Effects of a Continental Slope along the Western Boundary on the Abyssal Circulation

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To investigate effects of a continental slope along the western boundary on the abyssal circulation, numerical experiments using multi-level models were carried out. An ocean which extends over the northern and southern hemispheres is forced by cooling inside the ocean at the southwest corner of the basin and uniform heating through the sea surface. When the reference density for the cooling is vertically uniform, effects of the slope emerge clearly for the slope with considerably broad width. The deep western boundary current flowing over the slope feeds no bottom flows in the southern hemisphere, and carries the warmed deep water into the northern hemisphere. This leads to the increased meridional density gradient, which results in the modification of deep flow patterns. When the reference density is vertically distributed, the upper and lower northward flowing western boundary currents form in the deep layer. As the density stratification relaxes the topographic control, the westward intensification of the upper boundary current is achieved over the slope. The intensified flow is accompanied by the countercurrent and they form the horizontal recirculation over the slope. However, the effects are confined around the slope region and the interior flow patterns do not change. The lower boundary current is not significantly affected by the slope and has the large width with no countercurrent. It is found that the actual continental slope does not have significant effects on the gross feature of the thermohaline circulation.

1. Introduction

The ocean general circulation may be divided into the wind-driven and thermohaline circulations in terms of the driving forces. However, due to nonlinearity of the ocean system, superposition is not possible. The wind-driven circulation, in general, appears as ocean currents above the main thermocline, and the thermohaline circulation determines the deep circulation over the world ocean. For the study of the thermohaline circulation, numerical modelling has been utilized because of the importance of thermodynamical nonlinearity of the circulation, although long time scales inherent in the deep circulation could make the study very hard. In the recent years, developments of super-computers lead to promotion of the study of the thermohaline circulation. For example, Bryan (1987) demonstrated strong dependence of the circulation on the effects of vertical diffusivity in a general circulation model, i.e., with increasing diffusivity, the thermocline deepens and the circulation becomes stronger. Colin de Verdière (1988) demonstrated detailed dependence of the thermohaline circulation on the diffusivity within the context of planetary geostrophy. SuginoHara and Fukasawa (1988) (hereafter referred to as SF) investigated a set-up process of the deep circulation as the thermohaline circulation, paying attention to the formation process of the Stommel and Arons' deep circulation pattern (Stommel and Arons,
The present study deals with the thermohaline circulation, particularly, an abyssal circulation flowing over a bottom floor. It may be easily expected that the actual abyssal circulation is affected by the bottom topography, e.g., sea mounts, ridges, basins and trenches. In general, topographic effects on the abyssal current are investigated by the general circulation model for the world oceans (O-GCM). In the O-GCMs, complicated ocean circulations appear in association with introduction of the realistic bottom topography. However, due to the complexity in topography, it is difficult to understand their dynamics.

In this paper, we attempt to understand effects of a continental slope along the western boundary on the abyssal circulation as a first step. The present study is motivated by and is closely connected to recent numerical studies, namely, Suginoehara and Aoki (1991) and Suginoehara et al. (1992). Therefore, we give a brief review of these two studies here. They are the investigation of the thermohaline circulation by using multi-level numerical models. Effects of the bottom topography are not considered. Their model oceans are on a β-plane, and extend over the northern and southern hemispheres. But they are small in size compared with an actual ocean basin.

Suginoehara and Aoki (1991) (hereafter referred to as SA) examined a nature of a steady thermohaline circulation which is driven by differential cooling through the sea surface. They found the dependency of the effect of the vertical diffusivity on the steady thermohaline circulation in the thermodynamic balance. A set of alternating zonal jets (stacked jets) along the equator found in SF were reproduced, and details of the jets and associated meridional circulation were more clearly demonstrated. They concluded that the stacked jets is a realistic feature and is an essential nature of the thermohaline circulation.

Suginoehara et al. (1992) (hereafter referred to as SAO) made a detailed study of a circulation driven by the cooling inside the ocean (body cooling). As suggested by the experiment of Aoki et al. (1992), the body cooling easily controls the basic stratification. SAO demonstrated that when the reference density for the body cooling is vertically distributed, the meridional circulation has a double structure; reversal of the Stommel and Arons' pattern emerges in the deep layer well below the thermocline. Their result successfully reproduced the deep circulation pattern for the Pacific proposed by Fiadeiro (1982).

As another important finding in SAO, the deep western boundary layer has features of the diffusive regime in Masuda and Uehara (1992), where characteristics of the boundary layer are examined using a reduced gravity model. They classify the dynamics of the boundary layer into two regimes. One is the viscous regime, for which the boundary current accompanies the countercurrent and has the width of the Munk layer (Munk, 1950). The deep western boundary current that is seen in SF and SA is in this regime. The other is the diffusive regime, for which the boundary current is characterized by a broad width and no countercurrent. SAO explained the dynamics of the formation of the boundary layer of the diffusive regime in their model.

The reason we first consider the slope along the western boundary is that effects of the slope may lead to modification of the whole abyssal circulation, as the deep western boundary current forms over the slope and feeds the interior flow. Modification of the deep western boundary current due to effects of the slope may be expected from the study of SF. During the set-up process, coastally trapped wave-type density currents along the western boundary in the southern hemisphere may play an essential role in place of Kelvin wave-type density currents, and long Rossby wave-type density currents toward the western boundary are modified by the slope in both hemispheres. Further, it is interesting to know to what extent the boundary layer of the diffusive regime obtained in SAO is modified by effects of the slope.
We use a small basin model which has almost the same size as in SA, and perform several case studies. We take two cases for the cooling inside the ocean which drives the model ocean; first, the reference density for the cooling is vertically uniform and, secondly, is vertically distributed as in SAO. As the reference case for each case, experiments are carried out for the model with the flat bottom. Next, a simple geometry is considered, i.e. a continental slope along the western boundary and the flat bottom floor in the rest of the ocean. For actual ocean basins, the width of the continental slope is usually a hundred kilometers or so. Therefore, our main concern is finding the effects of such a narrow and steep slope. We also make one experiment with a broad slope for the case where the reference density is uniform. We discuss this case in detail to help understand the effects of the narrow slope.

2. Model

Figure 1 shows the model ocean we use. It extends over the southern and northern hemispheres on the spherical surface. The longitudinal width is 30° and the latitudinal width is 60°. For the reference experiments, the model ocean has a flat bottom. The slope is located along the western boundary below the depth 1500 m. The bottom floor is flat except the slope region and the depth there is 4000 m. We define the width of slope as a length between the coast, Long. 0° and the point at which the slope intersects the bottom floor. Main experiments with a slope are carried out taking 1.5° for the width (narrow slope case). We also make one experiment where the slope width is 5° (broad slope case).

We use a spherical coordinate system with longitude \( \lambda \), latitude \( \phi \) and height \( z \) (measured positive upward) relative to the earth’s mean radius, \( a \). Let \( u \), \( v \) and \( w \) be the components of velocity in the \( \lambda \), \( \phi \) and \( z \) directions, respectively, \( P \) pressure and \( \rho \) density. The equations of motion and the hydrostatic relation under the Boussinesq approximation are

\[
\frac{\partial u}{\partial t} + \mathcal{L}(u) - f v = -\frac{1}{\rho_0 a \cos \phi} \frac{\partial P}{\partial \lambda} + A_w \left[ \nabla^2 u - \left( \frac{1 + \tan^2 \phi}{a^2} \right) u - \frac{2 \sin \phi}{a^2 \cos^2 \phi} \frac{\partial v}{\partial \lambda} \right] + A_v \frac{\partial^2 u}{\partial z^2},
\]

![Schematic view of the model ocean.](image)
\[
\frac{\partial v}{\partial t} + L(v) - fu = -\frac{1}{\rho_0 a} \frac{\partial P}{\partial \phi} + A_H \left[ \nabla^2 v - \left( 1 + \tan^2 \phi \right) v + \frac{2 \sin \phi}{a^2 \cos^2 \phi} \frac{\partial u}{\partial \lambda} \right] + A_v \frac{\partial^2 v}{\partial z^2},
\]

\[
\frac{\partial P}{\partial z} = -\rho g,
\]

and the equation of continuity is

\[
\frac{1}{a \cos \phi} \frac{\partial u}{\partial \lambda} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (v \cos \phi) + \frac{\partial w}{\partial z} = 0,
\]

where \(A_v\) and \(A_H\) are the coefficients of vertical and horizontal eddy viscosity, respectively, \(\rho_0\) the mean density over the whole depth and \(f\) the Coriolis parameter defined on the sphere. The advection operator \(L(\cdot)\) is defined as

\[
L(\sigma) = \frac{u}{a \cos \phi} \frac{\partial \sigma}{\partial \lambda} + \frac{v}{a \cos \phi} \frac{\partial}{\partial \phi} (\sigma \cos \phi) + w \frac{\partial \sigma}{\partial z},
\]

and the horizontal Laplacian \(\nabla^2\) is given by

\[
\nabla^2 \sigma = \frac{1}{a^2 \cos^2 \phi} \frac{\partial^2 \sigma}{\partial \lambda^2} + \frac{1}{a^2 \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\partial \sigma}{\partial \phi} \right).
\]

The equation of density under the assumption that density and temperature have a linear relation is

\[
\frac{\partial \rho}{\partial t} + L(\rho) = K_H \nabla^2 \rho + K_v \frac{\partial^2 \rho}{\partial z^2} + S,
\]

where \(K_v\) and \(K_H\) are the coefficients of vertical and horizontal eddy diffusivity, respectively. Effects of salinity are not considered. The term, \(S\) represents a cold (dense) water formation inside the ocean as given below.

Heating and cooling to drive a thermohaline circulation are made as in SAO. The mass source, \(S\) is imposed inside the ocean in the southern part of the southern hemisphere,

\[
\frac{\mu(\lambda, \phi)}{\rho_0 C_v} \left( \rho_c(z) - \rho \right),
\]
where \( C_v \) is the specific heat at constant volume, \( \rho_c \) the reference density to which the sea water density is damped. The coefficient, \( \mu(\lambda, \phi) \) is distributed as shown in Fig. 2, and the maximum value, \( 10^{-2} \text{ cal} (\degree \text{C cm}^3 \text{day}^{-1}) \) (100 days in the damping time) exists around the southwest corner of the southern hemisphere. Figures 3(a) and 3(b) show the distributions of \( \rho_c(z) \). For the case where the reference density for the cooling is vertically uniform, the cold water is formed at depths deeper than 650 m in the southern part of the southern hemisphere and \( \rho_c(z) \) has 28.0 in \( \sigma_t \) units (Fig. 3(a)). For the case where the reference density is vertically distributed, the cold water formation is imposed on the total column, i.e. from the sea surface through the bottom floor, in the southern part of the southern hemisphere. The distribution of the reference density is taken to be the same as in SAO. From the sea surface to the depth 650 m, \( \rho_c(z) \) increases rapidly from 24.0 to 27.0 in \( \sigma_t \) units, and in the deep layer \( \rho_c(z) \) changes slowly to 28.0 (Fig. 3(b)). In addition to the cold water formation inside the ocean for each case, the density flux

\[
\frac{\gamma_w(\lambda)}{\rho_0 C_v} (\rho_c(z) - \rho)
\]

is imposed through the southern boundary. The maximum value of the coefficient, \( \gamma_w(\lambda) \) is \( 3.3 \times 10^3 \text{ cal} (\degree \text{C cm}^2 \text{day}^{-1}) \) (33.3 days in the damping time) between the western boundary and Long. 3.5\degree. The value then decreases linearly until it reaches zero at the eastern boundary. At the sea

Fig. 2. Horizontal distribution of the coefficient, \( \mu(\lambda, \phi) \). The contour interval is \( 10^{-3} \text{ cal} (\degree \text{C cm}^3 \text{day}^{-1}) \). The maximum value lies around the hatched southwest corner and is \( 10^{-2} \text{ cal} (\degree \text{C cm}^3 \text{day}^{-1}) \).
Fig. 3. Vertical distribution of the reference density, $\rho_c(z)$ when it is uniform (a) and distributed (b).

surface, the density flux

$$\frac{\gamma_s}{\rho_0 C_v} (\rho_s - \rho)$$

is imposed, where $\gamma_s$ has the constant value, 500 cal ($^\circ$C cm$^2$ day)$^{-1}$ (100 days in the damping time). The reference density for heating through the sea surface, $\rho_s$ is uniform and the value of it is 24.0 in $\sigma_t$ units.

The wind stress at the sea surface is not taken into consideration. Friction is imposed at the bottom (see SF). The side walls are the no-slip boundaries. On the bottom floor and the side walls except the southern boundary the normal gradient of density is zero, i.e., there is no density flux across these boundaries. As the initial condition, a weakly stratified ocean is considered to suppress density currents with large amplitude as discussed in SF.

The numerical model used in this study is the GFDL ocean general circulation model (Cox, 1984), except that the weighted upcurrent scheme is adopted for the density advection term (Suginohara et al., 1991). It is remarked that the vector $(u, v)$ point is located on the equator in the present model though the scalar $(\rho)$ point is so in the model of SF, SA and SAO. To accelerate the calculation and to reach a steady state efficiently, Bryan's accelerating method (Bryan, 1984) is adopted except in the initial transient stage. For the horizontal directions, the grid interval is taken to be 1.0$^\circ$ in longitude and latitude but 0.5$^\circ$ for the narrow slope case. There are twelve levels in the vertical direction as shown in Fig. 4. Calculations are carried out until a thermally and dynamically steady balance is established for each case. The following values are used for the numerical calculations: $K_H = 10^7$ cm$^2$s$^{-1}$, $A_H = 8 \times 10^7$ cm$^2$s$^{-1}$ and $K_V = A_V = 1.5$ cm$^2$s$^{-1}$.
3. Result

3.1 Case where reference density is vertically uniform

For the case where the bottom is flat, horizontal velocity fields (a) and velocity and density fields along meridional (b) and zonal (c) sections are shown in Figs. 5(a), (b) and (c), respectively. As seen in the meridional section of the density field (Fig. 5(b)), the deep layer is occupied by the homogeneous water formed in the cooling region in the southern hemisphere. It is remarked that the thermocline penetrates below the depth of 650 m where the top of the cold water formation is situated, as pointed out by SA. In the deep layer, the Stommel and Arons' pattern for the Pacific is formed (see the depth 3750 m in Fig. 5(a)); the northward flowing deep western boundary current feeds the interior flows crossing the equator, and in the interior region an eastward flow with the poleward component is accompanied by upwelling (see the velocity field in Fig. 5(b)) following vorticity balance, $\beta v = f \omega/\partial z$. As this circulation is the internal mode motion, the circulation is reversed in the surface layer above the thermocline (see the depth 175 m in Fig. 5(a)). The deep western boundary current has the characteristics of the viscous regime as mentioned in SA; it accompanies a countercurrent and an intense vertical motion, and has uniform width and thickness throughout its path (see the zonal velocity field in Fig. 5(b) and the meridional velocity in Fig. 5(c)). The stagnation point of the deep western boundary current in the northern hemisphere is located at a fixed latitude (about $15^\circ$N) throughout the deep layer. Along the equator concentrated zonal flows are found in the deep layer and form a higher vertical mode structure (see the zonal velocity in Figs. 5(b) and 5(c)), which is the stacked jets discussed in SF and SA.

The circulation and density fields are almost the same as those in SAO, although, in SAO, the model ocean is on a $\beta$-plane and is longitudinal wider, and the cooling region is horizontally
Fig. 5(a)

Fig. 5. Horizontal velocity fields (a) and meridional circulation, zonal flow and density fields along the meridional section (b) and zonal circulation, meridional flow and density fields along the zonal sections (c), when the reference density is vertically uniform for the flat bottom case. For the zonal flow, the contour interval is 0.2 cm s$^{-1}$ and the shaded areas indicate the westward flow. For the meridional flow, the contour interval is 1.0 cm s$^{-1}$ and the shaded areas indicate the westward flow. For the density, the contour interval is 0.2 in $\sigma_z$ units for the solid lines and 0.002 for the dashed lines for greater than 27.97 $\sigma_z$. Numerals on contours are subtracted by 20.0.
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Fig. 5(c)

Fig. 5(b)
Fig. 6. Same as Fig. 5 but for the broad slope case.
confined. As for the stacked jets, the different arrangement of grid points and the coarse vertical resolution in the deep layer in the present model lead to insignificant differences.

Next, we discuss the broad slope case. Figures 6(a), (b) and (c) show horizontal velocity fields (a) and velocity and density fields along meridional (b) and zonal (c) sections, respectively, for the case where the slope has the width of 5°. In the horizontal velocity field in the deep layer, remarkable differences from the flat bottom case are seen as follows. In the interior region in the deeper part of the deep layer (see the depth 3750 m in Fig. 6(a)), a weak westward flow dominates in the southern hemisphere, but, on the other hand, a strong eastward flow with the northward component from the western boundary current dominates in the northern hemisphere. In the upper part of the deep layer (see the depth 2250 m in Fig. 6(a)), a strong eastward flow is found in the southern hemisphere. Breakdown of the symmetry of zonal flows with respect to the equator is more clearly seen in the central meridional sections (Fig. 6(b)). In the deep layer a meridional density gradient in the interior region is considerably strengthened compared with the flat bottom case (compare Fig. 6(b) with Fig. 5(b)). This is associated with vertical shear of zonal flows following the thermal wind balance, \( \tau \delta / \tau z = g \delta \rho / \delta \phi \). Consequently, in the southern hemisphere, the flow at bottom levels becomes westward and the eastward flow in the upper part of the deep layer is strengthened (see the zonal velocity in Fig. 6(b)).

Remarkable differences from the flat bottom case are also seen in the deep western boundary current. The detailed structures are shown in the zonal sections (Fig. 6(c)). In the southern hemisphere the western boundary current upwells along the slope (see the velocity field at 10°S in Fig. 6(c)) and feeds the interior flows at the intermediate depths of the deep layer (see the zonal velocity in Fig. 6(b)); the interior region at the bottom levels seems to stagnate. The strong downward flow at the western boundary in the equatorial region and in the northern hemisphere (see the velocity field at EQ. and 10°N in Fig. 6(c)) carries the warm (light) water into the deeper part of the deep layer, which is clearly seen in the density distribution at 10°N in Fig. 6(c). This warm (light) deep water extends to the interior in the northern hemisphere, and this causes the large meridional density gradient.

Figure 7 shows a series of zonal sections for the meridional velocity near the western boundary, from north of the cooling region to the northern hemisphere. At 22°S, just north of the cooling region, the deep western boundary current, the core of which is located off the slope, starts to flow being accompanied by a countercurrent i.e., the boundary layer is in the viscous regime. As it approaches the equator, the core shifts westward and rides on the slope. At the same time, the lower part of the current loses the countercurrent and gets a large width (see also the meridional velocity at 10°S in Fig. 6(c)), i.e. the feature of the diffusive regime. When it crosses the equator, the countercurrent appears again. Then the upwelled water descends, and the western boundary current loses its strength quickly and the countercurrent as well.

The behavior of the deep western boundary current over the slope described above is explained in the context of one layer barotropic model; because of uniformity of density in the deep layer below the thermocline and constancy of the upper level of the deep western boundary current, the deep layer can be treated as one homogeneous layer. Then, we can define the ambient potential vorticity, \( fH \) (Fig. 8), where \( H \) is taken to be thickness between the bottom floor and the center of the thermocline, the depth of 650 m. The deep western boundary current starting from the cooling region tends to flow along the contour of ambient potential vorticity; it rides on the slope in the southern hemisphere and goes down the slope in the northern hemisphere, which causes upwelling and downwelling in the western boundary layer. These are just what happens in the broad slope case.
Fig. 7. Meridional flow along zonal sections, when the reference density is vertically uniform for the broad slope case. The western half of the basin is plotted. The contour interval is 1.0 cm s$^{-1}$ and the shaded areas indicate the southward flow.
Fig. 8. Horizontal distribution of ambient potential vorticity, \( f/H \) for the broad slope case. The contour interval is \( 5 \times 10^{-10} \text{ (cm s)}^{-1} \).

A vertically integrated horizontal mass transports must occur as the northward flowing deep western boundary current and the southward flowing western boundary current in the surface layer are, contrary to the flat bottom case, not stacked vertically. In general, it is expected that over the slope region a large transport should be induced by the bottom pressure torque (Holland, 1973). In fact, an anticlockwise circulation over the slope takes 20 Sv in mass transport.

Warming in the deep layer in the northern hemisphere for the broad slope case is found in the density distribution and the density balance at the bottom levels. In the flat bottom case, density is larger in the western region (Fig. 9(a)); the horizontal advection cools the northern bottom levels counterbalancing the vertical diffusion (Fig. 9(b)). On the contrary, in the broad slope case, warm (light) water exists around the western boundary (Fig. 10(a)); the horizontal advection heats the northern bottom levels and, with the vertical diffusion as heating, balances the horizontal diffusion (Fig. 10(b)). On the other hand, at the bottom levels in the southern hemisphere, cooling by the horizontal diffusion is dominant as far as \( 10^\circ \text{S} \) because the cooling region is largely relative to the basin size. Considering the fact that the southern bottom region is not fed by the deep western boundary current and is stagnant, it is concluded that the southern bottom levels in the broad slope case are, essentially, cooled by the horizontal diffusion from the southern formation region.

In a case where the slope has the width of \( 1.5^\circ \), velocity and density fields along meridional (a) and zonal (b) sections are shown in Figs. 11(a) and (b), respectively. At a glance, differences from the flat bottom case are indistinguishable. However, careful comparison with the flat
Fig. 9. Horizontal distribution of density at the depth of 3750 m (a) and terms in the density equation along the meridional section at 15° at the depth of 3750 m (b), when the reference density is vertically uniform for the flat bottom case. For the density, the contour interval is 0.001 in \( \sigma_z \) units. The hatched areas indicate greater than 27.995 \( \sigma_z \). For terms in the density equation, HA and HD are for the horizontal advection and diffusion terms and VA and VD for the vertical advection and diffusion terms, respectively. The unit of the originate is \( \sigma_z \) s\(^{-1}\).
Fig. 10. Same as Fig. 9 but for the broad slope case. Numerals on contours are subtracted by 20.0.
Fig. 11. Meridional circulation, zonal flow and density fields along the meridional section (a) and zonal circulation, meridional flow and density fields along the zonal sections (b), when the reference density is vertically uniform for the narrow slope case. The contour intervals are as in Fig. 5.
bottom case tells that the characteristics pointed out in the broad slope case are recognized (compare Fig. 11 with Fig. 5). Another effect of the slope can be seen as follows. The core of the deep western boundary current which feeds the stacked jets stands shallower over the slope in the equatorial zone. Considering the discussion in SF and SA, this implies that the deepest eastward jet increases its thickness. Indeed, comparison between the flat bottom case and the narrow slope case clearly shows that the top of the deepest eastward flow along the equator is shallower in the slope case (see the zonal velocity in Figs. 5(b) and 11(a)). But, for the broad slope case, it is not proper to discuss the thickness of the deepest flow, because the vertical shear of zonal flows at the bottom levels distorts the shape of the stacked jets (see Fig. 6(b)).

A degree of slope effects is found in the meridional stream function (Fig. 12). A cell pattern

Fig. 12. Stream function of the meridional circulation, when the reference density is vertically uniform for the flat bottom case (top), the narrow slope case (middle) and the broad slope case (bottom). Units are $10^{12}$ cm$^3$ s$^{-1}$.
is more distorted as the slope width is broader. The upward transport at the bottom levels in the southern hemisphere for the broad slope case reflects the upwelling of the western boundary current along the slope. In the narrow slope case, a slight upward transport can be seen. In the case where the reference density is vertically uniform, the deep western boundary current has, for the most part, the characteristics of the viscous regime. This width is that of the Munk layer and is about 130 km under the present parameters. To what extent the slope affects the behavior of the deep western boundary current may depend on the slope width relative to the width of the boundary current. Indeed, in the broad slope case, the slope is sufficiently wider than the boundary current, and significant topographic effects emerge. On the other hand, in the narrow slope case, the slope width is much the same as the Munk layer width, and the result shows small topographic effects.

3.2 Case where reference density is vertically distributed

For the flat bottom case, horizontal velocity fields (a) and velocity and density fields along meridional (b) and zonal (c) sections are shown in Figs. 13(a), (b) and (c), respectively. Basic features are well compared with the results of SAO, though there are differences in the flow fields in the southern hemisphere due to the large cooling region of this model. At the bottom levels (see the depth 3750 m in Fig. 13(a)), the Stommel and Arons' pattern is formed in the northern hemisphere, but the southern hemisphere is dominated by a westward flow which feeds the western boundary current. At the intermediate depths of the deep layer (see the depth 2250 m in Fig. 13(a)), the horizontal circulation is reversed. This contrasts with the case where the reference density is vertically uniform, where the reversal occurs in the upper part of the deep layer (see the velocity field in Fig. 5(b)). As seen in the central meridional section for the velocity field (Fig. 13(b)), the upwelling velocity has its maximum at the depth 2750 m in the northern hemisphere, which indicates the reversal of the Stommel and Arons' pattern above this depth following vorticity balance, $\beta v = f\omega/\partial z$. In the upper part of the deep layer (see the depth 1250 m in Fig. 13(a)), the horizontal flows are reversed again, i.e. the same circulation as at the bottom levels.

The deep western boundary current also shows the same structure as in SAO. Just north of the cooling region (see the meridional velocity at 20°S in Fig. 13(c)), the western boundary current has the first baroclinic mode structure and accompanies a countercurrent. As it approaches the equator (see 10°S in Fig. 13(c)), the third mode motion dominates. The upper northward flowing current is weaker than the lower one. At the bottom levels the broad western boundary current accompanied by no countercurrent is formed; this is the boundary layer of the diffusive regime as pointed out by SAO. Though the countercurrent appears at the bottom levels in the equatorial region (see Eq. in Fig. 13(c)), the western boundary current keeps this modal structure as north as the stagnation point (see 10°N in Fig. 13(c)). Farther north, the first baroclinic mode structure appears again (see 20°N in Fig. 13(c)). Details of the deep western boundary current are fully discussed in SAO.

For the narrow slope case, horizontal velocity fields (a) and velocity and density fields along meridional (b) and zonal (c) sections are shown in Figs. 14(a), (b) and (c), respectively. In the horizontal velocity field in the deeper part of the deep layer (see the depths 3750 m and 2250 m in Fig. 14(a)), the difference from the flat bottom case (Fig. 13) is indistinguishable. However, a striking feature emerges around the western boundary region in the upper part of the deep layer (see the depth 1250 m in Fig. 14(a)); the southward flowing countercurrent develops in the northern and southern hemispheres. It seems that together with the northward flowing western boundary current it forms the horizontal recirculation over the slope. But, the interior region
Fig. 13. Same as Fig. 5 when the reference density is vertically distributed for the flat bottom case. For the density, the contour interval is 0.2 in $\sigma$ units for the solid lines and 0.02 for the dashed lines for greater than 27.8 $\sigma$. 

Fig. 13(a)
Fig. 14. Same as Fig. 13 but for the narrow slope case.
remains almost the same as in the flat bottom case. The zonal sections of the meridional velocity (Fig. 14(c)) clearly show the detailed structure of the western boundary current. We can easily notice that the upper northward flow in the western boundary current is considerably intensified, and the lower one is slightly weaker than in the flat bottom case, though the gross modal structure is not changed (compare Fig. 14(c) with Fig. 13(c)). This intensification begins just north of the cooling region (see 20°S in Fig. 14(c)). Farther north in the southern hemisphere (see 10°S in Fig. 14(c)), this upper northward flow lies over the top of the slope and accompanies a countercurrent. And, the northward and southward flows in the western boundary current are not vertically stacked. This structure continues as north as the stagnation point in the northern hemisphere (see 10°N in Fig. 14(c)).

The behavior of the deep western boundary current described above is explained when we consider effects of density stratification. In the slope case, the southward flowing boundary current in the surface layer and the upper northward flow in the deep western boundary current apparently form the first baroclinic mode structure over the slope, where the thickness between the isopycnal and the bottom floor is small (see Fig. 14(c)). The intensified upper northward flow is associated with the strong western boundary current in the surface layer as the first baroclinic mode near the coast. The formation of such a first mode structure is permitted only under the existence of the density stratification. As demonstrated in the case where the reference density is vertically uniform, the western boundary currents are controlled by the bottom topography, and the core of the deep western boundary current does not reach the western end of the basin (see Fig. 7). However, in the case where the reference density is vertically distributed,
the density stratification formed in the deep layer (see the density field in Fig. 14(c)) relaxes the control of the bottom topography. Thus, the westward intensification of the upper northward flow over the slope becomes possible in this case. This intensified upper northward flow has the characteristics of the viscous regime, and consequently it accompanies the countercurrent.

Surprising in spite of considerable changes in the structure of the western boundary current, differences of interior fields between the flat bottom case and the slope case are not large as seen in Figs. 13 and 14. The enhanced northward transport due to the intensified upper northward flow in the western boundary layer is compensated by the countercurrent, and consequently no effects seem to escape into the interior. At the same time, as discussed in SF, the interior fields are set up by a Rossby wave-type density current extending westward from the eastern boundary where the topographic effects do not exist. This results in the identical pattern of the meridional velocity in the interior region between the flat bottom case and the slope case (compare Fig. 14(c) with Fig. 13(c)). On the other hand, the lower northward flowing current in the western boundary does not feel the bottom effects which are discussed in the case where the reference density is vertically uniform, and the structure of the boundary layer of the diffusive regime in the flat bottom case is not distorted.

As shown in the meridional stream function pattern (Fig. 15), the double structure which is characterized by a return flow at mid-depths can be seen in the flat bottom case. In the slope case, the double structure becomes more distinguishable; two cells are more clearly separated. This separation reflects the structure of the western boundary current; the intensification of the upper northward flow and the development of the intermediate southward flow.

4. Discussion and Conclusion

We have carried out numerical experiments on the abyssal circulation using multi-level models with bottom topography. Comparing with the case of the flat bottom, effects of the continental slope along the western boundary have been investigated. The thermohaline circulation is driven by two types of the body cooling, one is that the reference density is vertically uniform and the other is that the reference density is vertically distributed.

In the case where the reference density is vertically uniform, significant topographic effects emerge when the slope is sufficiently broader than the actual slope. The deep western boundary current flowing along the ambient potential vorticity contour over the slope feeds no interior flows at the bottom levels in the southern hemisphere and carries the warmed water into the interior region in the northern hemisphere. The southern bottom levels get cooled directly by horizontal diffusion from the cooling region. This forms the increased meridional gradient of density in the deep layer, which results in the breakdown of the symmetrical distribution of zonal flows about the equator in the flat bottom case. Effects of the slope depend on its width relative to that of the boundary layer of the viscous regime, i.e. that of the Munk layer. When the slope corresponds to the actual continental slope, the slope width is the same order as that of the Munk layer, and topographic effects are not so large.

In the case where the reference density is vertically distributed, the reversal of the Stommel and Arons' pattern well below the thermocline takes place as studied in SAO. Effects of the slope lead to intensification of the upper northward flowing deep western boundary current, which is fairly weaker than the lower one in the flat bottom case. This flow accompanies the countercurrent and forms the horizontal recirculation over the slope. It should be noted that the effects of the actual slope are confined to the western boundary region and the interior is not affected. Even if the slope is broad, significant effects of the slope on the interior region can not be expected for
the following reason. The core of the lower northward flowing current in the western boundary tends to flow along the contour of the ambient potential vorticity over the slope like the case where the reference density is vertically uniform. But, the density stratification in the deep layer presses down the top of the current. Therefore, the slope is, virtually, narrow for the current, and the topographic effects should not be so large.

As for the details of the deep western boundary current, its great width is found in observations (e.g., Stommel et al., 1973). Warren (1976) and recently Masuda and Uehara (1992) proposed the concept of the boundary layer of the diffusive regime which sufficiently explains the large width. And SAO reproduced this boundary layer in their numerical model where the reference density for the cooling is vertically distributed considering the stratification in the deep layer of the Pacific. Furthermore, the present study demonstrated that the continental slope does not change the boundary layer at the bottom levels for the case where the reference density is distributed as in SAO. Therefore, it is safe to say that in the South Pacific where no broad slope exists, the deep western boundary current which carries the bottom water is in the diffusive regime. We should be careful, however, in the case of the South Atlantic, where the broad deep western boundary current flows over the continental rise whose width is as broad as 1000 km. Stommel and Arons (1972) argued that the continental rise has a "slope-broadening" effect on the deep western boundary current. They considered the deep current flowing northward on the slope and conserving the potential vorticity. To simplify the conservation relation, they took the current to flow from the source of uniform potential vorticity and the fluid outside of the flowing portion to be at rest. But, it seems difficult to apply the setting of their model on the numerical models, when considering the result of the present model. Anyway, the problem in the South Atlantic is left for future study.

In conclusion, we point out that the actual continental slope is not expected to have significant effects on the gross feature of the thermohaline circulation.

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References