Interactions between thermohaline- and wind-driven circulations and their relevance to the dynamics of the Antarctic Circumpolar Current, in a coarse-resolution global ocean general circulation model

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Abstract. We present an analysis of the interaction between wind driven and thermohaline-driven circulations in a coarse-resolution global ocean general circulation model of the Bryan-Cox code. A series of experiments is described in which the flow is driven by wind forcing only, thermohaline forcing only, or both. In a global ocean with topography, the circulation driven by wind alone is strongly influenced by contours of constant potential vorticity, which limit the barotropic transport to relatively small values. With the same geometry the flow driven by relaxing to observed surface temperature and salinity fields alone contains deep overturning circulation (producing North Atlantic Deep Water and Antarctic Bottom Water (AABW)) and a large barotropic Antarctic Circumpolar Current (ACC) generated by bottom form stress. The ACC is dependent on the overturning circulation, the deep density field and the bottom form stress and increases with $A_{TV}$, the vertical mixing coefficient. If wind is added to this flow, the additional circulation is also dependent on the baroclinic structure but decreases with increasing $A_{TV}$. This applies especially to the ACC, where wind-induced additions are much larger than when the wind acts alone in a homogeneous ocean. An analysis of the dynamic balance of the ACC in the model shows that it is governed by lateral friction, bottom form stress, and wind. The mechanism for driving the ACC by deep convection and bottom topography is revealed in special experiments with simplified topography. In these runs, all topography is removed except for all or part of a submarine ridge across the Drake Passage; this topography alone causes an ACC with barotropic transport of 80 Sv, driven by a deep density difference and pressure gradient across the ridge. This ridge channels AABW formed in the Weddell Sea northward into the South Atlantic, and the ACC is driven by angular momentum conservation across these latitudes of the compensating southward flowing shallower flow (or, equivalently, by the Coriolis force acting on it) above the ridge. In most other parts of the model ocean, bottom form stress acts as a net drag on the zonal current.

1. Introduction

One of the most common ways of determining the present-day state of global ocean circulation from observed hydrographic data is to assimilate sea surface temperature (SST) and sea surface salinity (SSS) climatologies into an ocean general circulation model (OGCM) and to simultaneously apply observed wind stress at the model ocean surface. However, the relative importance of each of the simultaneous forcing constituents is not clear, since the model circulation obtained in this way is the resultant circulation driven by both the wind and the surface thermohaline fields. Also, the way in which the wind-driven and thermohaline-driven circulations interact is unclear. It will be demonstrated that in the presence of such interaction, the model circulation is not the sum of the pure wind driven circulation theory [e.g., Sverdrup, 1947; Stommel, 1948; Munk, 1950] and pure thermohaline-driven circulation and thus cannot be interpreted by these separately. This emphasizes the importance of understanding the interaction, which is the task of this paper. Special attention is given to the dynamical balance of the Antarctic Circumpolar Current (ACC).

Interest in the interaction between wind-driven and thermohaline-driven circulation can be traced back to the late 1960s, when numerical modeling of the ocean had just emerged. Bryan and Cox [1967, 1968a,b] con-
ducted studies on the relative importance of the wind and thermohaline forcing in driving ocean circulation in a flat-bottomed basin model and found that the circulation strengthens as the wind stress increases. These studies were later continued by Bryan [1987] who, using the commonly adopted dynamical balance [e.g., Bryan and Cox, 1988a, b; Holland, 1973], and through numerical experiments, found that the relative importance of wind forcing and thermohaline forcing varies with the vertical diffusivity $A_{TV}$. For $A_{TV}$ less than 0.5 cm$^2$ s$^{-1}$, wind forcing appears dominant, while for $A_{TV}$ greater than 1 cm$^2$ s$^{-1}$, the thermohaline forcing dominates.

Over the last 5 years or so, studies of the sensitivity of the thermohaline circulation to vertical mixing have been carried out in both idealized flat-bottomed basin OGCMs [Colin de Verdière, 1988, 1989; Cummins et al., 1990; Cummins, 1991], and in global OGCMs [e.g., Hirst and Cai, 1994; Cai, 1994]. However, the dependence of wind-forced circulation upon vertical mixing (as modeled by eddy diffusivity) has remained unexamined. As will be shown in this paper, the net wind effect, on both the thermohaline circulation and the barotropic transport, varies substantially with the vertical mixing, especially in the presence of realistic topography. Bottom topography has two main effects. First, large ridges, etc., constrain the location of deep flows, often causing them to be laterally displaced from where they might otherwise be. Second, the pressure field on the topography causes a bottom pressure drag, or form stress. We use the latter term here because this stress sometimes acts on the fluid in the same direction as the net local barotropic transport.

One of the most striking features of the flow in an OGCM with bottom topography is that substantial barotropic transports are generated with thermohaline forcing alone [Holland, 1973; Greatbatch et al., 1991; FRAM Group, 1991; England, 1993; Cai, 1994]. Holland [1973] and Greatbatch et al. [1991] demonstrated the role of bottom pressure in the generation of a barotropic Gulf Stream. In a global water mass study, England [1993] noted the effect of the bottom form stress on the barotropic ACC in the absence of wind forcing. Cai [1994] drove a global OGCM by the surface thermohaline field alone and obtained a substantial barotropic ACC. Killworth [1992] noticed a similarly substantial ACC in the FRAM Group study when the flow was driven by the observed thermohaline field alone, and suggested that the ACC was due to the implicit wind effect already present in the observed density (SST and SSS) field. This so-called implicit wind effect appears to be linked to a belief that the ACC is mainly wind-driven, and it can be traced back to Munk and Palmén [1961]. They considered the balance between lateral friction and wind stress over a flat-bottomed ocean and found that under this balance the ACC was too strong for any realistic lateral friction. They then suggested that the ACC had a strong barotropic component penetrating down to the bottom, where the ACC was retarded by the bottom topographic drag of the mid-ocean ridges along its path. Hidaka and Tsuchiya [1953] confirmed that mid-ocean topography was essential for a realistic model reproduction of the ACC. Through the balance between wind stress and lateral friction, they agreed that the magnitude of the ACC in the absence of topography is determined by $\tau D^2/\nu$ (where $\tau$ is the wind stress magnitude, $D$ is the current width, and $\nu$ is the lateral viscosity). Realistic values of $\nu(10^4$ m$^2$ s$^{-1}$), $D(1000$ km), and $\tau(10^{-4}$ m$^2$ s$^{-1}$) give an unrealistic ACC of about 1000 Sv, a feature sometimes known as the "Hidaka dilemma." The effect of mid-ocean topography has since been pursued in many studies [e.g., Johnson and Bryden, 1989; Wolff et al., 1991; Kryggitos and Cane, 1994].

So far, primitive equation numerical models (e.g., the most widely distributed model of the Bryan-Cox code) have failed to support the balance concept of Munk and Palmén. In a flat-bottomed global OGCM driven by wind only, Bryan and Cox [1972] reproduced the Hidaka dilemma and later found that the inclusion of more realistic topography reduced the ACC transport to the opposite extreme of less than 30 Sv due to the termination of potential vorticity contours at the Drake Passage. Gill and Bryan [1971] described a series of experiments showing the effect of varying the depth of the sill in the Drake Passage on the size of the model ACC, in an idealized Southern Ocean model with a Drake Passage. They found that under both thermohaline and wind forcing with the depth of the passage half of the depth of rest of the ocean, the ACC was 3 times larger than in a run with the passage as deep as the rest of the ocean. The enhancement of the ACC transport is due to the development of an east-west pressure difference below the sill, which drives the ACC in the same direction as the wind forcing. The relevance of this striking result was questioned because of the use of an unrealistic rotation rate, and its significance has since been largely ignored. Cox [1975] then demonstrated that a model with realistic topography and stratification can produce a realistic ACC. This result supported the role of the joint effect of bottom topography and baroclinicity in determining the ACC, as in Gill and Bryan's model, but the link was not made. Similar results have been described by Olbers et al. [1993]. To date, the sizes of ACC transport in OGCMs vary from about 200 Sv [FRAM Group, 1991; Sietzer and Chevree, 1988] to about 120 Sv [Moore and Reason, 1993], and the dynamics responsible for the difference are not completely clear. Here we examine the dynamical balance of the ACC in a coarse-resolution model and attempt to determine the relative importance of the role of wind forcing, thermohaline structure, and their interaction, as well as the role of Drake Passage topography, in driving the model ACC.

This paper is organized as follows. The model and runs are detailed in section 2. In sections 3 and 4 we present the ocean circulation driven first by wind only, and then by thermohaline forcing alone. In section 5 we present the solutions when both wind and thermohaline forcing are applied. Section 6 highlights the relative importance of wind and thermohaline forcing in
the generation of barotropic transports, especially the ACC. The key results are summarized and the conclusions are drawn in section 7.

2. The Model and Experimental Details

This study employs the Pacanowski et al. [1991] version of the Bryan-Cox OGCM, which is based on the primitive equation model described by Bryan [1969] and Bryan and Lewis [1979]. The global model configuration is identical to that of Cai [1994] and has a realistic geometry and topography represented by 12 model levels, with horizontal grid spacing of approximately 3.18° latitude and 5.625° longitude. Bottom stress is parametrized by a quadratic drag law. The vertical eddy viscosity \( A_{TV} \) has the value \( 1 \times 10^{-4} \) m² s⁻¹. The vertical diffusivity for tracer \( \dot{A}_{TV} \) and the horizontal viscosity \( A_m \) vary from run to run and can be found in Tables 1-4. The model uses Cox’s [1987a] parameterization to compute convection implicitly, with enhanced vertical diffusivity in regions of static instability set at 10 m² s⁻¹. When thermohaline forcing is used, the isopycnal diffusion scheme is switched on. The isopycnal mixing is that discussed by Cox [1987b], which is based on the work of Redi [1982].

The various model configurations are summarized in Tables 1-4 and 6. The annually averaged wind stress field used to force the model is that of Holleran and Rosenstein [1983]. The surface thermohaline forcing fields are the annual mean Lentus [1982] SST and SSS climatology, following Cai [1994].

Run W (Table 1) features a homogeneous ocean with topography, in which the ocean is subject to wind stress alone, while in runs II and 2I the ocean is subject to thermohaline forcing alone (without wind forcing). Run II features a large vertical diffusivity \( (\dot{A}_{TV} = 1 \times 10^{-4} \text{ m}^2 \text{s}^{-1}) \), in contrast to run 2I, where a fairly small value \( (\dot{A}_{TV} = 0.2 \times 10^{-4} \text{ m}^2 \text{s}^{-1}) \) is used. Runs IIW and 2IW are exactly the same as runs II and 2I except the model is also subject to the wind stress. The surface thermohaline field is assimilated by Haney relaxation following Cai [1994] using a relaxation time of 6 days. These five runs are the basic runs for this study. We have also carried out many other runs that address issues arising from these basic runs and these will be detailed as we proceed.

Bryan’s [1984] technique of asynchronous integration is applied, allowing the time step for temperature and salinity to vary with depth, increasing from 2 days for levels 1-5 to 16 days at level 12. The equilibrium is defined as a state in which rates of change of the global mean temperature and salinity at each level are less than \( 1 \times 10^{-3} \)°C and \( 1 \times 10^{-3} \) ppt per century.

3. Circulation in a Homogeneous Ocean Driven by Surface Wind Stress

In run W, only the annual wind forcing is applied, and the thermohaline forcing is switched off completely by setting both temperature and salinity to a uniform and constant value of 34.2 ppt and 4.5°C, respectively. The horizontal barotropic transports for this run are shown in Figure 1a and listed in Table 1. These results are consistent with those of Bryan and Cox [1972]. In the limit of low Rossby number, where relative vorticity is small, the flow is guided by the contours of \( f/H \), where \( H \) is the water depth and \( f \) is the Coriolis parameter. \( f/\sin\varphi \) (where \( \varphi \) is the latitude) is plotted in Figure 1b. We can see the similarity between Figures 1a and 1b. In the North Pacific the bottom is relatively flat, and the contours of \( H/\sin\varphi \) are relatively more zonally oriented, leading to the circulation least affected by the topography. In other western boundary current regions the topography reduces the circulation strength from those predicted by Munk [1950]. In the region of the Drake Passage, the sill in the passage blocks many contours, and the few that pass through are confined to the continental margins. As the flow must approximately follow \( f/H \) contours, it is therefore very weak, with an ACC of about 7.6 Sv. A further discussion on topographic effects will be presented in section 6.2.

The meridional overturning stream function for run W, Figure 1c, shows two major cells in each hemisphere: a tropic-subtropic cell associated with strong upwelling

<table>
<thead>
<tr>
<th>Run</th>
<th>W</th>
<th>TS</th>
<th>( A_{TV} ) ( 10^{-4} \text{m}^2 \text{s}^{-1} )</th>
<th>Fr.</th>
<th>Top.</th>
<th>( A_m ) ( 10^{-6} \text{m}^2 \text{s}^{-1} )</th>
<th>ACC, Sv</th>
<th>GS, Sv</th>
<th>K, Sv</th>
<th>IT, Sv</th>
<th>NADW, Sv</th>
<th>AABW, Sv</th>
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<td>no</td>
<td>yes</td>
<td>yes</td>
<td>real</td>
<td>9</td>
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<td>10.4</td>
<td>30.9</td>
<td>6.7</td>
<td>17.2(10.95)</td>
<td></td>
</tr>
<tr>
<td>2I</td>
<td>no</td>
<td>yes</td>
<td>0.2</td>
<td>yes</td>
<td>real</td>
<td>9</td>
<td>24.7</td>
<td>3.3</td>
<td>0.3</td>
<td>3.8</td>
<td>6.43</td>
<td></td>
</tr>
<tr>
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<td>yes</td>
<td>1.0</td>
<td>yes</td>
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<td>80.2</td>
<td>8.7</td>
<td>0.3</td>
<td>13.7</td>
<td>15.8</td>
<td></td>
</tr>
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<td>yes</td>
<td>real</td>
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<td>32.2</td>
<td>21.4</td>
<td>19.9</td>
<td>51.6(30.7)</td>
</tr>
</tbody>
</table>

The wind forcing (W) is either switched on ("yes") or switched off ("no"), as are the thermohaline forcing (TS) and bottom friction (Fr.). Bottom topography (Top.) is either realistic ("real") or flat, except in runs F1Lp and F1Lpp where either one grid point of topography or six grid points of topography are retained, and the rest of the ocean is flat-bottomed. \( A_{TV} \) is vertical diffusivity for temperature and salinity, and \( A_m \) is the horizontal viscosity. Also shown in the table are the Antarctic Circumpolar Current (ACC), the Gulf Stream (GS), Indonesian Throughflow (IT), the Kuroshio (K), the maximum North Atlantic Deep Water (NADW) formation rate, and the maximum Antarctic Bottom Water (AABW) formation rate, with the formation rate in the Weddell Sea shown in brackets.
4. Circulation Driven by Surface Thermohaline Forcing Alone

A description of some of the circulation features in the model driven by surface thermohaline forcing alone has been given by Cai [1994]. This is included in the following description.

4.1. Baroclinic and Barotropic Circulations

As was discussed above, runs 11 and 21 differ in having a large and small vertical diffusivity respectively (Table 1). The baroclinic flows associated with the “conveyor belt” [Broecker, 1987] arc produced as detailed by Cai [1994]. In the Southern Ocean, substantial barotropic transports are produced (top panels of Figure 2; Table 1) without explicit wind forcing. Although the observed SST and SSS fields already implicitly include some information about the wind, the barotropic transports are not driven by the implicit wind effect as suggested by Killworth [1992] but are indirectly driven by the bottom pressure field [Cai, 1994] arising from the joint effect of baroclinicity and bottom topography [Gill and Bryan, 1971; Holland, 1973; Cai, 1994]. In short, the presence of bottom topography leads to horizontal divergence of a fluid column in certain locations; this increases currents and pressure gradients near the bottom and generates vertically integrated barotropic transport. This is further discussed in sections 6.2, 6.4, and 6.5. Both the baroclinic and barotropic circulations (top panels of Figure 2) intensify with the increase in vertical diffusivity from run 21 to run 11, as in the Cai [1994] model (see Table 1).

4.2. Implied Free Surface Height, Air-Sea Heat Exchange, and Poleward Heat Transport

The implied free surface height fields referenced to 1500 m are shown in the top panels of Figure 3. In the absence of explicit wind forcing, the spatial variation of this field is mainly due to heating and cooling and to fresheining and salinization (increasing salinity due to implied evaporation). We see that the equator-to-pole density contrast generates a downward slope poleward. Due to convective sinking in the Gulf Stream and ACC regions, well-defined gradients in height across the ACC and the Gulf Stream are produced. The laws associated with these gradients are related to the strong air-sea heat exchange (Figure 4a, negative values mean that the ocean loses heat) in these regions. Even in the absence of explicit wind forcing, the air-sea heat exchange feature an intense heat loss in the polar and the mid-latitude western boundary current regions and a broad but weak heat gain in the interior. Like the barotropic transports in the previous section, the magnitudes of the implied free surface height variations and of the air-sea heat exchange increase with Argy, in response to a more intense vertical and horizontal heat exchange. The larger surface heat exchange in run 11 relative to run 21 is accompanied by a substantial increase in horizontal heat transport in run 11 over 21 (Figure 5).

4.3. Water Mass Properties and Overturning Circulations

The top panels of Figures 6, 7, 8, and 9 show the variation of global zonally averaged temperature, salinity, potential density, and zonally integrated meridional overturning stream function with depth, respectively. In both runs 11 and 21 the global main thermocline and the low-salinity tongue associated with the Antarctic
Intermediate Water (AAIW) are reproduced (Figures 6 and 7). In the absence of explicit wind forcing, the realistic production of water masses is not due to the implicit wind forcing (see the zonal mean forcing case of Cai [1994]), but is largely determined by the surface heat and freshwater fluxes in the polar, subpolar, and western boundary current regions, which are in turn determined by convection [Cai, 1994]. These fields are asymmetric about the equator; for example, in the region poleward of 40°S, the pycnocline is more diffuse and penetrates deeper than in the region poleward of 40°N (Figure 8). In both runs, qualitatively realistic meridional overturning stream function cells (Figure 9) are produced. These cells are a northern hemisphere convective cell, mainly the North Atlantic Deep Water (NADW) formation cell (from about 60°N to the equator), a southern sinking cell (from the equator to about 55°S), and an Antarctic Bottom Water (AABW) formation cell (from 55°S south). The latter two cells are connected as the AABW spreads northward and turns upward.

With a larger $A_{TV}$ in run 11, the thermocline and pycnocline are deeper and more diffuse, and the salinity minimum reaches greater depth; the meridional overturning cells are stronger and penetrate to greater depth, and the stronger AABW spreads further north. At abyssal depths a larger $A_{TV}$ yields a lower salinity and a higher temperature and hence a lower density.
5. Circulation Driven by Both Thermohaline and Wind Forcing

5.1. Baroclinic and Barotropic Circulations

The baroclinic flows associated with the conveyor belt in runs 1IW and 2IW, that is, in the presence of both thermohaline and wind forcing, are more realistic than those under thermohaline forcing alone. Since the qualitative features are similar to those without wind forcing, a description will not be repeated here. The barotropic transports in both runs are shown in the middle panels of Figure 2. A comparison of barotropic transports between runs with and without wind shows that with the exception of the Kuroshio, which is independent of the value of $A_{TV}$, the wind-induced increase in barotropic transport is larger with a smaller $A_{TV}$ (bottom panels of Figure 2; Tables 1-4). For example, the Gulf Stream in run 2IW increases by 22.2 Sv from run 2I, compared to an increase in run 1IW of 21.3 Sv from run 1I; the ACC in run 2IW increases by 50.4 Sv from run 2I, compared to an increase in run 1IW of 39.4 Sv from run 1I. These features are associated with the effect of bottom form stress, which is in turn associated with the wind-induced enhancement of deep and bottom water formation, as will be discussed in section 6.2.

5.2. Implied Free Surface Height, Air-Sea Heat Exchange, and Poleward Heat Transport

The middle panels (ii) of Figure 3 show enhancement of the poleward slopes of the implied free surface height in the presence of both thermohaline and explicit wind forcing. This enhancement is due to increased heating and cooling and to freshening and salinization as...
Figure 4. The same as in Figure 2, but for surface heat flux (in watts per squared meter, negative values mean that the ocean loses heat). Contour interval is 50 W m$^{-2}$ for top and middle panels and 20 W m$^{-2}$ for bottom panels.

As a result of wind effects (Figure 4), the free surface height variations associated with wind-driven subtropical gyres are now evident. Also evident is a region of low free surface elevation extending from the eastern Pacific westward along the equatorial Pacific.

The signatures of free surface height in the polar and equatorial regions are intimately linked via the thermohaline circulation and its interaction with the wind-driven circulation. This point becomes clear when the free surface height fields in the middle panels of Figure 3 are compared with the air-sea heat exchange fields in the middle panels of Figure 4. In the presence of explicit wind forcing, wind-driven equatorial upwelling brings up the subsurface cold water. This increases the intake of heat through its upper surface. The surface heat intake is almost independent of the vertical mixing coefficient because cool water is brought up near to surface depth (bottom panels of Figure 4) regardless of the strength of vertical mixing. The wind-induced heat intake is then transported by the wind-induced circulation, especially by the large-scale western boundary currents, to the convection regions, where enhanced heat losses take place through convective cooling. In this way, wind forcing increases the poleward heat transport (Figure 5) and enhances the overturning circulation (see section 5.3).

5.3. Water Mass Properties and Overturning Circulations

The middle panels of Figures 6, 7, and 8 show the global zonally averaged temperature, salinity, and po-
of heat intake by the ocean in the equatorial region, which is then transported poleward, convectively mixed downward, and carried around by the NADW outflow, so that the temperature increases throughout almost the entire ocean. Fifth, wind enhances evaporation in low latitudes and precipitation in high latitudes (with the restoring boundary conditions) so that salinity increases in the tropical-subtropical upper ocean and decreases in the polar and deep regions. In the northern hemisphere, as water with enhanced salinity is advected poleward to, and convectively mixed downward in, the NADW region, salinity in the NADW increases. In contrast, in the southern hemisphere the enhanced salinity is less readily advected across the ACC. The resulting low-surface salinity along the ACC penetrates downward and spreads equatorward, increasing salinity at depth. In general, subsurface density generally decreases, due to the combined effects of temperature and salinity, and more so near the surface than at depth.

The middle panels of Figure 9 show the zonally integrated meridional overturning stream function for runs 11W and 21W. The bottom panels of the same figure show the differences between runs with and without wind forcing. From Figures 9 and 1c we see that the overturning cells are not the simple superposition of the wind-driven and the thermohaline-driven cells. Most of the wind-induced cells seen in Figure 1c are lifted to a shallower depth. However, the depth reduction of the Deacon cell is less than that of the other wind-driven cells, especially that of the northern subtropical-subpolar cell. This hemispheric asymmetry is partially due to the smaller wind-driven northern subtropical-subpolar cell and partially due to the weaker stratification in the Deacon cell region. The Deacon cell downwells from the surface to 1500 m in run 11W and 2500 m in run 21W at latitudes from 35°S to 50°S and then upwells from 50°S to 65°S, similar to that in FRAM [Döös and Webb, 1994].

The bottom panels (ii) of Figure 9 also show that the influence of the wind-induced circulation intensifies the AABW and the NADW formation cells. The mechanisms of the intensification of these two cells are different. The intensification of the NADW formation requires a larger heat and salt transport. The wind-driven circulation drives the warm and salty subtropical water to the convection region. This leads to the warmer and saltier water seen in the bottom panels of Figures 6 and 7. In the southern polar region the ACC hinders the meridional heat and salt transport. The enhancement of AABW is made through the Deacon cell. The upwellings associated with the Deacon cell strengthen the vertical mixing, destabilize the water column, and hence enhance the convection and the AABW formation. The enhanced convection then mixes the surface fresh water downward, resulting in the freshening seen in the bottom panels of Figure 7. Again the above features are more pronounced in run 21W than in run 11W. For example, the AABW formation cell in run 21W increases by 13.3 Sv from run 21 but increases by only 5.1 Sv in run 11W from run 11. It will be demonstrated that this
dependence of the wind-induced change in the AABW formation cell upon $A_T$ plays a significant role in generating the dependence of the wind-induced ACC upon $A_T$.

To summarize briefly, we have seen that the qualitative features of the thermohaline circulation, the associated water mass properties, air-sea heat exchange, and poleward heat transport can all be produced in a coarse-resolution global model, without the explicit wind forcing. We have also seen that the role of the explicit wind forcing on these fields depends upon the vertical mixing coefficient. With a smaller vertical mixing coefficient, the explicit wind forcing modifies the ocean circulation to a greater extent. This is because wind forcing tends to increase vertical mixing, but with larger vertical diffusivity, there is a smaller role for wind forcing to play.

6. The Modeled ACC and Other Barotropic Transports

One of the most interesting features of the barotropic transport in the model is that in the presence of bottom topography the wind-driven barotropic transport in a homogeneous ocean (except that in the North Pacific) is very small, whereas in a stratified ocean, even without wind forcing, substantial barotropic transports (especially the ACC) are produced. We have also seen that when a smaller vertical mixing coefficient is used, larger wind-induced barotropic transports and greater wind-induced increases in thermohaline circulation cells are generated. The reasons for these features, and some related issues, are explored in this section.
6.1. Effect of Bottom Friction

The question arises as to whether the dependence of wind induced transport upon $A_{TV}$ is due to the effect of bottom friction. For example, in run 21W the circulation is much weaker at depth than that in run 11W (see Figure 9); the stronger bottom stress in run 11W, due to stronger bottom flow, may reduce the wind-induced barotropic transport. To test the effect of bottom friction, we have rerun experiments 11 and 11W, but without bottom friction. These two experiments are termed 11nf and 11Wnf. The barotropic transports are listed in Table 2. The barotropic ACC, Gulf Stream, and other transports are virtually unaffected. We have also carried out runs F11 and F21, which are identical to runs 11 and 21, respectively, except in a flat-bottomed ocean. In these two runs the only factor that generates barotropic transport is frictional bottom stress. The results show that the barotropic transports observed in runs 11 and 21 disappear (Table 2) and that the maximum transport is less than 1 Sv. These results demonstrate that the effect of frictional bottom stress is negligible, as was noticed in many previous studies [Holland, 1973; Bryan and Cox, 1972, Gill and Bryan, 1971].

6.2. Effect of Topography

Additional experiments are listed in Table 3. As is already clear, the effect of topography in a homogeneous ocean (section 3, run W) is drastically different from that in a stratified ocean (section 5.1, runs 11 and 21), and these will be addressed separately.
We first discuss the effect of bottom topography in a homogeneous ocean. For comparison, run FW was carried out, which is the same as run W except that it is in a flat-bottomed ocean. In this run, the flow in the western boundary current regions is analogous to that of Munk [1950], in which the vorticity balance is

$$
\beta \psi_x = \frac{1}{\rho_0 H} \text{curl} \tau + \frac{A_m}{H} \nabla^4 \psi.
$$

Here $\psi_x$ is zonal ($x$) derivative of the stream function ($\psi$), $\tau$ is the wind stress vector, $H$ is the water depth, $\beta$ is the latitudinal derivative of the Coriolis parameter, $\rho_0$ is the reference density, and $A_m$ is the lateral friction coefficient. Near the western boundary, the planetary vorticity advection is balanced by the diffusion of relative vorticity, while in the ocean interior the planetary vorticity advection balances the wind stress curl, that is, a Sverdrup balance. In the ACC region the momentum balance is between lateral friction and wind forcing, as in the model of Hidaka and Tsuchiya [1953]. In the flat-bottomed ocean, these balances for western boundary currents and the ACC hold regardless of whether the ocean is homogeneous or stratified. This means that the wind-induced barotropic transport is independent of stratification and hence of the vertical mixing coefficient. Note that the realistic ACC in run FW is a fortuitous happenstance, entirely due to the use of a large horizontal viscosity ($A_m$).
Figure 9. The same as in Figure 7, but for the global zonally integrated meridional overturning circulation (in sverdrups). The sign convention is as in Figure 1c. Contour interval is 5 Sv for top and bottom panels and 2 Sv for bottom panels.

Table 2. The Same as in Table 1, but for Runs Described in Section 6.1

<table>
<thead>
<tr>
<th>Run</th>
<th>W</th>
<th>TS</th>
<th>$A_{TV}$, $10^{-7}$ m$^3$s$^{-1}$</th>
<th>Fr.</th>
<th>$A_{m2}$, $10^5$ m$^2$s$^{-1}$</th>
<th>ACC, Sv</th>
<th>GS, Sv</th>
<th>K, Sv</th>
<th>IT, Sv</th>
<th>NADW, Sv</th>
<th>AABW, Sv</th>
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<td>1.0</td>
<td>no</td>
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<td>9</td>
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<td>30.0</td>
<td>32.2</td>
<td>21.4</td>
<td>19.9</td>
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<td>9</td>
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<td>no</td>
<td>flat</td>
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<td>0.7</td>
<td>3.7</td>
<td>1.0</td>
<td>4.9</td>
<td>8.1</td>
</tr>
</tbody>
</table>

We have seen in section 3 that the wind-driven circulation in a homogeneous ocean with topography (run W) has a tendency to follow the $H/\sin \phi$ contours. In this situation the vorticity balance is

$$\beta \psi - \frac{f}{H} J(\psi, H) = \text{curl} \frac{\mathbf{v}}{\rho_{\mathbf{H}}} + A_m \nabla^2 [\nabla (\nabla \psi / H)],$$

where $J$ is the Jacobian operator. A simple rearrangement leads to
Table 3. The Same as in Table 1, but for Runs Described in Section 6.2

<table>
<thead>
<tr>
<th>Run</th>
<th>W.</th>
<th>TS.</th>
<th>( A_{TV} ), ( 10^{-4} \text{m}^2\text{s}^{-1} )</th>
<th>Fr.</th>
<th>Top.</th>
<th>( A_m ), ( 10^{6} \text{m}^2\text{s}^{-1} )</th>
<th>ACC, Sv</th>
<th>GS, Sv</th>
<th>K, Sv</th>
<th>IT, Sv</th>
<th>NADW, Sv</th>
<th>AABW, Sv</th>
</tr>
</thead>
<tbody>
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<td>yes flat</td>
<td>9</td>
<td>130.7</td>
<td>21.6</td>
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<td>14.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>yes</td>
<td>yes</td>
<td>1.0 yes flat</td>
<td>9</td>
<td>131.0</td>
<td>21.6</td>
<td>31.8</td>
<td>14.5</td>
<td>27.3</td>
<td>69.1</td>
<td>(47.2)</td>
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<tr>
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<td>1.0 yes flat</td>
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<td>0.01</td>
<td>0.05</td>
<td>22.7</td>
<td>57.9</td>
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<td>yes</td>
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<td>9</td>
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<td>31.8</td>
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<td>0.01</td>
<td>0.05</td>
<td>8.1</td>
<td>25.8</td>
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<td></td>
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<td>yes</td>
<td>2.0 yes real</td>
<td>9</td>
<td>155.8</td>
<td>18.4</td>
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<td>27.6</td>
<td>28.8</td>
<td>67.5</td>
<td>(42.2)</td>
<td></td>
</tr>
</tbody>
</table>

\[
(\beta - \frac{f}{H} H_y) \psi_x + \frac{f H_x}{H} \psi_y -
\]
\[
\text{curl} \frac{\tau}{\rho_0 H} + A_m \nabla^2 \left[ \nabla \left( \nabla \psi / H \right) \right].
\]  

In comparison with (1), equation (3) contains terms that depend on slopes. All other things being equal, the same wind stress would generate different current systems in oceans with level and inclined bottoms. In particular, (3) shows that bottom slope can either intensify or diminish the \( \beta \) effect. An examination of the balance in run W shows that it is consistent with those found by Holland [1973] and Schulman [1970].

In stratified circulation without bottom topography, barotropic transports are independent of the stratification because of the absence of bottom form stress. This can be seen in a comparison between runs F1I1W and F2I1W with FW (Table 3). Runs F1I1W and F2I1W are identical to runs F1I and F2I, except for additional wind forcing. The barotropic transports in these two new runs are almost entirely driven by wind forcing, as comparison with runs F1I and F2I shows.

As was discussed in section 4.1, in the presence of both stratification and topography, the model ACC and other barotropic transports may be regarded as being generated by bottom form stress. This stress acts to drive the current in certain preferred locations (specifically the Drake Passage region, as discussed below). This stress is strongly dependent upon the bottom density field and bottom flow, which in turn are strongly dependent upon the strength of vertical diffusion, deep water formation, and the associated overturning circulation. Relative to run 2I, a greater vertical mixing in run 11 decreases the stability of a water column, leading to enhancement of deep and bottom water formation, which in turn generates larger bottom density gradients, and induces stronger barotropic transports (Table 3). The Gulf Stream is larger in run 11 compared to that in run 2I because the surface northward flow is stronger in the presence of a larger vertical mixing as a result of a larger “subduction” by the intensified sinking. The high sensitivity to the vertical mixing coefficient can be further seen in run JI, carried out under otherwise the same conditions as in run 11 (no wind forcing) except that \( A_{TV} \) is increased to 2 \( \text{cm}^2 \text{s}^{-1} \). The ACC reaches 155.8 Sv, the Gulf Stream reaches 18.4 Sv, and the Indonesian Throughflow reaches 27.8 Sv (Table 3).

When wind is added to the buoyancy-driven circulation, the heat and mass transports are generally increased as noted in section 5.3, and these additions are larger for smaller vertical diffusivity (or mixing). For example, when the wind is added to runs 11 and 2I, giving runs 11W and 21W, the resulting increases in NADW and AABW formation (Table 1) are larger for smaller \( A_{TV} \). This is because the wind-driven Ekman pumping and resultant advection cause de facto mixing effects that are more significant when the actual diffusivity is weaker.

### 6.3. Sensitivity of the ACC to Horizontal Viscosity

In this section we test the sensitivity of the model ACC to changes in \( A_m \). Additional experiments are listed in Table 4. In most of our model runs we use

Table 4. The Same as in Table 1, but for Runs Described in Section 6.3

<table>
<thead>
<tr>
<th>Run</th>
<th>W.</th>
<th>TS.</th>
<th>( A_{TV} ), ( 10^{-4} \text{m}^2\text{s}^{-1} )</th>
<th>Fr.</th>
<th>Top.</th>
<th>( A_m ), ( 10^{6} \text{m}^2\text{s}^{-1} )</th>
<th>ACC, Sv</th>
<th>GS, Sv</th>
<th>K, Sv</th>
<th>IT, Sv</th>
<th>NADW, Sv</th>
<th>AABW, Sv</th>
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<td>130.7</td>
<td>21.6</td>
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<td>9</td>
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<td>29.7</td>
<td>21.0</td>
<td>17.5</td>
<td>45.4</td>
<td>(27.5)</td>
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</tr>
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</table>
$A_m = 9 \times 10^3 \text{ m}^2 \text{s}^{-1}$. This relatively large viscosity is necessary to resolve Munk's boundary layer, which requires $A_m$ to be constrained by $A_m \geq \beta(\Delta 2\sqrt{3} \Delta a)^3$, where $\Delta$ is the grid spacing [Bryan et al., 1975]. This requirement limits the value of $A_m$ to about 5 to $9 \times 10^3 \text{ m}^2 \text{s}^{-1}$ in these experiments. Note that in the absence of some surface forcing the currents are weaker, and one may be able to use a smaller value [Huang and Chen, 1994] as in some of the following experiments.

In a flat-bottomed homogeneous ocean driven by wind forcing alone, the momentum balance of the ACC is between wind stress and lateral friction [Munk and Palmén, 1961] and is very sensitive to changes in $A_m$: it increases from 130.7 to 155.7, and to 267.4 Sv as the $A_m$ decreases from $9 \times 10^3 \text{ m}^2 \text{s}^{-1}$ to $7 \times 10^3 \text{ m}^2 \text{s}^{-1}$, and to $3 \times 10^3 \text{ m}^2 \text{s}^{-1}$ (see runs FW, FWsm1Am, and FWttnAm in Table 4). This variation is not directly proportional to $1/A_m$ because the current width also varies.

In a homogeneous ocean with topography driven by wind forcing alone, the ACC is strongly influenced by the potential vorticity contours, increasing from 7.6, to 7.9, and to 9.6 Sv as the $A_m$ decreases from $9 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ to $7 \times 10^3 \text{ cm}^2 \text{s}^{-1}$, and $3 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ (see runs W, Wsm1Am, and WttnAm in Table 4).

In an ocean with topography driven by thermohaline forcing alone, the ACC is under the balance between lateral friction and bottom form stress, and it increases from 80.2 to 85.5 Sv as the $A_m$ decreases from $9 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ to $7 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ (see runs 111Am in Table 4). When an $A_m$ of $20 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ is used, a value more than double that of run II, the ACC decreases by only 10 Sv from that of run II (see run 11WhgAm in Table 4). Thus in the presence of bottom topography the ACC is less sensitive to $A_m$ than in the flat-bottomed ocean driven by wind forcing alone because of the constraints of topography on the current.

In an ocean with topography driven by both thermohaline and wind forcing the ACC is under the balance between lateral friction, bottom topography, and wind forcing, and it increases from 119.6 to 130.1 Sv as the $A_m$ decreases from $9 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ to $7 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ (see runs 11W and 11Wsm1Am in Table 4). When an $A_m$ of $20 \times 10^3 \text{ cm}^2 \text{s}^{-1}$ is used, the ACC is 97.8 Sv (see run 11WhgAm in Table 4). These results confirm that in the presence of bottom topography the ACC is not as sensitive to $A_m$ as it is in the flat-bottomed wind-driven ocean. The detailed momentum balance is given in section 6.4. We note, however, that in the real ocean the driving of the ACC by wind and bottom form stress is probably also balanced by bottom form stress (acting as drag in most locations) rather than lateral friction, as suggested by Munk and Palmén [1951]. Much of this bottom drag is probably due to topographic features that are unresolved in this and higher-resolution numerical models.

### 6.4. Momentum Balance Along the ACC in the Presence of Bottom Form Stress

In the study by Gill and Bryan [1971], the joint effect of the baroclinicity and an idealized submarine ridge in the form of a submerged sluice gate (the water depth of the gate was set to one-half the depth of the rest of the ocean) was found to have accelerated the ACC in the eastward direction. Our model results of runs 21, 21, and 31 clearly support the result of Gill and Bryan [1971]. In view of this, we have carried out an analysis similar to that in their study. The steady state eastward component of the momentum equation (see Bryan [1969] for details) is:

$$
\frac{m}{a} [(u^2)_x + (uv/m)_x + (wu)_x - 2mnwv/a] - 2\Omega v +
\frac{m}{a} (P/p_0) \lambda - A_0 u_{zz} - F^\lambda = 0.
$$

(4)

Here $m = \sec \varphi$, $n = \sin \varphi$, $u = a \lambda \text{ m}^{-1}$, and $v = a \varphi$, where $a$ is the radius of the Earth, $\lambda$ is the longitude; $\varphi$ is the latitude ($\varphi < 0$); $\Omega$ is the angular velocity of the Earth; $P$ is the pressure; $p_0$ is average density; and $F^\lambda$ is the lateral friction term. Integrating equation (4) with respect to longitude and depth throughout the water column and over the zonal strip which passes through the Drake Passage yields

$$
\int_{-\varphi_1}^{\varphi_2} \int_{-h}^{0} \int_{0}^{2\pi} \frac{m}{a} [(u^2)_x + (uv/m)_x + (wu)_x - 2mnwv/a] \text{cos}\varphi d\lambda dz \text{d} \varphi + \int_{-\varphi_1}^{\varphi_2} \int_{0}^{2\pi} P h \lambda d\lambda d\varphi + \int_{-\varphi_1}^{\varphi_2} \int_{0}^{2\pi} -\lambda(0) \text{cos}\varphi d\lambda d\varphi + \int_{-\varphi_1}^{\varphi_2} \int_{0}^{2\pi} \lambda(-h) \text{cos}\varphi d\lambda d\varphi + \int_{-\varphi_1}^{\varphi_2} \int_{-h}^{0} \int_{0}^{2\pi} F^\lambda \text{cos}\varphi d\lambda dz \text{d} \varphi = 0.
$$

(5)

In (5), the first term is the nonlinear term, the second is the bottom form stress term, the third is the wind forcing term, the fourth is the bottom friction term, and the fifth the lateral friction term. Given a wind stress field, the wind forcing term is constant. Hence, for the purpose of intercomparison among runs the results are normalized by the magnitude of wind forcing term, and are listed in Table 5.

Several features emerge from the analysis. First, in the presence of both baroclinicity and topography, say runs 11, 21, and 31, the balance is mainly between the bottom form stress and the lateral friction terms. One normally expects bottom form stress to constitute a drag on the current, which would imply a positive sign here for the bottom form stress term, and this probably applies over most of the model ocean. That this term is large and negative here is largely due to the effects of the sill at the Drake Passage, as discussed below. If equation (5) is evaluated over a broader band of latitudes than just those passing through the Drake Passage, the term is less negative, indicating positive drag from the additional area. Second, the addition of wind forcing enhances the bottom form stress term, and with smaller
vertical mixing, the enhancement is greater. In particular, the wind-induced enhancement of the form stress in run 21W is larger than the wind term itself. This is the reason why the wind-induced change in ACC transport is greater with a smaller vertical mixing coefficient (e.g., run 21W) than with a larger vertical mixing coefficient (e.g., run 11W).

The strong ACC in the absence of wind forcing suggests that deep water on the Atlantic side of the Drake Passage must be denser than that on the Pacific side, so that the dynamic height is greater in the Pacific than in the Atlantic and hence drives eastward flow. Table 5 also shows the density difference (Pacific side minus Atlantic side, negative means that Atlantic water is denser) of the potential density at level 10, averaged over the Drake Passage band at the model points of two meridians immediately next to the model Drake Passage. The density on the Atlantic side is indeed greater than that on the Pacific side. Again, the wind-induced change in density difference is greater with a smaller vertical mixing coefficient. A question arises as to whether the real ocean behaves in a similar way. To examine this, Levitus [1982] data were interpolated to the depth of model level 10 at the horizontal grids, and the density difference across the Drake Passage was calculated in the same way as for the model results. The results confirm that the density is larger on the Atlantic side with a potential density difference of 0.0834 kg m$^{-3}$.

We should point out that the lateral viscosity and friction term (last term in (4) and (5)) in this model (and in others with comparable resolution) is unrealistically large, compared with the real ocean. In the latter we expect that the principal drag is due to bottom form drag, as was discussed above.

### 6.5. Sensitivity of ACC to the Topography of the Drake Passage

The role of the density difference across the Drake Passage in influencing the ACC and the sensitivity of this density difference to the topography of the model Drake Passage are further investigated in runs F11, F11p, and F11pp (Table 6). Run F11lp is the same as run F11 (flat-bottomed everywhere), except at one model grid point immediately north of the tip of the model Antarctica Peninsula, where the model depth is set to 2350 m. Run F11pp is also the same as in run F11 except that along the meridional line of the Drake Passage, the model topography of the Drake Passage is retained. This includes six model grid points, and the water depths of these six points from south to north are 2350, 3250, 3250, 1050, 215, and 215 m. The model barotropic transports are shown in Figure 10. The striking feature is that run F11lp produces an ACC of about 70 Sv, compared with an ACC of 0.8 Sv in run F11, while run F11pp produces an ACC of 110 Sv.

The question arises as to why there is such high sensitivity of the ACC to the Drake Passage topography.

### Table 5. Eastward Components of Forces along the ACC

<table>
<thead>
<tr>
<th>Run</th>
<th>Wind</th>
<th>Nonlinear</th>
<th>Bottom Friction</th>
<th>Lateral Friction</th>
<th>Bottom Form Stress</th>
<th>Density Difference, kg m$^{-3}$</th>
</tr>
</thead>
<tbody>
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<td>2I</td>
<td>0</td>
<td>0.04</td>
<td>0.02</td>
<td>0.75</td>
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</tbody>
</table>

The eastward momentum is first integrated with respect to longitude from the eastern to western boundary, to the depth throughout the water column, and to the latitudes of a zonal strip which passes through the Drake Passage. Shown are the average of the integrals over the latitudes covering the strip. These integrations include the wind forcing term, nonlinear term, bottom friction term, lateral friction term, and bottom form stress. The results are normalized by the magnitude of the wind forcing term. For comparison purposes the various terms in runs without wind forcing are also normalized by the wind forcing term, which is fixed. Also shown is the difference in the potential density normalized by 0.001 at level 10 between the Pacific side and the Atlantic side averaged over the Drake Passage band at the two meridional lines immediately next to the Drake Passage, one on the Pacific side and the other on the Atlantic side (negative means the Pacific side is lighter).

### Table 6. The Same as in Table 1, but for Runs Described in Section 6.5

<table>
<thead>
<tr>
<th>Run</th>
<th>W.</th>
<th>TS.</th>
<th>$A_{TV}$, $10^{-4}$m$^{2}$s$^{-1}$</th>
<th>Fr.</th>
<th>Top.</th>
<th>$A_{N}$, $10^{5}$m$^{2}$s$^{-1}$</th>
<th>ACC, Sv</th>
<th>GS, Sv</th>
<th>K, Sv</th>
<th>IT, Sv</th>
<th>NADW, Sv</th>
<th>AABW, Sv</th>
</tr>
</thead>
<tbody>
<tr>
<td>F11</td>
<td>no</td>
<td>yes</td>
<td>1.0</td>
<td>yes</td>
<td>flat</td>
<td>9</td>
<td>0.8</td>
<td>0.04</td>
<td>0.01</td>
<td>0.05</td>
<td>22.7</td>
<td>57.9 (35.2)</td>
</tr>
<tr>
<td>F11lp</td>
<td>no</td>
<td>yes</td>
<td>1.0</td>
<td>yes</td>
<td>1 point</td>
<td>9</td>
<td>67.3</td>
<td>0.0</td>
<td>0.0</td>
<td>0.7</td>
<td>19.3</td>
<td>58.9 (41.5)</td>
</tr>
<tr>
<td>F11pp</td>
<td>no</td>
<td>yes</td>
<td>1.0</td>
<td>yes</td>
<td>6 points</td>
<td>9</td>
<td>110.0</td>
<td>0.0</td>
<td>0.0</td>
<td>1.0</td>
<td>18.2</td>
<td>60.3 (42.9)</td>
</tr>
</tbody>
</table>
Stream Function (Sv)

(a) run F11

(b) Run F111p

(c) Run F11pp

Figure 10. Horizontal barotropic transport stream function (in sverdrups) for (a) run F11, (b) run F111p, and (c) run F11pp (see text for the details of these runs). Contour interval is 10 Sv.

In order to understand the associated processes, horizontal flows at level 6 (370 m) level 10 (2350 m), and level 12 (4150 m) are plotted in Figure 11. In run F11, the baroclinic currents at different levels almost completely compensate, leading to a near-zero barotropic transport everywhere (top panels of Figure 11). At surface and subsurface depths shallower than 1000 m (Figure 11a), velocities are eastward and oppositely directed to the deeper westward flows between 1300 and 3600 m (Figure 11b). This westward flow through the Drake Passage is also seen at the bottom water depth (level 12, deeper than 3800 m, Figure 11c). The density difference between the two sides of the Drake Passage is small (Table 5).

In run F111p the submarine ridge, represented by the one-point topography, blocks part of the westward flow and changes the flow system substantially. Although currents at the surface and subsurface depths are very similar to those in run F11, at deeper depths the westward flow through the Drake Passage of the Weddell Sea deep water weakens (middle panels of Figures 11b and 11c). The Weddell Sea deep water that cannot flow westward through the Drake Passage flows northward and creates a significant eastward pressure gradient, as reflected in the density difference (Table 5). This pressure gradient is maintained by the continuous northward flow of the Weddell Sea deep water. It is these changes in the deep and bottom flows that lead to the substantial change in the magnitude of the vertically integrated ACC transport.

In run F11pp the effect of the submarine ridges is more substantial. The flows through the Drake Passage are eastward throughout the water column (bottom panels of Figure 11). All the Weddell Sea deep water flows northward, and a greater density difference is set up (Table 5).

The effect of the submarine ridge is not confined to the region near the Drake Passage. It also channels AABW into the North Atlantic where it lifts the NADW to a shallower depth, as can be seen from middle and bottom panels of Figures 11b and 11c.

The above results have confirmed that in the presence of bottom topography, there is a linkage between the magnitude of the ACC and the AABW formation rate and its outflow. A stronger Weddell Sea deep water outflow flowing northward against the modeled submarine ridges establishes a larger density difference and generates a larger ACC. This process may be interpreted as follows. The AABW formed in the Weddell and Ross Seas flows northward across and under the ocean basins. This flow must be compensated by an approximately equal southward flux in the overlying waters. A ring taken along a latitude circle through the Drake Passage of southward moving fluid will conserve its angular momentum about the Earth's axis if no external forces act on it. In this case, angular momentum conservation will cause the ring of fluid to accelerate in the eastward direction as its radius decreases. Alternatively, the acceleration of the ring may be regarded as being due to the Coriolis force acting on the southward motion. The compensating northward moving fluid (the AABW from the Weddell Sea, etc.) is blocked by the topography, so that there is a net eastward stress on the vertically integrated water column, represented by the form stress term in Table 5. This process has the potential to generate large velocities, so that factors causing drag must be important in restraining this current, in the same manner as for wind forcing. In the model this is accomplished by the lateral friction.

The process may be quantified by considering the behavior of a single layer of homogeneous fluid of uniform depth $D$ on a spherical Earth of radius $a$, which flows at a zonally uniform rate toward the pole with net flux $Q$ over a given range of latitudes. This gives meridional velocity $v = -Q/(2\pi a D \cos \varphi)$ where $\varphi$ is the latitude, directed toward the pole. One may then determine the resulting zonal velocity $u$, assuming that it vanishes at
Figure 11. Horizontal velocities at (a) level 6, (b) level 10, and (c) level 12, for runs (top) F1I, (middle) F1I1p, and (bottom) F1Ipp.

the latitudinal boundaries. We retain lateral viscosity \(A_m\) but omit bottom friction and wind stress. The equations are given in an Appendix. If the latitudinal boundaries are taken at 45°S and 65°S, the resulting velocity profile when \(\epsilon = Q/2\pi A_m D \ll 1\) is

\[
u = \frac{Q \alpha \Omega}{\pi A_m D} f(\varphi),
\]

where \(f(\varphi)\) is shown in Figure 12. For the parameter values used in this paper, with \(A_m = 9 \times 10^5 \text{ cm}^2 \text{ s}^{-1}\), and taking a total AABW formation rate \(Q\) of 30 Sv and a returning southward flow of depth 2000 m, we have \(\nu = -0.065 \text{ cm} \text{ s}^{-1}\) at 55°S and a maximum of \(\nu\) of 5.27 cm s\(^{-1}\) at 56°S. These values are consistent with the speeds obtained in the numerical model, which reinforces the notion that this contribution to the ACC is due to conservation of angular momentum restrained by lateral friction. We may also note that if we take values of \(A_m = 10^4 \text{ m}^2 \text{ s}^{-1}\), \(Q = 20\) Sv, \(D = 3000\) m, which are more realistic for the real ocean, we obtain \(\epsilon = 0.1, \nu = -0.03 \text{ cm} \text{ s}^{-1}\) at 55°S, and a maximum \(\nu\) of 210 cm s\(^{-1}\). With still smaller \(A_m\) values these velocities become enormous, and clearly, as for the wind forcing, drag forces additional to lateral friction must be operating in the real ocean.

The linkage between the ACC and Weddell Sea deep water also explains the feature of a larger thermohaline-induced ACC and a weaker wind-induced ACC with a
in run 11 the combined effect of bottom pressure drag in other places has reduced the ACC from that in run F11pp. While it may be possible to identify the locations where the reduction takes place, it would require countless experiments. Instead, we have conducted several experiments to test the sensitivity of the ACC to topography in other places. In an experiment with a submarine ridge of the same depth as in run F11Ip but at the eastern side of the Weddell Sea, an ACC of only 3 Sv is generated. In another experiment with the same ridge as in run F11Ip but at the western side of Ross Sea, an ACC of only 12 Sv is produced. It is reasonable to speculate that the proximity of the Weddell Sea to the narrow Drake Passage may be important, since in this narrow passage a bottom flow is more easily guided by the submarine ridge. For example, if the passage is wider, the bottom flow may tend to veer around the submarine ridge and still flow westward.

The sensitivity of the ACC transport to topography at different locations is not much affected by the east-west variation of the surface thermohaline forcing fields. We have repeated runs reported in this section but with zonally averaged SST and SSS fields, and the sensitivity of ACC does not change much.

To further test our hypothesis that the Weddell Sea deep water formation process contributes significantly to the size of the ACC, we have repeated run 11W and 11, but with the deep water formation in the Weddell Sea switched off by setting the relaxation salinity to 32.00 ppt over the Weddell Sea region (see Figure 13, the shaded area) from the observed value (typically 33.80 ppt). The deep water formation rate in either of these two runs is less than 2 Sv. These two runs are called 11WnoWDF and 11noWDF, respectively. The ACC in run 11noWDF reduces to 51.2 Sv from 80.2 Sv in run 11 (Figure 13a), and in run 11WnoWDF reduces to 87.4 Sv from 119.6 Sv in run 11W (Figure 13b). This dependence is supported by the work of Olbers et al. [1993].

7. Summary and Conclusions

In this paper, we have described the circulation driven by wind stress only, by surface thermohaline forcing alone, and by both wind and thermohaline forcing in a coarse-resolution global model. The central issue is how important the wind and thermohaline forcing are in determining the global ocean circulation. It was found that strong interactions take place between the circulations driven by the two forcing constituents and that such interactions contribute significantly to the present-day state of the oceanic circulation.

In the basic runs (runs W, 11, 11W, 21, and 21W), we first examined the pure wind-driven circulation (run W) in a homogeneous ocean with topography, the dynamics are strongly influenced by potential vorticity conservation, leading to weak barotropic transports (except the Kuroshio). Another feature is that the wind-driven overturning cells associated with Ekman divergence (equatorial and subpolar regions) and conver-
Figure 12. Eastward velocity profile function $f(\varphi)$ as a function of latitude ($-\theta - 90$) for flow due to conservation of zonal angular momentum of uniform poleward flow, restrained by lateral friction with $v = 0$ at 45°S and 65°S and $\epsilon \ll 1$.

Figure 13. Horizontal barotropic transport stream function (in sverdrups) for (a) run 1noWDF and (b) run 1WnoWDF (see text for the details of these runs). Contour interval is 10 Sv. Over the shaded area, the relaxation salinity is reduced so that the Weddell Sea deep water formation is switched off.

gence (subtropical regions) in the homogeneous ocean penetrate to the bottom of the ocean. In an ocean with topography driven by the thermohaline field alone (runs 1I and 2I), qualitative features of the major observed overturning cells (the NADW formation cell, the southern sinking cell, and the AABW formation cell), the poleward heat transport, air-sea heat exchange, and water mass property are all produced, although there is a strong dependence upon vertical mixing coefficient. Barotropic transports, in particular a substantial ACC, are generated by the bottom form stress as a result of incomplete compensation for the surface flows by the bottom flows. Since deep water flows are controlled by the deep water density field, which in turn is strongly influenced by the overturning circulation, the barotropic transport and the overturning circulation are “coupled.” In an ocean with topography driven by both thermohaline and wind forcing (runs 1W and 2W), interaction of wind- and thermohaline-driven circulations takes place, and the resultant circulation is not simply the superposition of the pure wind-driven and pure thermohaline-driven circulations. Because of the coupling, the wind-induced barotropic transports and overturning circulations are larger in the presence of a smaller vertical diffusivity.

The coupling between the overturning circulation and the barotropic transport in the ACC region is most interesting. We have shown that in the absence of wind forcing, stronger vertical mixing leads to a larger AABW and hence a greater ACC. An analysis of the momentum balance shows that in the presence of bottom topography and stratification, the dynamic balance
of the ACC is mainly between lateral friction and bottom form stress. It is these features that are responsible for the substantial ACC without wind forcing. We have also shown that the direct effect of wind forcing in generating the ACC is small, but the indirect effect of wind forcing can be significant. This indirect wind effect plays a role largely via the wind-driven southern subtropic-subpolar cell, that is, the Deacon cell. The upwellings associated with the Deacon cell intensify the vertical mixing and strengthen the AABW cell (and the Weddell Sea deep water formation rate) and hence the ACC. In the presence of wind forcing, the dynamic balance of the model ACC is between lateral friction, bottom form stress and wind forcing. Further, in the latitudes of the Drake Passage the net bottom form stress is in the same direction as the wind forcing term, and is much larger. In particular, wind-induced enhancement of the bottom form stress can be larger than the wind forcing term itself. This driving force is not present in most previous theories for the ACC [e.g., Munk and Palmén, 1951; Johnson and Bryden, 1989; Wolff et al., 1991; and Krupitsky and Cane, 1994].

The way in which the bottom density distribution generates the model ACC has been further investigated in a series of experiments. In a flat-bottomed ocean driven by the surface thermohaline field alone, the deep water formed in the Weddell Sea flows westward, and the ACC is virtually zero. However, the addition of a submarine ridge, represented by one model grid point immediately north of the tip of the modeled Antarctica Peninsula, prevents part of the Weddell Sea deep water from flowing westward, and establishes a deep water density difference. The pressure gradient associated with this density difference then generates an ACC of about 70 Sv (with $\Delta TV = 1 \text{ cm}^2 \text{s}^{-1}$). When the complete model submarine ridge along the Drake Passage is included, all the Weddell Sea bottom water flows northward against the submarine ridge, a larger density difference is set up, and the ACC reaches up to 110 Sv. These experiments suggest that the ACC is at least partly driven by the thermohaline circulation associated with the AABW formation process. This result has important implications. There has been debate about the size of the ACC. The most acceptable figure to date is 134 Sv, which is from the International Southern Ocean Studies program [Nowlin and Klink, 1986] and is a mean over 12 months but with 20% variability. Since the ACC is influenced by the rate of AABW formation, which may vary significantly with time, the magnitude of the ACC could be highly dependent upon the time an observation is taken.

As Munk and Palmén [1951] suggested, in the real ocean the wind- and buoyancy-driven forcing of the ACC are probably balanced by bottom drag rather than by lateral friction. Much of this drag is probably due to topography on horizontal scales that are too small to be resolved by this model, in which the principal drag on the current is due to the unrealistically large lateral friction term. However, we feel that the basic conclusions of this paper about the nature of the driving mechanisms for the ACC should still apply regardless of whether the drag on the current is attributed to topographic form drag or large lateral viscosity.

### Appendix: Zonally Symmetric Flow From a Line Source to a Line Sink on Latitude Circles

We consider a shallow layer of homogeneous fluid of uniform depth $D$ on a sphere rotating with angular velocity $\Omega$ ($\Omega > 0$) and adopt spherical polar coordinates $(r, \theta, -\lambda)$, where $\theta$ is the colatitude measured from the south pole. We also have $r = a$ (effectively), and $\lambda$ denotes longitude measured eastward. If $u$ and $v$ are the zonal and meridional velocity components measured in the directions of $\lambda$ and $\theta$ increasing respectively, then with lateral viscosity $A_m$ and $\partial / \partial \lambda \equiv 0$ the zonal momentum equation for $u$ is

$$
\begin{align*}
\frac{v}{a} \frac{\partial u}{\partial \theta} + \frac{u \cot \theta}{a} - 2\Omega \cos \theta &= \frac{A_m}{a^2 \sin^2 \theta} \frac{\partial}{\partial \theta} \left( \sin \theta \frac{\partial u}{\partial \theta} \right) - \frac{u}{a^2 \sin^2 \theta},
\end{align*}
$$

(A1)

We now specify uniform meridional flow $v$ with total flux $Q$ toward the south pole between colatitudes $\theta_1$ and $\theta_2$, where

$$
\begin{align*}
v = \frac{Q}{2\pi D \sin \theta},
\end{align*}
$$

(A2)

Substituting into (A1) then gives

$$
\begin{align*}
\frac{d}{d\theta} \left[ \sin \theta \frac{du}{d\theta} \right] + \epsilon \frac{du}{d\theta} + \left( \cot \theta - \frac{1}{\sin \theta} \right) u - c_2 \Omega \cos \theta = 0,
\end{align*}
$$

(A3)

where $\epsilon = Q / 2\pi A_m D$. If we require that the zonal velocity vanish at $\theta_1, \theta_2$, the boundary conditions are

$$
\begin{align*}
u = 0, \quad \theta = \theta_1, \theta_2.
\end{align*}
$$

(A4)

If $\epsilon \ll 1$, one may solve (A3) to leading order in $\epsilon$ by variation of parameters by substituting

$$
\begin{align*}
Y_i = \ln(\tan(\theta_i/2)), \quad i = 1, 2,
\end{align*}
$$

(A5)

to obtain

$$
\begin{align*}
u = -\frac{Q a \Omega}{\pi A_m D} \left[ A \sinh(Y - Y_1) + B \sinh(Y_2 - Y) - \cosh(Y - Y_1 - (\tanh(Y_2 - \tanh Y_1)) - \sinh Y \ln(\frac{\cosh Y_2}{\cosh Y_1}) \right],
\end{align*}
$$

(A6)

where

$$
\begin{align*}
A = \frac{Y_2 - Y_1 - (\tanh Y_2 - \tanh Y_1) \cosh Y_2}{\sinh(Y_2 - Y_1)},
\end{align*}
$$

and

$$
\begin{align*}
B = \ln(\frac{\cosh Y_2}{\cosh Y_1}) \frac{\sinh Y_1}{\sinh(Y_2 - Y_1)}.
\end{align*}
$$

Taking $\theta_1 = 25^\circ$ and $\theta_2 = 45^\circ$ with latitude $\varphi = \theta - 90^\circ$ gives the profile shown in Figure 12.

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References


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