Effects of a Bottom Boundary Layer Parameterization in a Coarse-Resolution Model of the North Atlantic Ocean

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ABSTRACT

The bottom boundary layer model approach of Beckmann and Döscher has been adopted for application in a coarse-resolution model of the North Atlantic Ocean. Both components of the approach (advective and conditional diffusive) are found to affect the deep water stratification and circulation. A significant deepening of the downward spreading North Atlantic Deep Water (NADW) is the major effect. This is associated with an enhanced spatial coverage of the NADW cell in the meridional circulation.

1. Introduction

The thermohaline circulation of the global ocean plays a major role in the earth's climate system. The large-scale overturning circulation in single ocean basins is not only driven locally from the surface, but also by interbasin throughflows and overflows through narrow and shallow passages, which control the exchange of dense bottom water. One of the most important arteries of dense water flow, the Denmark Strait overflow, for example, is 20–40 km wide and only 200 m "thick" in places. This dense overflow is directly connected to deep boundary currents and the conditions of deep water formation at sites downstream of the overflow. Model studies in the framework of the World Ocean Circulation Experiment–community modeling effort (WOCE–CME) have revealed the critical importance of the North Atlantic overflow water masses for driving the thermohaline circulation (Döscher and Redler 1997). In comparison, the effect of subpolar deep water formation south of the North Atlantic sill is of minor importance in the model (Böning et al. 1996). It is concluded that ocean models should represent dense downslope flow properly in order not to underestimate the consequences for the large-scale overturning circulation.

However, many ocean models are characterized by major deficiencies concerning flux across topographic obstacles. This is particularly true in cases of coarse resolution or geopotential coordinate system. The latter prevents the preservation of tracer concentrations during downslope transports of dense water masses.

In the case of overflow water masses on a sill, they cannot spread along their isopycnals. Instead, tracer transport in an overflow situation is accompanied by strong vertical mixing and a loss of water mass and tracer properties. This occurs when dense water crosses a topographic step in the horizontal direction. Due to static instability, it is involved in a convective homogenization instead of being transported downslope. This deleterious effect of artificial dilution has been investigated recently by Winton et al. (1998).

The artificial dilution can be prevented by either resolving the viscous bottom layer or implementing a bottom boundary layer (BBL) model. The first possibility is unsuitable for large-scale studies because of excessive resolution requirements. An example of the latter has recently been suggested by Beckmann and Döscher (1997, hereafter BD97). Their BBL model is based on a hybrid approach, coupling a terrain-following bottom boundary layer to the standard ocean general circulation model (OGCM). Communication between bottom boxes is enabled by diffusion and advection. The advective velocities are inferred from the OGCM. This approach combines some of the benefits of a σ-coordinate model (bottom following) with the simplicity of a geopotential coordinate model. The basic concept of combining bottom boxes has been extended by Killworth and Edwards (1999). They propose a bottom layer of varying thickness inside the bottom boxes, which is diagnosed from local dynamical properties by empirical formulas. Gnanadesikan (2000, manuscript submitted to J. Phys. Oceanogr.) proposes a bottom layer of constant thick-
ness with a large number of adjustable parameters. Both works utilize an extra momentum equation for the bottom layer and have to take measures to keep the numerical pressure gradient error (as known from \( \sigma \)-coordinate models; see, e.g., Beckmann and Haidvogel 1993) as small as possible. This error tends to create artificial currents, especially over steep slopes, that is, at sites where the BBL is needed most. The simpler approach of BD97 avoids this error by not solving a momentum equation for the BBL.

In an idealized configuration, BD97 have demonstrated a significant improvement in spreading of a density anomaly over a linearly sloping bottom compared to the reference case of a pure steplike bottom. Their results give patterns similar to the plume model of Jungclaus and Backhaus (1994). The method allows a freely adjustable combination of advective and diffusive bottom-following transports not disturbed by the steps of topography in a geopotential coordinate GCM.

As a next step, we present an application of the BD97 method under more realistic conditions in a North Atlantic model of GFDL type, thereby choosing a configuration typical for the oceanic components of climate models. As an adequate representation of North Atlantic overflow is crucial for the large-scale circulation, a distinct effect can be expected. In order to make the method operational, several combinations of BBL advection and diffusion are tested with respect to tracer preservation in downslope flows and consequences for the overturning circulation. In the more realistic and more complex environment of the North Atlantic, some modifications to the basic method were found necessary.

2. The model

For the purpose of demonstrating the effects of the BD97 bottom layer approach, a North Atlantic realization of the Geophysical Fluid Dynamics Laboratory (GFDL) ocean primitive equation model MOM2 (Pacanowsky et al. 1993) has been chosen because the North Atlantic has one of the most noticeable overflows in the Denmark Strait and the GDFL-type models are widely used and provide the basis for most climate studies. A horizontal resolution of 1.5° aims at the next generation of climate models.

The model domain encompasses 20°S–75°N. The gridpoint distance is 1.5° in latitude and longitude. The vertical is discretized in 21 levels of increasing thickness (25–447 m). The topography has been inferred from the Scripps 1° topography by area averaging, which is the standard technique built into MOM2. As a result, the Denmark Strait was nearly closed and needed some extra broadening. This hand-tuning is commonly done in many models of the North Atlantic. Isopycnal diffusion is enabled. Thereby a small horizontal diffusivity remains to prevent numerical problems (Cox 1987). (The model parameters are summarized in Table 2.) The model starts from a resting state of temperatures and salinities given by the annual mean climatology of Levitus (1982). It is driven by annual mean wind stresses from Hellermann and Rosenstein (1983) and by a surface restoring of potential temperature and salinity to Levitus values. The restoring timescales of 50 days enables the model to follow the annual cycle. Convective mixing is implemented as full convection, as described by Rahmstorf (1993). Because of the restricted model domain, zonal slabs of 7.5° (south) and 6° (north) latitude are used as restoring zones at the boundaries. The restoring time varies between 3 (outer wall) days and 30 days. The northern restoring zone is completely located north of the North Atlantic sills, therefore it keeps the Nordic Seas at realistic mean temperatures and salinities but does not drive the meridional circulation of the Atlantic.

a. The BBL treatment

The GCM is supplemented by the bottom boundary layer approach of Beckmann and Dösch (1997). The primary goal of this parameterization is to prevent the deleterious dilution of descending dense water masses, which occurs near the bottom in \( \sigma \)-coordinate models by vertical convective homogenization after static stabilities. This method employs a coupling between a simple terrain-following bottom boundary layer and a \( z \)-coordinate model (Fig. 1). The bottom boundary layer consists of all the boxes at the bottom of the GCM basin (see also Fig. 2 of Beckmann and Dösch 1997), which are directly connected to their neighbors.

In BD97 two components of transport in a bottom boundary layer are described: an advective and a diffusive contribution. The diffusive part of the BBL is named “slope convection” and represents a conditional diffusion. As pointed out by BD97, an unrestricted application of this bottom-following diffusion would result in an overall smoothing of tracer fields within the bottom layer. The conditional scheme parameterizes the effect of a downslope tracer flow in a GCM with steplike
topography by applying a diffusion between bottom cells only if the bottom is not flat and if the upper density is higher than the lower [Eq. (1)]. Therefore, we impose a constraint on the Laplacian diffusion coefficient $A^\sigma$: \begin{equation}
A^\sigma(x, y, t) = \begin{cases} 
A^\sigma_{\text{max}} & \text{if } \nabla \rho > 0 < 0 \\
A^\sigma_{\text{min}} & \text{else} 
\end{cases} \tag{1}
\end{equation}
of the tracer diffusion equation in the BBL, \begin{equation}
D^\sigma = \nabla A^\sigma \nabla \rho, \tag{2}
\end{equation}
($\rho_b$ symbolizes potential density in the bottom tracer box, $\sigma$ indicates the bottom-following coordinate, $h$ is depth), which affects the tracer fields in addition to the diffusion of the GCM. An appropriate diffusivity is determined from the sensitivity studies in section 3. A small minimum diffusivity appeared to be necessary to prevent numerical instabilities.

For advection, only tracer tendencies (not momentum) are calculated within the boundary layer. The advective velocities are inferred from the level-coordinate GCM. This approach prevents numerical problems concerning the calculation of the horizontal pressure gradient. The most important mechanism of this technique is to prevent convective vertical homogenization in the near-bottom tracer field.

b. Modifications to the BBL scheme

Both methods (advection and conditional diffusion) have been successfully applied in idealized configurations with linearly sloping bottom (BD97). Here, we apply the BBL to a more realistic GCM of the North Atlantic. When using the original BD97 formulation, we encountered a potential problem: the downward advection of a less dense upper layer water to deeper layers. This occurs at the boundary of the Carribean Sea where warm water enters the Atlantic. The GCM responds with vertical homogenization, which is unrealistic in this case. The problem is especially noticeable at steep slopes, which in our model appear to be up to 700 m within two grid points of 165-km distance. Therefore, a modification of the BD97 method was necessary to restore the universal applicability of the BBL advection scheme. We therefore chose to add a condition to the advection equation to eliminate this difficulty. Thereby, two alternative ways are pursued:

- **Conditional advection;** that is, advection is allowed only if dense water overlies less dense water on the slope (i.e., $\nabla \rho \nabla h < 0$) and if the velocity is directed towards greater depth ($u \nabla h > 0$, “dense downward condition”). Under these conditions cross-isobath bottom layer flow of the level model is rotated completely into the terrain-following system:

\begin{equation}
\rho = -\frac{1}{h}[(\alpha h^2 \Omega), + (\alpha h^2 \Omega h), + (\alpha h^2 \Omega),_h] - [(1 - \alpha) u p, + (1 - \alpha) v p], \\
+ (1 - \alpha) w p], \tag{3}
\end{equation}
(where $\Omega$ is the vertical velocity in the $\sigma$ coordinate system of the bottom layer; and $u$, $v$, and $w$ are, respectively, zonal, meridional and vertical velocities). More specifically, the tracer fluxes are calculated in the topography-following coordinate only if the dense downward condition is fulfilled. Formally, this can be expressed as

\begin{equation}
\alpha^\sigma = \begin{cases} 1 & \text{if } \rho \nabla h < 0, u \nabla h > 0 \\
0 & \text{else} \end{cases} \tag{4a}
\end{equation}
\begin{equation}
\alpha^\sigma = \max(\alpha^\sigma, \alpha^\sigma), \tag{4b}
\end{equation}

Note that the tuning factor $\gamma$ [Eq. (5) of BD97] has been eliminated (essentially is set to 1) in order to focus on our goal of preventing artificial dilution due to convective overturning during an overflow event. This choice corresponds to the bottommost layer of a $\sigma$-coordinate model, thereby combining the virtues of both $z$- and $\sigma$-coordinate models. Note that over flat topography, the solution remains unaffected by the BBL model. As in the GCM, advection is discretized in central differences.

- **Velocity-dependent diffusion;** that is, downward advection of relatively dense water is parameterized by conditional diffusion as described in Eq. (1) and the additional constraint of downward advection velocity in the bottom layer (dense downward condition). Thereby a diffusivity coefficient

\begin{equation}
A^\sigma_{\text{max}}(x, y, t) = \begin{cases} 
\Delta x \Delta y & \text{if } \rho \nabla h < 0, u \nabla h > 0 \\
A^\sigma_{\text{min}} & \text{else} 
\end{cases} \tag{5}
\end{equation}
is applied for each of the two individual directions. The term $A_{\text{BBL}}$ is limited between a minimum of $A_{\text{BBL}} = 10^2 \text{ cm}^2 \text{ s}^{-1}$ and a maximum of $A_{\text{BBL}} = 10^4 \text{ cm}^2 \text{ s}^{-1}$ in order to avoid numerical instabilities. This method is introduced as an alternative to conditional advection. It is even simpler to implement in numerical code than BBL advection and represents a step toward local dependence of the diffusivity coefficient.

The activity of this BBL varies with space and time according to the dense downward condition described above. The BBL is active only if it is needed to assist dense water descending in order to prevent artificial convective dilution during an overflow event. This system can be envisaged as a set of tubes at the bottom of the model topography connecting only selected bottom boxes. The location and number of tubes varies with space and time.

The method does not include a prediction of dense plume thickness. The strength of our method is to provide an efficient prevention of artificial convective mix-
Table 1. Parameters common to all experiments.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_{v}$</td>
<td>Vertical diffusion coefficient</td>
</tr>
<tr>
<td>$A_{m}$</td>
<td>Vertical viscosity coefficient</td>
</tr>
<tr>
<td>$A_{h}$</td>
<td>Horizontal diffusion coefficient</td>
</tr>
<tr>
<td>$A_{h.iso}$</td>
<td>Isopycnal diffusion coefficient</td>
</tr>
<tr>
<td>$c_{d}$</td>
<td>Bottom friction</td>
</tr>
</tbody>
</table>

Table 2. Experiments with diffusion coefficients for the bottom-following BBL in cm$^2$ s$^{-1}$.

<table>
<thead>
<tr>
<th>Exp.</th>
<th>$A_{v,\text{max}}$</th>
<th>$A_{m,\text{max}}$</th>
<th>BBL advection</th>
<th>Velocity-dependent diffusion</th>
</tr>
</thead>
<tbody>
<tr>
<td>NO.BBL</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>DIFF</td>
<td>1.0e8</td>
<td>1.0e2</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>DIFF2</td>
<td>2.0e7</td>
<td>1.0e2</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>DIFF5</td>
<td>5.0e7</td>
<td>1.0e2</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>DIFF15</td>
<td>1.5e8</td>
<td>1.0e2</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>DIFF20</td>
<td>2.0e8</td>
<td>1.0e2</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>DIFF.VD</td>
<td>1.0e8</td>
<td>1.0e2</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>DIFF.VD0</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>ADV</td>
<td>No</td>
<td>Yes</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>DIFF ADV</td>
<td>1.0e8</td>
<td>1.0e2</td>
<td>No</td>
<td>Yes</td>
</tr>
</tbody>
</table>

Fig. 2. Meridional streamfunction for expts (a) NO.BBL, (b) DIFF, (c) ADV, and (d) DIFF ADV. Contour interval = 0.5 Sv.

ing during overflow events, thus preventing excessive dilution of a core water mass. We do not aim at providing a formulation for entrainment/detrainment other than continuity. This GCM feature is unchanged, thus we do not reach a “minimum entrainment solution” but cut the most deleterious shortcoming of mixing in overflows. This approach is universally applicable. In this paper, we show its success in improving water mass representation in a more realistic configuration.

c. Additional forcing

As with other circulation models, this model cannot produce the correct source water mass for the Denmark Straits Overflow Water (DSOW) in the Nordic Seas on its own. The physics of DSOW formation in the Nordic Seas are not fully understood. Different circulation and mixing schemes are proposed (Strass et al. 1993; Mauritzen 1994). Correct cooling of the Norwegian Current seems to be as important as adequate mixing processes in the subsurface East Greenland Current. Even high-resolution models so far have not been successful in producing a correct DSOW. Improved ice models and surface forcing may cure the situation. Therefore, the DSOW in our model still has to be introduced by restoring the temperature in selected boxes on the saddle of the Denmark Strait, as indicated in Fig. 7 at 65°N with a timescale of 20 days. There is no restoring on the southern rise of the strait’s saddle. The existence of a dense water mass at the sill is a precondition for the activity of the BBL.

Different versions of the model have been spin up for 21 years using different versions of the BBL. A reference state is defined by the pure GCM without the BBL after 21 years. Parameters common to all experiments are shown in Table 1. All experiments are listed in Table 2.

3. Results

In this section we look at the large-scale effects of the presence of various BBL model formulations. We expect significant effects on the meridional overturning, the tracer distribution in the deep ocean, and moderate effects on the vertically integrated mass transport.

a. Meridional circulation

The streamfunction of the meridional overturning (Fig. 2) shows a pattern typical of many North Atlantic models. Its strength is relatively weak compared to models with restoring zones south of the North Atlantic sill (like the WOCE-CME models; e.g., Döscher and Redler 1997), but similar to OGCMs without sidewall forcing (Gerdes and Köoberle 1995). The meridional transport shows a maximum of 7–8 Sv at 44°N. These low values are caused by the absence of a full restoring at the southern descent of the North Atlantic sills, covering the whole water column and both connections between the Atlantic and the Nordic Seas east and west of Iceland. Instead, only the DSOW is embossed onto the model by a restoring in the deep and western parts of the Denmark Strait sill.

The application of a BBL shows a clear deepening of the upper cell of the North Atlantic Deep Water. The diffusive part is dominating as is revealed by employing both BBL components together and as single applica-
tions. In experiment DIFF the depth of the zero streamline is increased from 1800 m to the bottom between 45° and 60°N, that is, deepened by 2700 m at most. Experiment ADV with pure advective BBL, causes only a 200-m deepening. The meridional streamfunctions with and without an active BBL component differ by about 3.5 Sv in the DIFF case and 1.5 Sv in the ADV case (Fig. 3). In the combined case of an advective–diffusive (DIFF-ADV) BBL, the individual effects are approximately added. The intersection of the zero streamline with the bottom is shifted by 10° to the south and the transport difference is 5.5 Sv. As a consequence, the bottom cell, representing the circulation of Antarctic Bottom Water (AABW) is reduced from 4.3 Sv in its core to 2–2.5 Sv for the advective–diffusive case. In all cases including the BBL, the Antarctic Bottom Water is literally pushed southward and the vertical extent of this bottom cell is reduced. Maximal vertical displacements are 700 m in the Tropics.

These experiments employing a BBL show a distinctly deeper penetration of NADW. This is confirmed by a constant release of a passive tracer at and north of the North Atlantic sills (release concentration = 1). The tracer enters the North Atlantic and is mixed in shallow NADW components for the reference case NOBBL. A core depth of 1200 m is prevalent in a depiction of the zonal-mean tracer concentration (Fig. 4). The BBL leads to a weakening of the shallow 1200-m core and an additional deeper penetration path of the tracer associated with denser water at the Denmark Strait sill. Instead of a pure horizontal spreading as in the NOBBL case, the tracer also shows a deep maximum in the DIFF-ADV case near 3300 m in the midlatitudes. Several tracer observations (e.g., Doney and Jenkins 1994) show a deep core at this depth associated with a transport maximum of the lower NADW (Roemmich and Wunsch 1985). Döscher et al. (1994) indicate that the deep transport maximum requires an adequate DSOW at the foot of the Denmark Strait sill. (In their WOCE–CME model, the northern overflow is restored in a broad restoring zone covering the southern slope of the Denmark Strait sill. Thus, the absence of a BBL in the WOCE–CME model is much less problematic.) In our model, this is given by restoring directly at the sill and by the bottom layer.

As an alternative to the advection within the BBL, we have considered a velocity-dependent diffusion. As revealed by the cases DIFF-VD0 and DIFF-VD (Fig. 5), which closely resemble the meridional streamfunctions of ADV and DIFF-ADV (Figs. 2c,d), this can be regarded as an adequate replacement for coarse-resolution models.

b. Temperature in the bottom layer

The direct effect of the BBL can be seen in the temperature field in the bottom layers connecting all bottom boxes (Fig. 6). DSOW at temperatures between 0° and 1°C is prevalent in the deeper and shallower parts of
the Denmark Straits. Southward of the sharp topographic drop, warmer water masses of around 5°C can be seen. This sharp gradient is caused by vertical convection, which alternates with horizontal southward transport of DSOW in the sill-level layers. There is no connection between the cold DSOW and the deep Labrador and Irminger Basin cold water masses. This is in clear contrast to observations (e.g., Lazier 1988) that identify the deep water masses of the Labrador Sea as deep NADW originated in the Denmark Straits. In our reference model without BBL, the deeper layers of the Irminger Basin are filled and renewed with Antarctic Bottom Water.

Employing the BBL, the picture looks quite different. The artificial convective mixing, which alternates with horizontal southward transport of DSOW in the sill-level layers, is prevented in this case. Instead, cold water from the Denmark Straits sill descends southwestward, establishing a connection between the sill and the deep Labrador Sea.

Figure 7 illustrates the southward penetration of the dense overflow water. The minimum temperature at the bottom of the Irminger Basin is displayed against the latitude. Cold DSOW at the Denmark Strait's sill at 65.5°N is common to all experiments. The sharp southward rise of temperature in NO, BBL is in contrast to the distinctly stronger preservation of the source water in DIFF, DIFF, VD, and DIFF, ADV on its way south.
Fig. 7. Minimum temperatures at latitude circles at the bottom of the Irminger Sea for various experiments. OBS indicates observational data from the *Atlas of Oceanographic Sections* (Grant 1968).

For the purpose of comparison, observational data from the *Atlas of Oceanographic Sections* (Grant 1968) are added in the curve marked “OBS.” The observed minimum bottom temperature is much closer to the BBL cases than to the reference case. An improved spreading of bottom water masses is crucial for the ability of a GCM to simulate the deep circulation. The close resemblance of observed and modeled temperature indicates that this ability can be achieved even in a coarse-resolution model. As for other parameterizations in ocean models, the effect can be tuned. An appropriate choice and the parameter dependence are explored later in this section.

c. Horizontal circulation

An effect of the additional BBL dynamics on the horizontal circulation can be detected as well. The barotropic streamfunction (Fig. 8) is affected by the BBL north of 35°N and along the western boundary. All BBL model formulations (the diffusive, advective and the combined cases) show a concentration and strengthening of the subpolar gyre off the east Greenland coast. This is associated with a stronger deep western boundary current. The difference between the barotropic streamfunctions with and without BBL reach a maximum of 20 Sv off the east Greenland coast within a 5° latitude range south of the Denmark Straits. A tongue of slightly stronger barotropic transport appears along the western boundary south to the equator. Previous studies have investigated the southward spreading of the signal (Döscher et al. 1994) and attribute similar changes originating in the Denmark Straits to the JEBAR term of the barotropic vorticity equation (Gerdes and Köberle 1995), communicating changing density structures to the barotropic mode.

Fig. 8. Horizontal streamfunction for expts (a) NO, BBL, (b) DIFF, (c) DIFF, ADV, and (d) the difference DIFF, ADV − NO, BBL. Contour interval = 5 Sv for (a–c), 2 Sv for (d).
**Fig. 9.** Influence of the diffusivity coefficient $A^\sigma$ on the meridional streamfunction. Key quantities: (a) Depth of zero streamline separation NADW cell from AABW cell, (b) strength of NADW cell transport, (c) strength of the bottom water cell, and (d) maximum transport difference.

d. Parameter dependence

In order to explore the sensitivity of our results, we varied the diffusivity in the BBL, $A^\sigma$, between 0 and $2 \times 10^8 \text{ cm}^2 \text{ s}^{-1}$ as listed in Table 2 in section 3. Figure 9 shows results for different key quantities: the depth of the zero streamfunction line at 0°N, the NADW transport, the AABW cell transport, and the maximum transport differences between the meridional circulations of the experiments and the reference (as in Fig. 3). The characteristic quantities show an asymptotic approach to a certain level for $A^\sigma \geq 2 \times 10^8 \text{ cm}^2 \text{ s}^{-1}$.

Figure 7 shows the resemblance between observed and modeled minimum temperatures in the deep Irmin-ger Sea, the key region for the North Atlantic thermohaline circulation. This gives us an indication for the choice of $A^\sigma$: The choice of $A^\sigma = 1 \times 10^6 \text{ cm}^2 \text{ s}^{-1}$ corresponds to a tracer transport velocity scale of $10^8 \text{ cm} \text{ s}^{-1}$. It is within the order of magnitude of streamtube the plume model results for different locations in the North Atlantic (Price and Büringer 1994; Jungclaus and Backhaus 1994). Together with the advective contribution the diffusivity is sufficient to give an adequate bottom layer transport. Thus, a $A^\sigma$ near $10^8 \text{ cm}^2 \text{ s}^{-1}$ can be regarded as a reasonable choice for the major overflow region of the North Atlantic.

The choice of $A^\sigma$ near $10^8 \text{ cm}^2 \text{ s}^{-1}$ corresponds to the less sensitive parameter range in Fig. 9. A further increase of $A^\sigma$ would have a less significant effect than a decrease on the system. The NADW cell transport (Fig. 9b) has reached a nearly constant level for $A^\sigma > 1 \times 10^8$. Summarizing, the value of the bottom layer diffusivity can be justified by its effect in comparison with observations. Interestingly, the system shows only a weak sensitivity to a further increase of $A^\sigma$.

Another point of interest is a comparison between the effects of diffusion and advection. Although the diffusion appears to dominate the results, advection contributes significantly. A comparison between both effects is enabled by the horizontal lines in Fig. 9. The dashed line refers to the pure advective case ADV. In three characteristic quantities (Figs. 9a,c,d) it resembles
the weak diffusive case $A^w = 2 \times 10^7$. The pure advective effect is of the opposite sign to that of the diffusive effect only for the NADW transport (Fig. 9b), though the magnitude of the sensitivity to advection in the BBL is small. (The reason for this is a weaker zonal density gradient in the shallower level. The pure advection transports dense water downward without maintaining the horizontal transport leaving the bottom. Because the BBL is primarily active near the western boundary, this results in a slight reduction of zonal pressure gradient in the upper layer.)

The combined advective and diffusive effect is given by the full line. For most key quantities, it approximately represents an addition of $A^\text{DIFF}$ ($A^w = 1 \times 10^7$) with the pure advective effects.

Velocity-dependent diffusion ($A^\text{VD}$) can be regarded as an alternative to BBL advection. The effects, given by the dash–dotted lines in Fig. 9 closely match the pure advection case, which itself resembles the weak diffusive case. Even the NADW transport (Fig. 9b) makes no exception here.

$A^\text{DIFF}$, $A^\text{ADV}$ represents velocity-dependent diffusion plus an explicit bottom layer diffusion (thick dash–dotted line in Fig. 9, see also Table 2). It does not completely match $A^\text{DIFF}$, $A^\text{ADV}$ but shows similar key quantities. Therefore, velocity-dependent diffusion can be regarded as an adequate replacement of an advective BBL, at least for very coarse resolution models.

**4. Summary and discussion**

One of the most deleterious shortcomings of z-coordinate models, the artificial dilution of water masses descending over sloping topography, has been improved using a simple bottom boundary layer (BBL) parameterization in a model of the North Atlantic. This major problem of modeling tracer spreading in the North Atlantic has been identified by several authors (e.g., Duffy et al. 1997). Tracers such as $\delta^{14}$C, entering the Atlantic from the Nordic Seas via the North Atlantic sills, generally reach a depth level too shallow compared to observations. In our model, the bottom-following boundary layer guides dense water from the Denmark Strait sill to deeper layers. Thereby, the artificial convective dilution known from z-coordinate models is prevented. This leads to a better preservation of tracer quantities entering the North Atlantic via the Denmark Straits. As a consequence, the downward penetration of NADW is distinctly stronger than usual in z-coordinate models. This causes a deeper penetration of the NADW circulation cell in the meridional overturning. Furthermore, the barotropic circulation is found to be stronger and concentrated off the east Greenland coast.

In this coarse-resolution application the effect of the diffusivity dominates the advection. In our coarse-resolution model, this is due to generally unrealistic low bottom velocities. An increase of bottom friction will presumably increase the effect of the pure advective BBL. An application in a high-resolution model with generally higher bottom velocities can be expected to show a stronger advective effect, thus enabling a reduction of the diffusive part. Other algorithms for finding a spatially varying diffusivity in the bottom layer model are conceivable. Clearly, there is more work to be done in this area. It is expected that the method improves overflows at sites other than just the Denmark Strait. The direction of the density gradient is the crucial criterion to make the BBL work.

Despite its simplicity, our BBL method improves the capabilities of a z-coordinate model in a configuration typical for coarse-resolution studies or climate models. The general principle of combining bottom boxes to a bottom-following layer (BD) provides a robust basis for different realizations, either diffusive and/or advective.

Even a simple bottom boundary layer (such as ours) seems highly desirable for climate and biogeochemical ocean models. A more realistic representation of the deep tracer distribution and the corresponding circulation can thus be achieved. Aspects of stability of the meridional circulation are also affected by a bottom layer representation. Lohmann (1998) shows a strongly reduced sensitivity to surface forcing fluctuations while his model (a simple rectangular representation of the Atlantic with idealized topography) uses a simple approximation of the BBL presented here. This model behavior concerning the meridional circulation is in accordance with sensitivity tests in the presence of an overflow representation by Döscher and Redler (1997). Both papers conclude that global climate models are likely to overestimate the ocean response to atmospheric fluctuations because they cannot describe the spreading of a dense overflow through the Greenland–Iceland–Scotland Ridge system. Thus, a BBL in climate models could lead to improved near-bottom flows and, as a consequence, to a more stable oceanic response behavior.

Specific locations in the world ocean are of particular interest for an application of the BBL. In the Arctic Ocean, brine release during ice formation on the extended continental shelves leads to dense descending plumes. Furthermore, the representation of deep or bottom water formation in contact with sloping topography (Antarctica, Labrador Sea), the Mediterranean outflow, and overflows through narrow passages in oceanic ridges could be improved in z-coordinate models as a result of application of the BBL.

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**REFERENCES**


