outside.

Figure 1 illustrates the algorithm discussed above for a water column with fixed salinity (the YS scheme includes salinity effects). The initial profile contains two unstable regions, which are identified after step 1. Step 2 adjusts each individual region iteratively. After step 2, the two regions overlap and step 3 produces the final profile. Note that the M scheme and the YS scheme are physically equivalent, and their difference is the method of iteration. While the YS scheme is more efficient than the IVD in spinning up the model, the M scheme is less efficient than the IVD (Marotzke 1991).

3. A comparison of the IVD and YS schemes in an OGCM

The major difference between the IVD scheme and the complete adjustment scheme is that the IVD scheme may leave residual static gravitational instability at each time step. If a time step of 1 day is used for the temperature and salinity equations as in asynchronous integration of many large-scale OGCMs with 12 vertical levels, the adjustment would be sufficient after about 12 days. Depending on the magnitude of surface buoyancy flux and stratification, which determines the convective time scale, the IVD scheme may not adjust the water column quickly enough, and at each time step, convection may not go as deep as it should. Because the thermohaline circulation is nonlinear, the incomplete adjustment by the IVD scheme may affect the subsequent dynamic adjustment. This can actually lead to significant difference in variability of the model in which convective rates vary over a short period of time. To illustrate this point, we compare thermohaline interdecadal oscillations in models with the IVD and YS schemes.

The model used is the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM) (Bryan 1969; Cox 1984; Pacanowski et al. 1991). It uses primitive equations. Potential temperature and salinity are governed by transport equations with convective adjustments. Turbulent transfer of momentum, heat, and salt are parameterized by eddy viscosities and diffusivities. The model geometry is chosen to be a rectangular

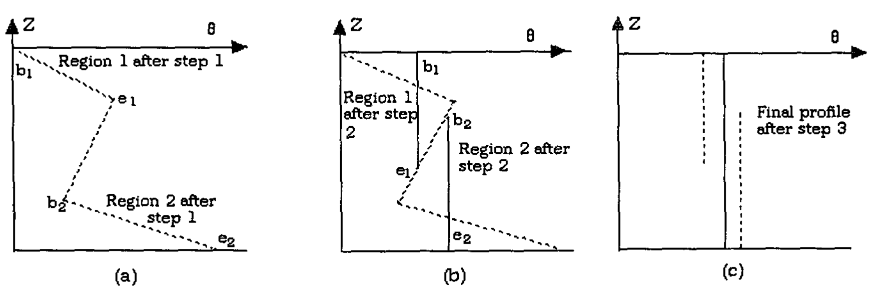

FIG. 1. Schematic diagram showing the algorithm of the convective scheme for an idealized water column with fixed salinity. (a) Initial potential temperature and the two unstable regions determined in step 1. (b) Adjusted potential temperature within the two regions after step 2. (c) Final profile after step 3. Solid lines represent the result after the step and dashed lines plot the result of the previous step.

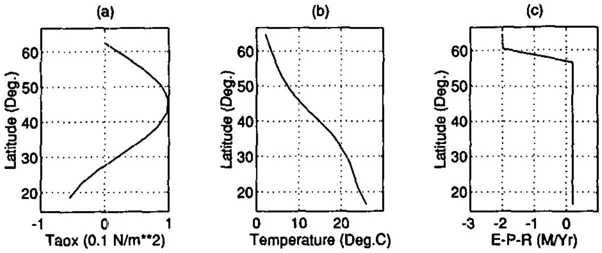

FIG. 2. Surface forcing function. (a) Zonal wind stress (0.1 N m^{-2}). (b) Reference restoring temperature (°C). (c) Freshwater flux (m yr^{-1}).

box in the earth's surface coordinates with a 5000 meter depth from 14.5° to 66.5°N, covering 60 degrees of longitude. The resolution of the model is 4° lat by 3.75° long, with 12 vertical grid points ranging from a grid size of about 50 m near the surface to about 780 m near the bottom. The lateral boundary is nonslip and it is also insulating to both heat and salt. There are no bottom stress, and no bottom heat and salt fluxes. The model is forced by zonal wind stress (meridional wind stress is not included), heat flux that is parameterized as restoring the surface temperature to reference value, and a freshwater flux (it is converted into an equivalent salinity flux with a constant reference salinity of 35 psu and a year of 360 days). Figure 2 shows these three forcing functions. They are zonally uniform and only vary meridionally.

The eddy parameters are chosen as follows: horizontal and vertical eddy viscosities $A_h = 2.5 \times 10^5$ $m^2 s^{-1}$ and $A_v = 10 \times 10^{-4}$ $m^2 s^{-1}$; horizontal and vertical diffusivities $K_h = 10^3$ $m^2 s^{-1}$ and $K_v = 1 \times 10^{-4}$ $m^2 s^{-1}$. The restoring time constant for surface temperature is 30 days. The large horizontal eddy viscosity is required to resolve the Munk-type western boundary layer because of the coarse resolution as in many large-scale ocean modeling studies. This choice is consistent with our main objective of this study: illustrating the role of convective schemes in thermohaline variability in a non-eddy resolving and hydrostatic ocean-only model.

The model was initialized with zero velocities, constant S = 35 psu and $\theta = 5$°C. It was first integrated for $t_{TS} = 8000$ years with an asynchronous time integration of $\Delta t_{uv} = 1$ h and $\Delta t_{TS} = 6$ days. At this time the model is well spun up, as confirmed by mean surface heat flux and basin mean kinetic energy. Asynchronous integration is a good approximation to a synchronous one when changes in temperature and salinity is small over one time step. However, when an ocean model is in a limit cycle state, there are often large changes in temperature and salinity during the strong phase of convection as we will see in the following. So the model was switched to a synchronous integration for a further 1100 years with $\Delta t_{uv} = \Delta t_{TS} = 6$ h. Note that the time step has been optimized to be very close to the limit that Courant–Friedrichs–Lewy criteria allow for an explicit time integration scheme.

Two experiments (A and B) were carried out. The YS scheme was used in A, and the IVD scheme in experiment B. Figure 3 shows basin mean surface heat flux (H_f) and basin mean specific kinetic energy (K_e) for both experiments during the last 200-year synchronous integration. As shown, while the model solution with either of these schemes can be characterized by interdecadal oscillations, the variability is different. The primary oscillation with both schemes has a period of about 11 years (obtained by a power spectral analysis, and the same method hereafter), but basin mean kinetic energy shows large differences. The 11-year cycle is modulated by a 35-year oscillation with the IVD scheme, but it is modulated by a 22-year cycle with a complete scheme. The amplitude of kinetic energy variation with the IVD scheme is also about twice as large as that with a complete adjustment scheme, presumably caused by persistent dynamic adjustment to the density field that has residual static instability with the IVD scheme. These interdecadal oscillations are a result of two competing horizontal advective processes associated with convection: subsurface warming and

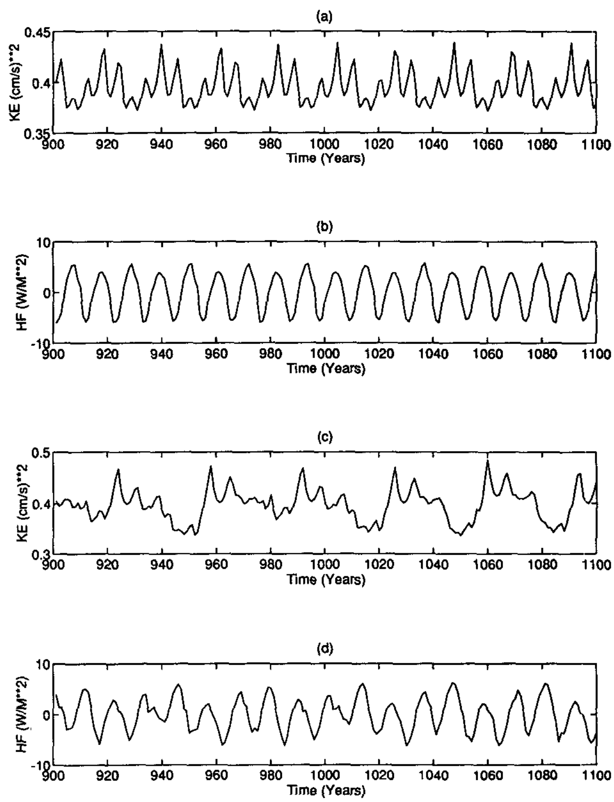

FIG. 3. Basin mean specific kinetic energy (E_k) ($cm^2 s^{-2}$) and mean surface heat flux (H_f) (W m^{-2}, positive represents downward) over the last 200-year synchronous integration. (a) E_k with the YS scheme. (b) H_f with the YS scheme. (c) E_k with the IVD scheme. (d) H_f with the IVD scheme.

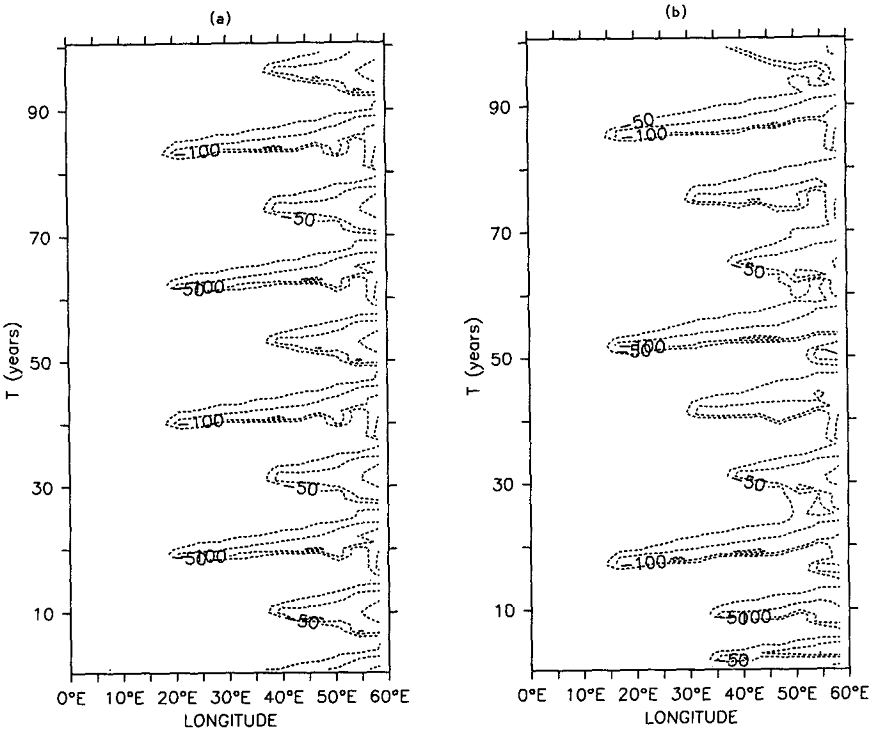

FIG. 4. Surface heat flux (W m^{-2}, negative represents upward) over the last 100-year synchronous integration in a longitudinal slab around 52°N where the largest variation occurs. (a) With the YS scheme. (b) With the IVD scheme. CI: 50 W m^{-2}, zero contour not drawn.

surface freshening in the subpolar region, and have been discussed in detail elsewhere (Yin and Sarachik 1994; Yin 1994). Here we only focus on the difference of these oscillations with the YS and IVD schemes in the model.

Figure 4 shows the surface heat flux over a 100-year integration in a slab around 52°N. This is the region where large variation of surface heat flux occurs during the oscillation in association with the east–west movement of convection during a cycle. As shown, both experiments have a similar primary oscillation of 11 years, which is mainly associated with the deep water formation [for details see Yin and Sarachik (1994)]. While this primary oscillation is modulated by a 35-year cycle with the IVD scheme, it is modulated by a 22-year cycle with the YS scheme. This difference reflects the effects of incomplete and complete convective adjustment schemes in the interdecadal thermohaline oscillation.

Figure 5 shows the zonally integrated meridional transport function at 2500 m. There is a localized cell counter (dashed lines) to the main thermohaline cell (solid lines) that periodically appears. The variation of transport function values in both experiments are similar, indicating that the difference between A and B is not associated with deep-water formation, but with intermediate-water formation in the interior of the ocean as shown in Fig. 4. Note that the only difference in experiments A and B is the convective adjustment schemes. While the density field that the velocity field adjusted to may contain residual static instability with the IVD scheme (an incomplete adjustment scheme), the YS scheme guarantees that the density field has no static instability. The difference in these oscillations with the two schemes is a result of the nonlinear coupling of dynamic and convective adjustments.

4. Summary and concluding remarks

The thermohaline overturning is the major circulation that transfers water property from the upper ocean into the deep ocean. Gravitational deep convection plays a dominant role in this process. Since convection occurs over a small scale, it is impossible to resolve in large-scale OGCMs and therefore has to be parameterized. Most large-scale ocean models use convective adjustment to remove gravitational static instability. The adjustment is carried out through vertical mixing with conservation of temperature and salinity. Major emphasis has been placed on the efficiency in removing the static instability.

We have devised an efficient convective scheme. Compared to the implicit vertical diffusion scheme (Cox 1984; Killworth 1989), this explicit and iterative one guarantees a complete removal of static instability of a water column at each time step, and it also takes

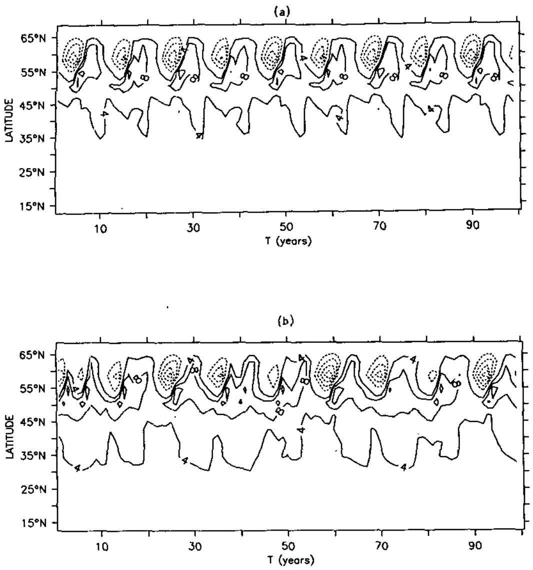

FIG. 5. Zonally integrated transport function (Sv $\equiv 10^6$ m^3 s^{-1}) at 2500 m over the last 100-year synchronous integration (solid lines represent the main cell of deep water formation and dashed lines correspond to the counter cell). (a) With the YS scheme. (b) With the IVD scheme.

about 10% less CPU time for a complete spinup of the model on a scalar machine. This scheme is based on the same physical consideration as the IVD scheme, and produces the same adjusted stable profile as the implicit one would were the IVD scheme run with enough iterations (about the number of vertical grid points).

The two schemes are compared in a model that is in a state of interdecadal limit cycles. While the model solution with either of these schemes can be characterized by interdecadal oscillations, the variability is different. The primary oscillation with both schemes has a period of about 11 years (obtained by a power spectral analysis), but basin mean kinetic energy shows large differences. The 11-year cycle is modulated by a 35-year oscillation with the IVD scheme, but by a 22 year one with the YS scheme. The amplitude of kinetic energy variation with the IVD scheme is also about twice as large as that with a complete adjustment scheme. Since the IVD scheme can suffer residual static instability at each time step, the density structure that the velocity fields adjusted to may be gravitationally unstable. This leads to the difference in these oscillations because of the nonlinear coupling of dynamic and convective adjustments. It is therefore suggested that complete and incomplete convective schemes can lead to different model variability when convective changes in temperature and salinity have a large variation over a short period of time.

Most existing parameterizations of convection in large-scale ocean models have employed rather simple physics since deep convection is not fully understood subject. Different parameterizations are mainly classified in terms of the efficiency in removing gravitational instability. If enough iterations are carried out, they are all physically equivalent in a sense that no gravitational instability exists in a vertical water column at each time step. While the M scheme is reported to be computationally less efficient than the IVD scheme, the present one is more efficient. So the scheme reported in this paper may be included in the GFDL ocean model as an efficient complete convective adjustment scheme. With more computational power available in the future, more sophisticated parameterizations may be incorporated into ocean models to improve simulations of the ocean.

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