Effects of Bottom Boundary Layer Parameterization on Reproducing Deep and Bottom Waters in a World Ocean Model

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ABSTRACT

A simple bottom boundary layer (BBL) model is developed to be incorporated into a $z$-coordinate medium resolution ocean general circulation model. In the BBL model, velocity is calculated from the pressure gradient, which is also calculated within the BBL. Preliminary experiments using an idealized basin model clearly document that, for reproducing realistic overflow/downslope flow, it is essential to adopt the horizontally distributed Rayleigh drag coefficient in the BBL model and also that, for avoiding warming of the abyssal ocean owing to unphysical strong flows created by the pressure gradient error along the steep slope, it is necessary to limit the area of the BBL. This BBL model is successfully incorporated into a World Ocean model with $1^\circ \times 1^\circ$ resolution, producing the overflow/downslope flow in the northern North Atlantic and around Antarctica. The dense overflow/downslope flow water provides the nucleus of the abyssal water in the World Ocean, leading to the better representation of the abyssal water. Thus, the incorporation of the BBL model can reduce the warming bias for the abyssal water in a coarse resolution model without modifying surface boundary conditions. In addition, it can also alleviate the model bias of too shallow extension of North Atlantic Deep Water.

Without the BBL model, the dense water from the Nordic seas does not flow southward and remains south of Iceland, forming a strong front there. The effect of Gent and McWilliams parameterization for mesoscale eddies to flatten isopycnal surfaces leads to unrealistic cooling south of Iceland. Unrealistic cooling also takes place around Antarctica. The incorporation of the BBL model drastically reduces this cooling by preventing the formation of the artificial front.

1. Introduction

In the present state of climate, most of the abyssal water originates from the two major water masses, that is, North Atlantic Deep Water (NADW) formed in the northern North Atlantic and Antarctic Bottom Water (AABW) around Antarctica. In the northern North Atlantic, a substantial amount of the dense water formed north of the Greenland–Scotland Ridge crosses over the ridge as overflow water and contributes significantly to the production of NADW (e.g., Dickson and Brown 1994). In the Weddell and Ross Seas, the dense shelf water formed on the continental shelf flows down to the ocean floor, increasing its volume through entrainment of overlying less dense waters, and it becomes AABW (e.g., Price and Baringer 1994).

In this way, the overflow/downslope flow from marginal seas and continental shelves provides the nucleus of the abyssal water in the World Ocean. But in state-of-the-art ocean models, the abyssal water is formed through a completely different process; that is, it is formed through open ocean convection at the polar rim of the open ocean. Accordingly, in the models, NADW is formed in the Irminger Basin and the Labrador Basin, and AABW is formed off the Weddell and Ross Sea, whereas dense waters in the marginal seas and on the shelves are isolated from the open ocean. Recent modeling studies show that not only the property of the abyssal water is modified through this unrealistic formation process but also adjustment to climate change is distorted severely (e.g., Döscher and Redler 1997; Wood et al. 1999).

There are two major reasons why the overflow/downslope flow is not represented properly in existing climate models. One is that coarse resolution models that do not resolve the Rossby radius cannot generate sufficient shear and eddy activity, which leads to an ageostrophic flow crossing the geostrophic contour. Accordingly, dense waters stay in marginal seas or on shelves. The other is that $z$-coordinate models cannot properly represent a flow along the ocean floor, even when the overflow/downslope flow is reproduced, because a dense water is artificially diluted through convective entrain-
ment when it falls down a cliff surrounded by less dense ambient waters (Winton et al. 1998).

It has been considered that these problems can be solved by incorporating a bottom boundary layer (BBL) parameterization (e.g., Hirst and McDougall 1996). Although the importance of BBL parameterization has been recognized for a long time, the incorporation of it into an ocean general circulation model (OGCM) started in the late 1990s (Beckmann and Döscher 1997).

Proposed BBL parameterizations can be classified into two types. One type does not use an extra bottom layer added to an OGCM, but it rearranges mass fluxes along the bottom. Beckmann and Döscher (1997) proposed to use advection in an OGCM, but to treat it as if mass fluxes were along the bottom when the bottom is not flat and also when a denser water is in the upper cell in order to represent the downflow slope. In that case large diffusion is also adopted for the bottom cell. Döscher and Beckmann (2000) applied this BBL parameterization to a coarse resolution model for the North Atlantic, showing that the incorporation of such a BBL parameterization can solve the long-standing model problem of too shallow extension of NADW. Deng et al. (1999) also applied this BBL parameterization to an eddy-permitting model for the northern North Atlantic (with a grid size of $\frac{1}{2} \times \frac{1}{2}$), demonstrating that the Denmark Strait overflow water has a pronounced influence on the water mass and tracer distribution in the subpolar North Atlantic. But their model was integrated only for 20 years, which is too short to set up the thermohaline circulation. Thus the effect of Beckmann and Döscher’s (1997) BBL parameterization on the abyssal circulation in an equilibrium state is left to be clarified. Campin and Goose (1999) proposed a simple parameterization of the downflow slope for z-coordinate coarse-resolution models. At the shelf break, when density on the shelf is higher than that in the neighboring deep water column, downwedge transport is imposed. Then they applied their parameterization to a coarse-resolution World Ocean model (with a grid size of $3^\circ \times 3^\circ$), showing some improvement in representing the downflow from the continental shelf around Antarctica.

The other type of parameterization explicitly represents the BBL by including it underneath the OGCM cells. Gnanadesikan (2001, submitted to J. Phys. Oceanogr., hereafter GNAN) proposed to compute horizontal flows and the pressure gradient for the BBL. Here the overflow/downslope flow is represented by the flow in the BBL, not by that computed in the OGCM cells. His scheme generates an ageostrophic flow crossing the geostrophic contour, succeeding in dragging the dense water from the marginal sea into the open ocean when the value for the Rayleigh drag coefficient is taken equal to the Coriolis parameter. Killworth and Edwards (1999) extended GNAN’s approach so as to enable the thickness of the BBL to vary temporally and spatially to represent entrainment and detrainment. This new approach is undoubtedly more realistic. However, its computational costs are too high to carry out a World Ocean experiment. Recently Song and Chao (2000) proposed a more elaborate BBL model in calculating the pressure gradient than Killworth and Edwards’ (1999).

In the present study, we develop a BBL model following GNAN. After examining the performance of the BBL model in an idealized basin model, we incorporated it into a medium-resolution World Ocean model (with a grid size of $1^\circ \times 1^\circ$). This is the first attempt to incorporate the BBL parameterization of the latter type into a World Ocean model.

This paper is organized as follows. The BBL model is described in section 2. Section 3 presents preliminary experiments for examining the performance of the BBL model. In section 4, this BBL model is incorporated into a world ocean model. Finally, discussion and conclusions are presented in section 5.

2. Description of BBL

a. Basic equations

The BBL is represented as a set of thin cells added to the bottom of the OGCM. The thickness of the BBL is taken constant. Vertically, the BBL cell interacts with the OGCM cell in the normal way, that is, through vertical diffusion, vertical advection, and convective adjustment when the density of the BBL cell is lighter than that of the OGCM cell above. But, horizontally, they interact only with the neighboring BBL cells, through horizontal advection.

The basic equations for the BBL model are

$$\frac{\partial u}{\partial t} - fu = -A(u) + \frac{uv \tan \theta}{a} - \frac{1}{\rho_0 a \cos \theta} \frac{\partial P}{\partial \lambda} + V(u) - au,$$

(1)

$$\frac{\partial v}{\partial t} + fu = -A(v) + \frac{u^2 \tan \theta}{a} - \frac{1}{\rho_0 a \cos \theta} \frac{\partial P}{\partial \theta} + V(v) - av,$$

(2)

$$A(1) = 0,$$

(3)

$$\frac{\partial T}{\partial t} = -A(T),$$

(4)

$$\frac{\partial S}{\partial t} = -A(S),$$

(5)

where $t$ is time, $\lambda$ and $\theta$ are the longitude and the latitude respectively, $u$ and $v$ are the zonal and meridional components of velocity, $P$ is pressure, $\rho_0$ is density of seawater, $a$ is the radius of the earth, and $V(\cdot)$ is the viscosity term. Here $A(\cdot)$ represents the advection term

$$A(\psi) = \frac{1}{a \cos \theta} \frac{\partial}{\partial \lambda} (u \psi) + \frac{1}{a \cos \theta} \frac{\partial}{\partial \theta} (v \psi \cos \theta)$$

$$+ \frac{\partial}{\partial z} (w \psi),$$

(6)
where $\psi$ is an arbitrary function. Pressure gradient terms in Eqs. (1) and (2) are calculated for the BBL cell. The parameter $\alpha$ is the Rayleigh drag coefficient, which is explained in detail in the next subsection.

When the depths of the neighboring BBL cells are different, the horizontal pressure gradient between them is calculated by bringing these cells to the mean depth of them. The basic concept is the same as that of $\sigma$-coordinate models. Specifically, it is expressed by the equation

$$\frac{\partial P}{\partial x} \bigg|_z = \frac{\partial}{\partial x} P(z = H(x)) + \frac{\partial P}{\partial z} \frac{\partial H}{\partial x},$$

where $P$ is pressure, $H$ is the depth of BBL, and $\rho$ is in situ density. The expression is based on GNAN. The additional computational cost for the incorporation of the BBL model roughly corresponds to that for increase in vertical resolution by one grid cell and is not a severe burden.

b. Parameters for the BBL model

1) BOTTOM STRESS TERM

For a coarse-resolution model, a dense water plume on the slope does not flow down, being isolated from the open ocean due to the geostrophic constraint. How does a dense water plume on the slope descend the slope when the Rossby radius and BBL are well resolved and mesoscale eddies are generated? Jiang and Garwood (1996) investigated three-dimensional features of a downslope flow of a dense water plume over the continental slope using an eddy-resolving $\sigma$-coordinate model. In their experiment, the plume is not constrained by geostrophy, flowing nearly 45° to the left of the geostrophic contour with vigorous eddy activity. When the plume reaches a certain depth, it starts to flow along the geostrophic contour. They show that this depth depends on combined effects of the initial overflow velocity, width, properties of a source water, planetary rotation, and slope steepness.

To break the geostrophic constraint without such eddy activity, Rayleigh drag is introduced into the BBL [Eqs. (1) and (2)] when the dense water is on the slope. When the value for the Rayleigh drag coefficient is nearly equal to the Coriolis parameter, $\alpha \approx |f|$, the dense water descends 45° to the left of the geostrophic contour. GNAN demonstrated that a Rayleigh drag coefficient of $\alpha \approx |f|$ is most efficient to generate flow that crosses the geostrophic contour: a smaller $\alpha$ cannot break the geostrophic constraint enough and a larger $\alpha$ results in a weak flow.

2) ADVECTION

GNAN used the upcurrent advection scheme to keep temperature and salinity of the BBL stable, but properties of the overflow/downslope flow water are modified quickly within a few grid points because the numerical diffusion caused by the upcurrent advection scheme is very large in the region of a strong horizontal flow, which corresponds to that of the overflow/downslope flow. We found that the use of a high-accuracy tracer advection scheme, Uniformly Third-Order Polynomial Interpolation Algorithm (UTOPIA: Leonard et al. 1993, 1994), that has third-order accuracy with little numerical diffusion and dispersion for the BBL model instead of an upcurrent advection scheme dramatically improves the problem, in particular, on the overflow around the Greenland–Scotland Ridge system.

3) HORIZONTAL DIFFUSION

Horizontal diffusion is usually applied to suppress numerical noise. However, this modifies properties of the overflow/downslope flow water. In addition, since the difference in temperature between the continental shelf and the ocean floor is large, even the small background horizontal diffusion coefficient ($\sim 10^2 \text{ m}^2 \text{ s}^{-1}$) brings about large diapycnal diffusion, resulting in noticeable warming of the abyssal ocean. This feature is demonstrated in section 3.

This is severe when the BBL is applied to the whole ocean floor. It has been known that there is a similar problem in a $\sigma$-coordinate model. For example, Ezer and Mellor (1997) integrated a $\sigma$-coordinate Atlantic model for 30 years, showing that weak Newtonian body forcing is needed to avoid the warming bias for the abyssal ocean owing to this diapycnal diffusion. We do not apply the horizontal diffusion among the BBL cells. Consequently, noises in the BBL cells are suppressed vertically.

4) OTHER PARAMETERS

There are other unknown parameters such as the vertical viscosity and the vertical diffusivity between the BBL and OGCM cells, and the thickness of the BBL. This indicates that the BBL model contains uncertainty as well as flexibility. These parameters, except for the thickness, are not important compared to the other parameters because the effect of advection dominates in the vertical mass exchange between the BBL and OGCM cells. We set the thickness to 100 m based on the results of a plume model of Jungclaus and Backhaus (1994), which shows thickness about 100 m for the Denmark Strait overflow. We will not be applying this BBL model to the whole of the world ocean to avoid unphysical flows created by pressure gradient error shown in the following section.
3. Preliminary experiments

a. Model description

We use the Center for Climate System Research (CCSR) Ocean Component model (COCO) described by Hasumi (2000). The governing equations are the primitive equations in spherical coordinates with the Boussinesq, hydrostatic, and rigid-lid approximations. COCO has isopycnal diffusion with weak background horizontal diffusion (Cox 1987). A high-accuracy tracer advection scheme, UTOPIA, is incorporated. The performance of COCO with UTOPIA is described by Hasumi and Suginohara (1999).

Before applying the BBL model to a World Ocean model, we perform preliminary experiments using an idealized basin model to examine its performance. We focus on the introduction of the BBL model to medium-resolution models, which marginally resolve the bottom topography around the important sills.

The model domain covers 20°S–70°N latitude, 0–60°E longitude. We consider two basins separated by a zonal ridge at 60°N (see Fig. 1). The northern and the southern basin, and the ridge between them, correspond to the Nordic Seas, the Atlantic Ocean, and the Greenland–Scotland Ridge, respectively. To be specific, they correspond to the Iceland Sea, the Irminger Sea, and Denmark Strait, although their scales are enlarged. The model grid spacing is 2° × 2° with 23 unequally spaced vertical levels. The maximum basin depth is 3500 m. This resolution can marginally represent this model topography.

The surface temperature is restored to the prescribed temperature, which is 25°C south of the equator and linearly decreasing to the north until 5°C at 60°N. The restoring time is 30 days. Wind stress and salinity flux are not applied. A constant salinity of 34.5 psu is used only to calculate density.

We carry out six experiments (Table 1). Each experiment is identical except for the treatment of the BBL model. Experiment I is a run without a BBL. Experiment II includes the BBL model without Rayleigh drag. Next, Rayleigh drag is applied for all BBL cells. In experiment III its coefficient is set to 10⁻² s⁻¹, and in experiment IV it is set to zero south of 45°N and at depths deeper than 3100 m. For experiment V the BBL model includes the horizontal diffusion; the coefficient is 10⁻³ m² s⁻¹. For experiment VI we apply the BBL model only to the north of 30°N.

The following values for the model parameters are used: $A_H = 2.5 \times 10^4$ m² s⁻¹, $A_V = 0.5 \times 10^{-4}$ m² s⁻¹, $K_H = 5.0 \times 10^{-2}$ m² s⁻¹, and $K_V = 0.5 \times 10^{-4}$ m² s⁻¹. Here $A_H$ and $A_V$ are the coefficients of horizontal and vertical viscosity, and $K_H$ and $K_V$ are those of horizontal and vertical diffusion. We use 100 m for the BBL thickness. The vertical viscosity and vertical diffusivity between the OGCM cells and the BBL are taken to be $1.0 \times 10^{-4}$ m² s⁻¹ and $5.0 \times 10^{-4}$ m² s⁻¹, respectively. Each model integration has been carried out for 4000 years to reach an equilibrium state.

b. Results

Figures 1 and 2 show the temperature and velocity fields in the OGCM bottom cells (expt I) and the BBL (expts II–V), respectively. When the BBL is not incorporated (expt I), the cold water is isolated to the north of the ridge (Fig. 1a), and a strong front is formed along the ridge. There is almost no flow crossing the ridge and the deep western boundary current appears south of 40°N (Fig. 2a). At the northwestern corner of the southern basin there is a weak signal of the cold water originating from the northern basin along the western boundary, but it hardly cools the southern basin as seen in the horizontally averaged temperature in the southern basin (Fig. 3). The zonally integrated streamfunction shows two circulation cells separated by the ridge (Fig. 4a). Note that, in contrast to rectangular basin cases, the core of the circulation cell in the southern basin is not attached to the ridge.

When the BBL model without Rayleigh drag is incorporated (expt II), the southern basin cools (Fig. 3). The cold water from the northern basin tends to cross the ridge along the western boundary (Figs. 1b and 2b). However, most of the cold water is trapped along the topography and it does not flow down to the bottom.

When Rayleigh drag with a coefficient of $10^{-4}$ s⁻¹ is applied throughout the basin (expt III), the southern basin cools significantly (Fig. 3). The overflow crosses the ridge and the cold water flows down to the deepest part of the ocean. There is no deep western boundary current south of 50°N. This is because Rayleigh drag works not only to break the geostrophic constraint along the ridge but also to damp the western boundary current.

This deficiency can be avoided when Rayleigh drag is applied only to the ridge area (expt IV). After reaching the prescribed depth where the Rayleigh drag coefficient is taken to be zero, the overflow water flows along the topography, forming the western boundary currents (Figs. 1d and 2d). In addition, there are other boundary currents along the topography, which are discussed below. The overturning circulation clearly shows that the incorporation of the BBL model having Rayleigh drag

<table>
<thead>
<tr>
<th>Expt</th>
<th>BBL region</th>
<th>$K_H$ (BBL)</th>
<th>$\alpha$</th>
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<tbody>
<tr>
<td>I</td>
<td>off</td>
<td>—</td>
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<tr>
<td>II</td>
<td>on</td>
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<tr>
<td>III</td>
<td>on</td>
<td>on</td>
<td>off</td>
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<tr>
<td>IV</td>
<td>on</td>
<td>on (depths ≲ 3100 m and north of 45°N)</td>
<td>off</td>
</tr>
<tr>
<td>V</td>
<td>same as IV</td>
<td>on</td>
<td>all</td>
</tr>
<tr>
<td>VI</td>
<td>same as IV</td>
<td>off</td>
<td>North of 30°N</td>
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leads to a vigorous water mass exchange between the northern and southern basins (Fig. 4b). The difference in overturning circulation between the experiments with and without the BBL model with Rayleigh drag indicates that the thermohaline circulation is enhanced both over the ridge and at the bottom (Fig. 4c).

Rayleigh drag should ideally be determined by the flow field itself. However, this is difficult to do unless downslope flow processes are well resolved in a model. In medium-resolution models we need to prescribe the Rayleigh drag coefficient as there are less than 10 grid points from the ridge to bottom.

When the BBL model includes horizontal diffusion (expt V), the overflow water is severely diffused (Fig. 1e). In addition, diffusion between the abyssal cold and shallow warm waters leads to warming of the southern basin as for \(\sigma\)-coordinate models (Fig. 3). Consequently, horizontal diffusion should be taken as small as possible in the BBL model.

In experiment IV, in addition to the southward deep
western boundary current south of the ridge, there are extremely strong boundary currents along the topography at low latitudes. They are created by the pressure gradient error. When the stratification is horizontally uniform, the pressure gradient between the BBL at the different depth should be zero. This means that the two terms on the right-hand side of Eq. (7) cancel each other. However, for the centered finite differencing of the two terms the cancellation does not occur unless $P$ and $\rho$ are the linear function of $z$, leading to the artificial pressure gradient. Although there are some methods to reduce this pressure gradient error (e.g., Song and Chao 2000), it is difficult to completely suppress it without increasing model resolution. The error is serious at low latitudes where the Coriolis parameter is small: even a small pressure gradient error causes strong unphysical flows. Consequently, the southern basin is warmed: the unphysical flows enhance the diapycnal fluxes between shallow warm and deep cold waters as they are not along isopycnals. However, the BBL is thought to be unim-

Fig. 2. Horizontal velocity field at the lowest levels and topography for expts (a) I, (b) II, (c) III, (d) IV, and (e) V. The lowest levels are the OGCM bottom cells for (a), and they are the BBL for (b)–(e). Contour intervals for the topography are 300 m.
important to global water masses except for the overflow region. In the overflow region, pressure on the northern side of the ridge is much greater than that on the southern side, and hence the direction of the pressure gradient can be estimated correctly. Therefore, it seems better to limit the area of BBL only to the northern part of the southern basin.

The overflow for experiment VI is quite similar to that for experiment IV. Figure 5 shows the temperature and the velocity field in the OGCM bottom cells and the BBL for experiment VI. The BBL flows are not distorted severely where the BBL ends. They simply flow upward into the OGCM cells and continue to flow southward. By suppressing the unphysical flows in the southern basin, the horizontally averaged temperature of experiment VI is lower than that of experiment IV (Fig. 3). Since the topography of the World Ocean is much more complicated and the ratio of the deep-water formation region to the upwelling region are much smaller than that of the ideal basin model, this warming effect becomes much greater for the World Ocean model. Consequently, we conclude that the BBL model with

![Fig. 3. Horizontally averaged temperature in the southern basin (20°S–60°N).](image)

![Fig. 4. Zonally integrated transport streamfunction for expts (a) I, (b) IV, and (c) IV – I. Contour intervals are 1 Sv for (a) and (b), and 0.5 Sv for (c).](image)
the horizontally distributed Rayleigh drag coefficient should be applied only to the location where the effects of the BBL are indispensable.

4. World Ocean model experiments
   a. Model description

   For the World Ocean model, the horizontal resolution is $1^\circ \times 1^\circ$ and there are 40 levels in the vertical with high resolution near the sea surface (50 m) and decreasing resolution toward the bottom (200 m). This resolution is relatively high for a climate model, which needs to be integrated for several thousand years in order to obtain an equilibrium state. The topography data for the model are taken from a 5-min earth topography (ETOPO5). First, ETOPO5 data are interpolated to the model grid, and then lightly smoothed. In the present $1^\circ \times 1^\circ$ grid model, the presentation of basins, which divide an
ocean into pieces, is significantly improved compared to a $3 \times 3$ grid model. This leads to better representation of the abyssal circulation. Furthermore, the $1^\circ \times 1^\circ$ resolution may yield the coarsest limit for incorporating a BBL model into a World Ocean model. For example, when a $3^\circ \times 3^\circ$ resolution is adopted, the continental shelf around Antarctica is represented by only one or two grid points and Denmark Strait is nearly closed unless Iceland is removed. Although the basins are properly represented in the present model, paths and straits that connect basins cannot be resolved. We dredge some important paths and straits to enable flow to pass through them. For example, Denmark Strait is set to 830 m and the sill east of Iceland is set to 940 m. It should be remarked that modification, which is common among OGCMs, is necessary even for eddy-permitting and eddy-resolving models. For example, Redler and Bönning (1997) pointed out that the World Ocean Circulation Experiment (WOCE) Community Modeling Effort (CME) model, whose resolution is $\frac{1}{3}^\circ \times \frac{1}{3}^\circ$, cannot resolve the Faroe Strait adequately without this modification and demonstrate that insufficient representation of mass flux across the strait has a significant influence on the flow pattern of North Atlantic Current.

The Arctic Sea is excluded for computational efficiency. The Arctic Sea indeed plays a critical role in forming the dense water in the Nordic seas (Mauritzen 1996), but the formation process, which involves sea ice, convection, and river runoff, may be too complicated to be reproduced properly. We do not predict the water in the northern Nordic seas but prescribe it (see below).

For the parameterization of mesoscale eddies, we use that of Gent and McWilliams (1990) and Gent et al. (1995) (hereafter GM parameterization). It is reported that the incorporation of GM parameterization leads to improvement in reproducing the abyssal water (e.g., England and Rahmstorf 1999). Its coefficient is usually set to the order of $10^4$ m$^2$ s$^{-1}$ (e.g., Danabasoglu and McWilliams 1995; Hirst and McDougall 1996; England and Rahmstorf 1999). Recently Visbeck et al. (1997) pointed out that in a coarse-resolution model with GM parameterization, the thickness diffusivity that best matches an eddy-resolving model differs among situations such as gyre, channel, and open ocean convection. There are several experiments that use the spatially variable thickness diffusivity of Visbeck et al. (1997) (Gordon et al. 2000; Wright 1997). However, their integration times are only a few decades, and their effects to an equilibrium state is not clear yet. We use the constant value of $5.0 \times 10^2$ m$^2$ s$^{-1}$, which is one of the best choices in England and Rahmstorf (1999).

Based on the result obtained in the preliminary experiments, we apply the BBL only north of 49°N in the North Atlantic and south of 54°S in the Southern Ocean. Next, we need to determine the depth for the transition from a downslope flow to a geostrophic flow, imposing $\alpha = |f|$, $\alpha = 10^{-4}$ s$^{-1}$ above it and $\alpha = 0$ below it. After some test runs, we decided to prescribe the constant value of 2000 m in the northern North Atlantic, resulting in a realistic flow overflow around the Greenland–Scotland Ridge system. Around Antarctica, the depth is set to 4000 m. These values roughly correspond to the base of the continental slope.

The following values are used for the model parameters: $A_H = 2.5 \times 10^4$ m$^2$ s$^{-1}$, $A_v = 1.0 \times 10^{-4}$ m$^2$ s$^{-1}$, $K_I = 1.5 \times 10^3$ m$^2$ s$^{-1}$, and $K_I' = 1.0 \times 10^2$ m$^2$ s$^{-1}$. Here $K_I$ is the coefficient of isopycnal diffusion. For the vertical diffusivity, $K_I'$, we adopt Tsujino et al. (2000) type, which takes the value of $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ in the upper thermocline, $1.0 \times 10^{-4}$ m$^2$ s$^{-1}$ in the lower thermocline, and a larger value of $2.7 \times 10^{-4}$ m$^2$ s$^{-1}$ in the bottom layer. We use 100 m for the BBL thickness. The vertical viscosity and vertical diffusivity between the OGCM cells and the BBL are $1.0 \times 10^{-4}$ m$^2$ s$^{-1}$ and $5.0 \times 10^{-4}$ m$^2$ s$^{-1}$, respectively.

**Forcings**

The surface forcings are seasonally varying. The wind stress is taken from Hellerman and Rosenstein’s (1983) monthly dataset. Following Haney (1971), the sea surface temperature is restored to “apparent atmosphere temperature,” which is derived from longwave radiation, shortwave radiation, and latent heat flux from da Silva et al. (1994). Salinity flux, which corresponds to evaporation minus precipitation $(E - P)$ from da Silva et al. (1994), is imposed. In addition, the sea surface salinity is weakly restored to the monthly mean value of Levitus et al. (1994) with a damping time of 200 days to avoid climate drift. For time interpolation of these sea surface data, the method of Killworth (1996) is used so that the values imposed on the model equal to those of the original monthly datasets when averaged over a month.

It is the dense water in the Nordic Seas that overflows the Greenland–Scotland Ridge and becomes the nucleus of NADW. The dense water is formed not only by convection in the Nordic Seas, but also through complicated water mass conversion in the Nordic Seas and the Arctic (Mauritzen 1996). In the present model, the temperature and salinity in the northern Nordic Seas (north of 75°N and below 1500 m) are restored to $-1.2^\circ$C and 35.1 psu.

The surface forcings in winter on the continental shelf around Antarctica are also modified to include the effects of catabatic winds and brine ejection. The modification is limited to the shelf region so as to avoid unphysical deep convection and heat flux. Thus, this model does not produce Toggweiler and Samuels’ (1995) spurious salinity enhancement.

**b. Experimental procedures**

We carried out two experiments that are identical except for the incorporation of the BBL model: experiment
ABBL has the BBL model and experiment NOBBL does not. Following the method of Danabasoglu et al. (1996), we obtained an equilibrium state by carrying out more than ten thousand years of accelerated integration followed by 30 years of synchronous integration. The accuracy of this method is assessed by Nakano et al. (1999). All results presented in this study are the annual mean for the last year of synchronous integration. For reference, bathymetric features referred to in the text are plotted in Figs. 6 and 7 for the northern North Atlantic and for the vicinity of the Ross Sea.

c. Results

1) GENERAL FEATURES OF THE MODEL WITH THE BBL

The meridional overturn for ABBL (the experiment with the BBL model) is shown in Fig. 8. In the Atlantic, the overturning circulation indicates NADW flowing southward from the Nordic Seas and AABW flowing northward from the Southern Ocean. Details are discussed when we compare the two experiments in section 4c(3). In the deep Pacific, a layered structure in meridional overturn is formed, which is composed of the northward transport at the bottom and the southward transport at middepths (e.g., Obata et al. 1996; Tsujino et al. 2000). Details are discussed by Nakano and Suginohara (2001, manuscript submitted to *J. Geophys. Res.*).

The zonally averaged temperature and salinity for ABBL are shown in Figs. 9 and 10 together with Levitus et al. (1994) climatology. Major characteristic features seen in the observation are well reproduced in the model, that is, NADW and AABW in the Atlantic and Antarctic Intermediate Water.

2) LOCAL EFFECT OF OVERFLOW/DOWNSLOPE FLOW

(i) Northern North Atlantic

Figure 11 shows the horizontal velocity and the density field in the BBL (OGCM bottom cells) in the northern North Atlantic for ABBL (NOBBL). For ABBL, the dense water in the Nordic Seas crosses over the Iceland–Scotland Ridge, descends along the eastern flank of the Reykjanes Ridge, enters the Irminger Basin passing through the Charlie-Gibbs Fracture Zone, and merges with the overflow water from Denmark Strait. Then it flows into the Labrador Basin as the deep western boundary current (DWBC). This flow pattern agrees with that of Dickson and Brown (1994). The overflow water changes its properties owing to strong entrainment of the overlying warm water. The entrainment takes place as the vertical compensation flow for the divergence of horizontal flow in the region where the gradient of the slope and density is large.

On the other hand, for NOBBL, the dense water from the Nordic Seas cannot cross over the Iceland–Scotland Ridge owing to the geostrophic constraint: it adheres to the topography south of Iceland and induces a strong density front there. The dense water does not flow as a DWBC at the bottom. In addition, the water along the eastern flank of Greenland is less dense than waters surrounding it. Accordingly, there is no remnant of the overflow in the Irminger Basin and the Iceland Basin.

The flow pattern obtained for ABBL cannot be reproduced when the Rayleigh drag coefficient, \( \alpha \), is constant everywhere. When \( \alpha = |f| \) everywhere, the overflow water that crosses over the Iceland–Scotland Ridge does not pass through the Charlie-Gibbs Fracture Zone but flows southeastward (not shown). When \( \alpha = 0 \) ev-
erywhere, the dense water does not cross over the ridge because it cannot break the geostrophic constraint (not shown). When $\alpha = |f|$ only at the Iceland–Scotland Ridge and $\alpha = 0$ at the Reykjanes Ridge, the overflow does not enter the Irminger Basin passing through the Charlie–Gibbs Fracture Zone but it flows over the Reykjanes Ridge whose depth is equal to that of the Iceland–Scotland Ridge. A flow pattern similar to that of Dickson and Brown (1994) is obtained only when $\alpha = |f|$ both at the Iceland–Scotland Ridge and in the shallow part of the Reykjanes Ridge.

Although the model results show basically good agreement with observations, there are some differences. In the model, the Iceland–Scotland overflow enters the Iceland Basin just east of Iceland where in the observations the main stream of the overflow passes through Faroe Strait.

(ii) Weddell Sea

The densest AABW is formed on the Filchner Ice Shelf at the southern end of the Weddell Sea. It descends northward along the continental shelf off the Antarctic Peninsula and reaches the bottom north of the Weddell Sea (Foster and Carmack 1976). Then it flows eastward to the Sandwich Trough ($\sim 58^\circ$S, $25^\circ$W) along the ridge. This feature is detected by tracing the high chlorofluorocarbon concentration ($\geq 0.3$) (Fig. 7 of Orsi et al. 1999).

Figure 12 shows the horizontal velocity and density fields in the BBL (OGCM bottom cells) around the Weddell Sea for ABBL (NOBBL). For ABBL, the dense downslope flow water descends northeastward along the continental shelf off the Antarctic Peninsula and reaches the bottom. This downslope flow water flows eastward to the Sandwich Trough along the northern rim of the Weddell Sea. On the other hand, for NOBBL there is no such dense downslope flow water and the densest water remains on the continental shelf. Also, flows are along the topography for NOBBL.

(iii) Ross Sea

The dense bottom water formed in the Ross Sea is prevented from going farther north to the Southeast Indian Ridge and the Pacific Antarctic Ridge. Accordingly, it does not have a direct relation with Circumpolar Deep Water (CDW) coming from the west: CDW from the west flows northward east of New Zealand, while the bottom water from the Ross Sea enters only the Amundsen Abyssal Plain (e.g., Mantyla and Reid 1983; Orsi et al. 1999). The bottom water from the Ross Sea is characterized by its high salinity. On the continental shelf in the Ross Sea, the western high salinity ($\geq 37.4$ psu) presents a striking contrast to the eastern one ($\leq 36.2$ psu) whereas there is no difference in potential temperature ($\leq -0.8^\circ$C).

Figure 13 shows the horizontal velocity and the density field in the BBL (OGCM bottom cells) around the Ross Sea for ABBL (NOBBL). For ABBL, the dense (saline) bottom water formed on the western continental shelf flows northward. When the bottom water reaches the ridge, it turns eastward and enters the Amundsen Abyssal Plain. This pattern is similar to that derived
Fig. 9. Zonally averaged temperature in the (a) Atlantic, (b) Pacific, and (c) Indian Oceans. Figures on the left are for Levitus et al. (1994) data (interpolated to the model grid) and those on the right for ABBL. Contour intervals are 1°C. The Atlantic for the Southern Ocean is defined as the area between Drake Passage and Cape of Good Hope, the Indian Ocean between Cape of Good Hope and south of New Zealand, and the Pacific between south of New Zealand and Drake Passage.

3) INFLUENCE OF THE OVERFLOW/DOWNSLOPE FLOW ON THE WORLD OCEAN

The difference in zonally averaged potential density ($\sigma_z$) between ABBL and NOBBL is shown in Fig. 14. Density of the abyssal water for ABBL is greater than that for NOBBL because the dense overflow/downslope flow water is produced for ABBL. Large positive differences are found along the path of the lower NADW in the Atlantic and at the bottom in the Southern Ocean. In the Atlantic, a large negative difference is seen around 63°N at the depth of 1000 m. This is related to the formation of the artificial, strong density front south of Iceland for NOBBL (Fig. 11), which is discussed in the next subsection. State-of-the-art coarse resolution models in general have warming bias for the abyssal water. The incorporation of the BBL model can reduce this bias without modifying boundary conditions.

from the CFC distribution (Fig. 7 of Orsi et al. 1999). In addition, we can see that the dense water extends northwestward from the shelf at about 150°E; that is, the bottom water formed off Adélie Land flows northwestward. For NOBBL, on the other hand, the densest water remains on the continental shelf. There is no bottom-water formation off Adélie Land, and the bottom water in the Australian–Antarctic Basin originates from the west. It is worth mentioning that linkage between the bottom water of the Australian–Antarctic Basin and the Southwest Pacific Basin has recently been proposed based on CFC observations from WOCE (Orsi and Bullister 1996). Thus the bottom water formed off Adélie Land is important not only for the deep eastern Indian (Foster 1995; Fukamachi et al. 2000), but also for the abyssal Pacific. Further details are discussed by Nakano and Suginohara (2001, manuscript submitted to J. Geophys. Res.).
Figure 15 shows the difference in meridional overturn between ABBL and NOBBL in the Atlantic. When the BBL model is incorporated, the overflow induces a circulation cell confined to the ocean floor south of the Greenland–Scotland Ridge, leading to the deeper NADW circulation. This does not mean enhancement of the NADW circulation because the center of the NADW circulation cell is located at 1500 m (see Fig. 8a). However, the BBL model works well to alleviate the model bias of too shallow extension of NADW. As Hirst and McDougall (1998) pointed out, GM parameterization on its own does not work because the GM parameterization cannot transport the dense water to the bottom even though it drags the dense water from the shelf and the marginal sea to the open ocean. Thus, only the incorporation of the BBL model can solve the long-standing problem of too shallow extension of NADW.

The northward heat transport is almost the same for the experiments with and without a BBL: 1.583 PW for ABBL and 1.597 PW for NOBBL at 24°N. This is because the heat transport is closely related to the maximum value of the streamfunction associated with the NADW circulation, that is, it slightly decreases at this latitude when the BBL model is incorporated. The cooling of the abyssal layer is not as effective for increasing the heat transport as the enhancement of the NADW transport. The introduction of the BBL model leads to the temperature decrease of ~0.5°C. This difference cannot be neglected for the abyssal water, but this is still very small compared to the difference between the abyssal and upper water (~20°C).

4) EFFECTS OF BBL ON GM PARAMETERIZATION

Many studies have been carried out to examine the sensitivity of the abyssal circulation to the GM parameterization (e.g., Danabasoglu and McWilliams 1995; Hirst and McDougall 1996; England and Rahmstorf...
Fig. 11. Properties in the BBL for ABBL (left panels) and the OGCM bottom cells for NOBBL (expt without the BBL model; right panels) in the northern North Atlantic: (a) horizontal velocity fields and topography, (b) potential density ($\sigma_z$). Counter intervals are 500 m for (a) and 0.02 $\sigma_z$ for (b). In (b), shading areas are for density greater than 37.05 $\sigma_z$.

Fig. 12. As in Fig. 11 but around the Weddell Sea.
Until now, the sensitivity of the abyssal circulation to GM parameterization has been examined in models that do not reproduce the overflow/downslope flow. Here, we carry out other experiments without the GM parameterization (ABBL $-\text{NOGM}$ and NOBBL $-\text{NOGM}$) and examine this sensitivity in the model where the abyssal water is realistically formed through the overflow/downslope flow.

It is known that the introduction of the GM parameterization leads to upward bowing of the thermocline because GM parameterization acts to flatten steep isopycnal surfaces (Gent et al. 1995). This effect is significant in particular at the subantarctic front, resulting in the northward intrusion of the dense Circumpolar Deep Water CDW and hence more stable stratification that is observed. In a typical coarse resolution model, unrealistic deep convection tends to take place at the subantarctic front, leading to a warm and less saline bias in the deep ocean. GM parameterization diminishes this deficiency thanks to the stable stratification. This effect is thought to represent effects of vigorous eddy activity in the circumpolar region (e.g., Danabasoglu and McWilliams 1995). This is the improvement common to previous studies.

Figure 16 shows the difference in potential density ($\sigma_z$) near the bottom in the northern North Atlantic between the experiments with and without GM parameterization for the model with and without the BBL, respectively. When the BBL model is not incorporated, GM parameterization leads to a significant increase in density south of Iceland. This density increase is much larger than that in the circumpolar region. And the large density increase extends southward along the path of NADW. On the other hand, when the BBL model is incorporated, the artificial increase south of Iceland is significantly reduced.

As was shown in Fig. 11 for NOBBL with GM parameterization, the dense water from the Nordic Seas does not flow southward along the Reykjanes Ridge but remains south of Iceland, creating the strong density front there. The flattening of isopycnal surfaces by GM parameterization acts on this artificial density front, resulting in unrealistic cooling (Figs. 14 and 16). This unrealistic cooling takes place also in the Weddell and Ross Seas. Thus, GM parameterization plays an unphysical role when the BBL model is not incorporated although it is favorable for reproducing the observed features.

5. Discussion and conclusions

We have developed a bottom boundary layer model, in which velocity is calculated from the pressure gra-
Fig. 14. Difference in zonally averaged potential density ($\sigma_z$) between ABBL and NOBBL in the (a) Atlantic, (b) Pacific, and (c) Indian Oceans. Contour intervals are 0.01 $\sigma_z$. Shaded areas indicate positive values.

Fig. 15. Difference in zonally integrated transport streamfunction between ABBL and NOBBL in the Atlantic. Contour intervals are 0.5 Sv.

The gradient calculated within the BBL, and incorporated it into a World Ocean model with 1° × 1° resolution. The incorporation of the BBL model succeeds in producing the overflow/downslope flow in the northern North Atlantic and around Antarctica. The dense overflow/downslope flow provides the nucleus of the abyssal water in the World Ocean, yielding better representation of the abyssal water. Thus, without modifying surface boundary conditions, the incorporation of the BBL model can reduce the warming bias for the abyssal water in state-of-the-art coarse-resolution models. In addition, it also solves the long-standing model problem of too shallow extension of NADW.

The GM parameterization plays an unphysical role when the BBL model is not incorporated. Without a BBL, the dense water from the Nordic Seas does not flow southward along the Reykjanes Ridge but remains south of Iceland, bringing about a strong density front there. The GM parameterization flattens this artificial density front, creating unrealistic cooling. The cooling also occurs in the Weddell and Ross Seas. The incorporation of the BBL model drastically reduces this cooling by preventing the formation of an artificial front.

To reproduce a realistic overflow/downslope flow, we adopted a horizontally distributed Rayleigh drag coefficient in the BBL model: it is nearly equal to the Coriolis parameter where the dense water plume flows down to the bottom and is zero where it flows along the geostrophic contour. In addition, to reduce the warming of the abyssal ocean owing to unphysical flows created by the pressure gradient error along the steep slope, it is necessary to limit the area of the BBL. These features are well documented in sensitivity experiments using an idealized basin ocean model. For the World Ocean model, we applied the BBL model only north of 49°N in the North Atlantic and south of 54°S in the Southern Ocean. To successfully apply the BBL model to a World Ocean model, its resolution also needs to be fine enough to represent topographic features of the abyssal water formation regions such as the northern North Atlantic and the adjacent seas around Antarctica. Our experience tells one that the resolution of 1° × 1° may be the coarsest limit for incorporating a BBL model into a world ocean model.

Although this treatment of the BBL is very simple, a realistic pattern of the overflow/downslope flow has been obtained. Better representation of the overflow/downslope flow is needed for simulating the abyssal circulation. In addition, the additional computational cost for incorporating the BBL model roughly corresponds to that for an increase in vertical resolution by one grid cell, thus is not a severe burden. Therefore, this parameterization is useful for climate studies that cannot explicitly include the physical BBL process. Of course, in the future, more complicated BBL processes will be incorporated, such as Killworth and Edwards (1999) and Song and Chao (2000) proposed. And it is
also necessary to develop a BBL model which is applicable for all of the World Ocean.

The GM parameterization is not the only parameterization for mesoscale eddies. Recently, other parameterizations for eddy-induced transport have been proposed that may be more physically justifiable (Treguier 1999; Killworth 1998). Although they have not been incorporated into OGCMs yet, these parameterizations will also flatten isopycnal surfaces. Since this flattening acutely affects the abyssal circulation by restricting overflow/downslope flow, the incorporation of a BBL parameterization is also necessary.

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