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Geothermal heating and its influence on the meridional overturning circulation

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Abstract. The effect of geothermal heating on the meridional overturning circulation is examined using an idealized, coarse-resolution ocean general circulation model. This heating is parameterized as a spatially uniform heat flux of 50 mW m^{-2} through the (flat) ocean floor, in contrast with previous studies that have considered regional circulation changes caused by an isolated hot spot or a series of plumes along the Mid-Atlantic Ridge. In our model results the equilibrated response is largely advective: a deep perturbation of the meridional overturning cell on the order of several sverdrups is produced, connecting with an upper level circulation at high latitudes, allowing the additional heat to be released to the atmosphere. Rising motion in the perturbation deep cell is concentrated near the equator. The upward penetration of this cell is limited by the thermocline, analogous to the role of the stratosphere in limiting the upward penetration of convective plumes in the atmosphere. The magnitude of the advective response is inversely proportional to the deep stratification; with a weaker background meridional overturning circulation and a less stratified abyss the overturning maximum of the perturbation deep cell is increased. This advective response also cools the low-latitude thermocline. The qualitative behavior is similar in both a single-hemisphere and a double-hemisphere configuration. In summary, the anomalous circulation driven by geothermal fluxes is more substantial than previously thought. We are able to understand the structure and strength of the response in the idealized geometry and further extend these ideas to explain the results of *Adcroft et al.* [2001], where the impact of geothermal heating was examined using a global configuration.

1. Introduction

As the radioactive elements in our Earth's interior decay over geologic time, heat is lost through the lithosphere. Approximately 70% of this heat flux is absorbed into the abyssal oceans and marginal basins [*Sclater et al.*, 1980]. Prior work has argued that both local hot spots (forming rising plumes) along the mid-ocean ridge axes [e.g., *Stommel*, 1982] and the more pervasive diffusive heating away from the ridge systems [e.g., *Thompson and Johnson*, 1996] play important roles in the abyssal circulation. Nonetheless, to the best of our knowledge the weaker widespread geothermal heating at the ocean bottom has been omitted from all previously published ocean general circulation model results (the only exception is *Adcroft et al.* [2001] using a global ocean model). In this study, we examine the impact of geothermal heating in the context of an idealized ocean geometry to understand better the structure and scaling of the circulation response.

To begin, let us compare the estimated 32 TW (1 TW = 10^{12} W) of geothermal heating through the ocean floor [*Stein et al.*, 1995] with other important energy fluxes in the climate system. The meridional oceanic heat flux is ~2000 TW [*Macdonald and Wunsch*, 1996], and the solar flux of energy at the ocean surface is greater still by an order of magnitude. Presumably, the contrast between these numbers underpins modelers' justification for the omission of the geothermal flux. However, as first noted by *Sandström* [1908], the net heating of the oceans occurs in the tropics, at high geopotential, and is therefore unable to drive directly a strong large-scale overturning circulation. More recently, *Huang* [1999] used a simple tube model of the ocean circulation to suggest that surface heating can only produce a weak circulation that is rate-limited by diffusive mixing, as first suggested by *Jeffreys* [1925], whereas deep heating can produce a very strong circulation, rate limited by friction. Thus it is not clear a priori that even a weak geothermal energy flux is necessarily negligible.

Until recently, nearly all attention has focused on the dynamics of plumes at mid-ocean ridges. Hydrothermal venting at high temperatures (roughly 350°C) can produce steady state plumes or transient thermals, also known as "megaplumes" or event plumes. A typical black smoker can release 60 MW, whereas event plumes are thought to produce a heat flux several orders of magnitude greater [*Lupton*, 1995]. These local sources entrain surrounding waters as they rise to a level of neutral buoyancy, or spreading level. The amount of entrained water is not negligible, owing to the disparity between the ambient and venting water temperatures; *Kadko et al.* [1995] estimated that the total vertical transport may

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be as large as 12 Sv, i.e., the same order of magnitude as the meridional overturning circulation. The influence of rotation causes entraining fluid to circulate cyclonically, whereas the flow is anticyclonic in the spreading phase (see *Helfrich and Speer* [1995] for a review of hydrothermal convective plume growth). In the absence of background flow, the β effect would propagate tracers and temperature perturbations westward [*Stommel*, 1982]. Evidence of this westward propagation has been observed at depth [*Lupton and Craig*, 1981]. *Joyce and Speer* [1987] showed that if a background *Stommel and Arons* [1960] abyssal flow is included (directed north and eastward), characteristics originating from a plume may tend westward or eastward, depending on the ratio of the background flow velocity to the long baroclinic Rossby wave phase speed [see also *Speer*, 1989].

Although less remarkable for their local dynamical effect, low temperature venting occurs at both mid-ocean ridge axes and along ridge flanks. Ocean crust that is older than about 50–70 million years does not produce significant hydrothermal circulations [*Stein et al.*, 1995], yet continued cooling and radiogenic decay maintain a background heat flux of 38–46 mW m^{-2} [*Sclater et al.*, 1980]. What is perhaps most notable from a thermodynamic perspective is the division of heat flux into the ocean. *Stein et al.* [1995] estimated 34 ± 12% is released through hydrothermal heat flux, and of this, only 29% occur directly along the ridge axes. *Joyce et al.* [1986] employed a scaling argument to show that the weak background heat flux has a negligible effect on the local potential vorticity balance. However, in this study we will consider the nonlocal effect of the background heating through the ocean floor, which we will show has a measurable effect on the abyssal circulation.

In keeping with the exploratory nature of this paper our general circulation model is highly idealized, as described in section 2. This paper extends the *Adcroft et al.* [2001] study, where a similar geothermal forcing was employed in a realistic global configuration, by examining the transient response and by performing several sensitivity runs. Our simpler configuration facilitates an improved dynamical understanding of the geothermal response. In section 3 we begin with a perturbation analysis of the flow and temperature structure that results from geothermal heating in a single-hemisphere ocean. A dynamical interpretation of these results is provided. In section 4 the sensitivity of these results to different model forcings is examined. Several experiments in a double-hemisphere ocean basin are presented in section 5, with a more explicit comparison to the results by *Adcroft et al.* [2001]. We conclude with some discussion of what these results suggest about the dynamics of the model ocean and some thoughts on how these results might apply in the real ocean.

2. Model Description

We employ the z coordinate, primitive equation model MOM2 (beta version 2.0), as described by *Pacanowski* [1996]. The ocean configuration and forcings for our single-hemisphere runs are those by *Marotzke* [1997] and J. R. Scott and J. Marotzke (The location of diapycnal mixing and the meridional overturning circulation, submitted to *Journal of Physical Oceanography*, 2001; hereinafter referred to as SM), with the addition of geothermal heating as represented by a uniform heat flux of 50 mW m^{-2} through the ocean floor. The domain is a 60° wide sector, ranging from the equator to 64°, with a constant depth of 4500 m. Surface temperature and salinity are relaxed toward an identical, zonally uniform cosine profile with peak-to-peak amplitudes of 27°C and 1.5 psu, respectively, using a 30 day relaxation time constant. For simplicity, no wind stress is imposed. Horizontal resolution is 1.875° zonally by 2° meridionally using 30 vertical levels ranging from 50 m at the surface to 250 m at the lowest level. At this resolution it is not possible to resolve

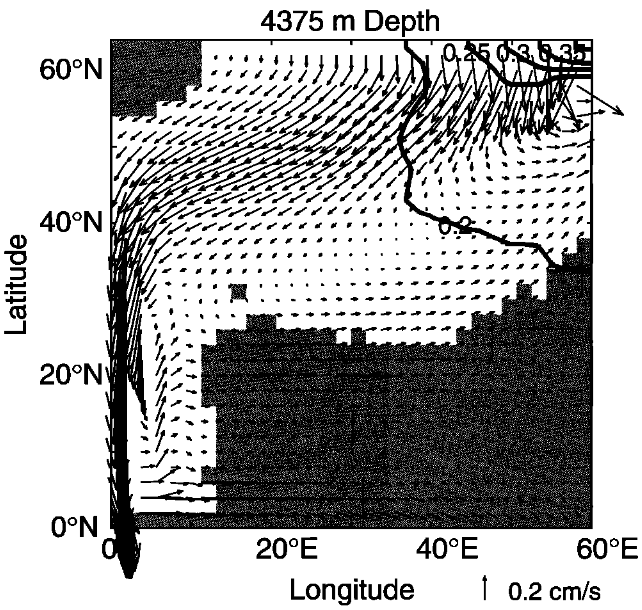

Figure 1. Plan view of bottom model layer 50 years after uniform geothermal heating is first applied. Flow is indicated by vectors and is representative of the background state flow as the model's advective response requires on the order of centuries to develop. Shading indicates grid cells undergoing convective mixing (i.e., cooling these locations). As compared to the background state prior to geothermal heating, the deep convective region in the northwest is increased in area, and all convection in the tropics is new.

dynamically active convective plumes, so the effect of local geothermal hot spots is ignored.

In our model runs we will assume that diapycnal mixing rates are determined through a prescribed diffusivity, investigating the role of deep heating rather than the likelihood that geothermal heating leads to measurable deep mixing (the latter being a more difficult problem, in our opinion). As by SM, diapycnal mixing is imposed in the columns adjacent to the north, south, east, and west side walls and is set to zero elsewhere; this is thought to mimic the effect of enhanced mixing due to a sloping lateral boundary. Diapycnal mixing at the equator is a surrogate for the global integral of mixing throughout the rest of the world's oceans. Unless specified otherwise, boundary diapycnal diffusivity is set to 10×10^{-4} m^2 s^{-1}, which produces an area-averaged diffusivity of 1.15×10^{-4} m^2 s^{-1}. For comparative purposes we also ran our single-hemisphere geothermal configuration with the more traditional approach of uniform diffusivity (using this area-averaged value), which showed only small quantitative differences (see SM for further discussion of boundary versus uniform mixing).

In section 5 a second hemisphere is added to the model so that the northern and southern basins are identical. In this configuration we impose our boundary mixing scheme as above but do not parameterize diapycnal mixing along the equator (as there is no longer a boundary located there).

For computational efficiency some of the experiments were run at half resolution, i.e., 3.75° × 4° and 16 vertical levels. In these runs the diapycnal mixing coefficient along the boundaries was decreased 50% for comparative purposes with the standard runs, and the horizontal viscosity was changed to resolve the Munk boundary layer at the new zonal grid spacing. There was no qualitative difference in the geothermal response between similarly configured runs at different resolution.

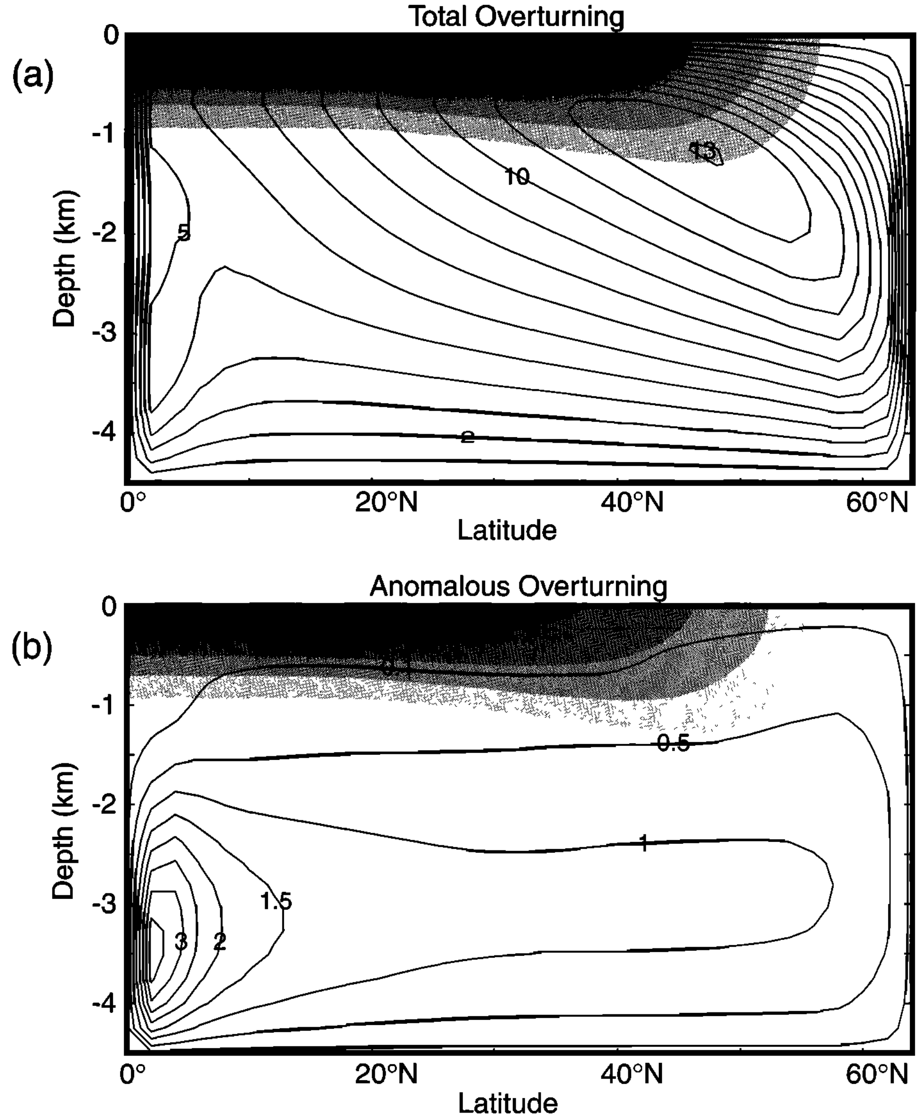

Figure 2. Meridional overturning stream function (contours) and zonally averaged temperature (shading). (a) Steady state run with uniform geothermal heating; (b) same as (a) except showing anomalous overturning as compared to a similarly configured run without geothermal heating. Contours are labeled in sverdrups, with solid contours indicating clockwise flow; isotherms are 1.35°, 2.7°, 5.4°, 10.8°, and 21.6° (0.05, 0.1, 0.2, 0.4, and 0.8 of ΔT = 27°C).

All model runs were integrated to equilibrium, as defined by a basin-averaged surface heat flux within 5 mW m^{-2} of the area-integrated geothermal heat flux and/or when overturning is discernibly within 0.1 Sv of its final value. All experiments with geothermal heating were spun up from a similarly configured steady state run without applied geothermal heating. Unless specified otherwise, geothermal heating was implemented via a uniform heat flux through the bottom boundary. We did not parameterize any hydrothermally generated salinity flux through the bottom boundary.

3. Single-Hemisphere Geothermal Heating

3.1. Preliminaries

Before proceeding to the model results, consider a scaling analysis of a simple uniform diffusive response to the geothermal heating. In steady state balance the extra heat input through the bottom boundary must eventually be released to the atmosphere through the surface heat flux. Given our surface restoring timescale of 30 days and a mixed layer depth of 50 m, a sea surface temperature increase of only 0.006°C is required to "radiate" away an additional 50 mW m^{-2}. The vertical temperature structure in this diffusive model can be deduced by balancing the additional deep heat input with the diffusion of an anomalous temperature gradient [see also *Adcroft et al.*, 2001]:

$$c_p \rho_o \kappa_v \partial_z T' \approx c_p \rho_o \kappa_v T'/H = 50 \mathrm{mW\,m}^{-2} \qquad (1)$$

We find that for an ocean depth H of 4500 m the top-to-bottom anomalous temperature difference is 0.5°C (assuming $\kappa_v = 1.15 \times 10^{-4}$ m^2 s^{-1}, our area-averaged vertical diffusivity). Our calculation of T' at the surface implies that virtually all of this temperature difference occurs in the abyss. In other words, the presence of geothermal heating leads to a leading order change in the oceans' deep water properties. In light of this simple calculation it is

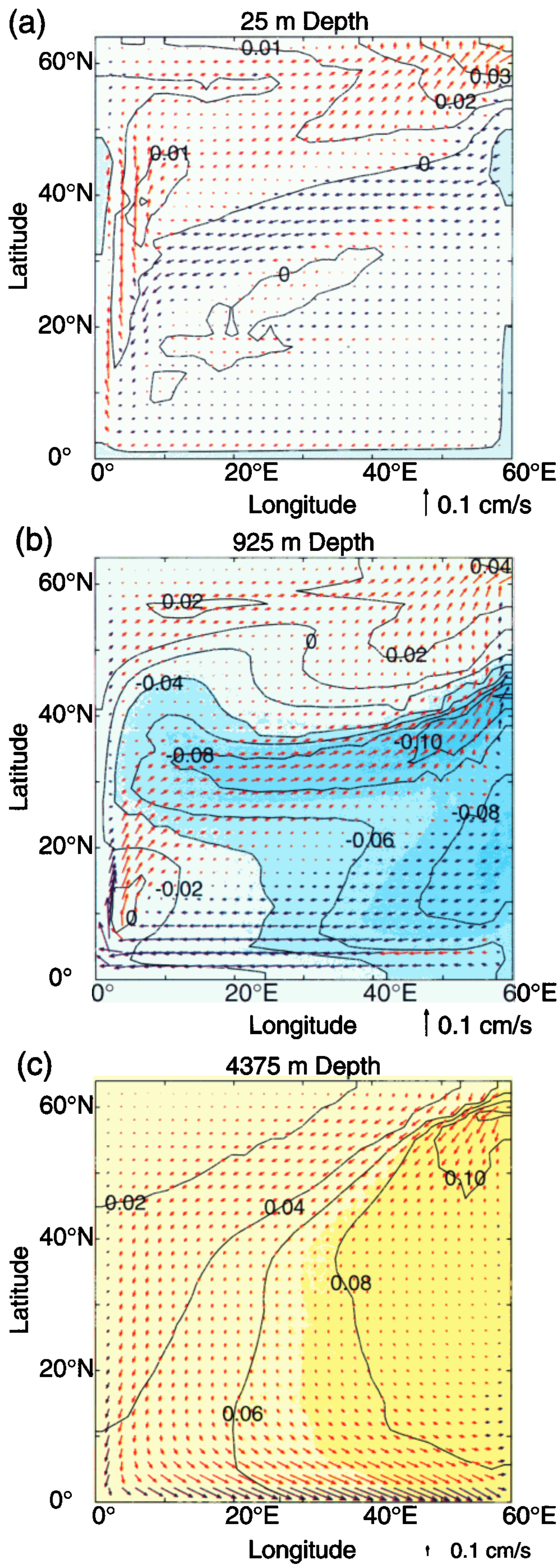

Plate 1. Plan view of the anomalous circulation and anomalous temperature (contours and color shading) obtained by subtracting the geothermal run results from the background state: (a) surface model layer (contour interval 0.01°C), (b) layer near the base of the thermocline (contour interval 0.02°C), and (c) lowest model layer (contour interval 0.02°C). Blue vectors indicate anomalous upwelling, and red arrows indicate anomalous downwelling.

somewhat surprising that the effect of geothermal heating has not received more attention. An important caveat, however, is that this analysis presumes that a purely diffusive response is dynamically possible; on a rapidly rotating planet, even small changes in the temperature structure produce an advective response through geostrophic balance.

3.2. Initial Transient Response

In response to uniform geothermal heating of 50 mW m^{-2}, convective mixing in the lowest model layers is triggered in the tropics, except near the deep western boundary current (Figure 1). Convection occurs here because the flow is sluggish and near the terminus of the abyssal flow pathway and has therefore been exposed to warming from the bottom boundary longer than more newly injected deep water. Since the abyssal waters are very weakly stratified, even weak heating from below is sufficient to initiate convective adjustment. Continued geothermal heating leads to increased convective penetration, effectively distributing the boundary heat flux over an expanding column of water. Analogous to a convective plume in the tropical atmosphere, however, penetration is ultimately limited by an area of sharply increased stratification. In the atmosphere this increase in stratification is delineated by the tropopause, whereas the analog in the ocean is the base of the thermocline. Present-day geothermal heat fluxes are too small to allow for any significant penetration of convective mixing into an existing thermocline.

On the timescale of a few hundred years a perturbation circulation develops in response to changes in the deep temperature structure. At steady state, flow through the abyss is increased, and tropical deep convective mixing is no longer triggered from below. We will not attempt to further diagnose the spin-up of this circulation but will instead focus on the steady state response. This transient behavior provides a qualitative link between the local hydrothermal plume models and our steady state geothermal response, as the lack of maintained convection from below in the latter might otherwise obscure this connection.

3.3. Perturbation Analysis

In Figure 2a we show the steady state meridional overturning circulation and zonally averaged thermal structure of our standard run. Compared to an identically configured run without geothermal heating (see SM's Figure 2a), the main meridional overturning circulation is similar, and its maximum near 50°N at 1 km depth is increased by only 0.4 Sv. However, the response at low latitudes is much stronger (Figure 2b): a deep meridional overturning cell is produced, with an overturning maximum of 3.8 Sv. All rising motion in the zonal mean occurs alongside the equator; net sinking occurs at all other latitudes but is concentrated primarily in the tropics and at high latitudes. Only 0.1 Sv flows through the thermocline.

The perturbation in surface temperature and flow is shown in Plate 1a. Anomalous warming occurs primarily in the northeast corner, where in the background state, surface waters downwell to the lowest layers of the abyss. The maximum temperature anomaly is 0.04°C, which is still small in comparison to the ~27°C difference between the equator and pole but is nearly an order of magnitude larger than the response estimated from scaling (section

3.1). Except for an area near the western boundary current much of the ocean surface is actually slightly colder (anomalous temperature of <0.01°C) in the geothermal simulation.

This structure in the surface perturbation temperature is accompanied by slight changes in surface flow patterns. Given a background state of nearly zonal flow into the eastern boundary (in the absence of applied wind forcing), the flow vectors indicate less downwelling at the eastern boundary between 40° and 50°N, with increased downwelling farther north. Closing this loop requires anomalous meridional flow, advecting warmer waters northward that in turn produce the patch of anomalous warming. This alteration in eastern boundary flow leads to a midlatitude cold tongue near the base of the thermocline, as observed in Plate 1b. The background flow into the eastern boundary advects warm water downward, so the cold tongue is a direct result of less downwelling at these latitudes. Perhaps surprisingly (i.e., in view of the fact that our forcing is heating the ocean), the tropics are also slightly colder in the geothermal run, reflecting weak anomalous upwelling into the thermocline. Given that the tropics are sharply stratified, any anomalous flow that continues upward into the thermocline results in anomalous cooling. Note, however, that the perturbation flow in the thermocline is over an order of magnitude smaller than the background flow, so the impact of geothermal heating on the upper ocean circulation is quite small.

In contrast, the magnitude of the perturbation flow in the abyss is significant in the context of the relatively sluggish background flow. A comparison of the perturbation flow pattern in Plate 1c and background flow vectors (see Figure 1, which shows the deep flow prior to any significant advective response) suggests that geothermal heating serves to augment the existing deep flow, except perhaps in the tropics where the background flow is more zonal. In the background state, flow is injected into the northeast as the warmest water in the abyssal layer. This flow is subsequently modified by the spread of dense water mass properties (through the *Gent and McWilliams* [1990] parameterization) from the northwest corner and finally reheated by diffusion from above along the boundaries in the tropics (see SM for a comprehensive discussion of the abyssal heat balance). Geothermal heating along this flow pathway produces anomalous warming throughout the bottom layer. Plate 1c indicates an increase in the volume of downwelling water into the northeast, where a maximum temperature anomaly of 0.12°C occurs. In the northwest corner, however, the anomalous warming is much weaker. Here the convective connection to the surface effectively transports any anomalous heat out of the abyss, creating an anomalous zonal temperature gradient at high latitudes. In the tropics, water is warmed as it flows eastward, completing the abyssal pathway of the large-scale overturning circulation; this heating helps to maintain an anomalous zonal temperature gradient, albeit weaker than at higher latitudes. We will subsequently show that this zonal gradient is not localized in the bottom layer and that the vertical structure of the east-west temperature difference is an essential feature of the dynamical response. The vigorous anomalous flow into the equator is consistent with the upwelling noted in our description of the overturning cell.

The steady state dynamical response invites several questions. Why is the anomalous upwelling narrow and the downwelling broadly distributed? Why does geothermal heating produce an anomalous meridional circulation at all? In the remainder of section 3 we will show that the anomalous deep circulation is ultimately governed by the requirement that the geothermal heat flux is exported out of the abyss to the surface, where it can be "radiated away" to the atmosphere through our restoring boundary condition.

3.4. Thermodynamic Interpretation of Response

In steady state, geothermal heat input into the ocean's bottom layer must be balanced by a compensating heat loss through

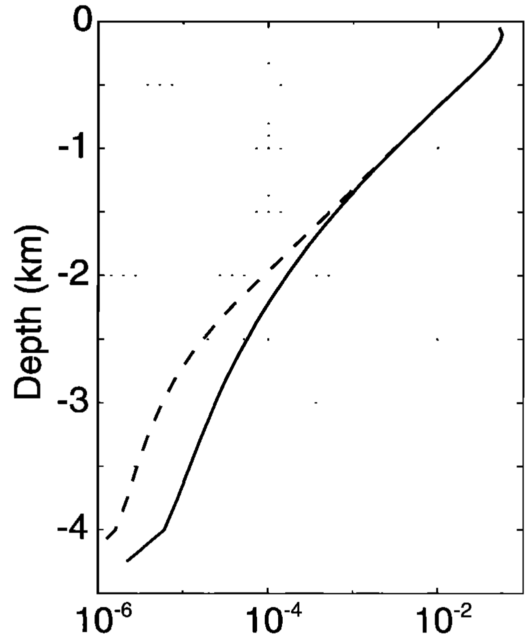

Figure 3. Plot of vertical potential temperature structure $\partial T/\partial z$ (in °C m^{-1}) versus depth along the equator at 60°E. The geothermal run results are represented by the dashed curve, and the solid curve shows the background state.

other processes. The most direct route of heat loss is through convective exchange to the surface. However, this mechanism is quite ineffective at removing heat input not local to the site of deep convection. Because deep convection occurs at high latitudes, it is near the beginning of the abyssal flow pathway; hence only brief exposure to geothermal heating has occurred. Moreover, given that the coldest water in the bottom layer occurs at the deep convection site [*Marotzke and Scott*, 1999], geostrophic dynamics diverts flow around the deep convection site, not through it. Thus any anomalous convective heat loss is quite limited, even if it were possible to effect a large-scale change in the deep circulation, i.e., rerouting the abyssal flow pathway back to high latitudes after extended exposure to geothermal heating. Penetrative convective mixing from below, as observed in the tropics during the transient response, can effectively distribute the geothermal heat across the abyss but cannot penetrate the thermocline. Hence any convective heat exchange to the surface is prohibited at low latitudes.

Weaker diffusive heat flow into the bottom layer constitutes a second conceivable means by which heat balance can be achieved. The decrease in stratification on the eastern boundary (Figure 3; there is little change in the western boundary stratification due to the rapid flow) indeed leads to smaller diffusive heat flux in the steady state geothermal run. However, a rough estimate of the heat flux from the term $c_p\rho_o\kappa_v\partial_z T$ suggests that even in the nongeothermal run, the area-averaged deep diffusive flux is 5 mW m^{-2}, about an order of magnitude smaller than the geothermal flux. Hence any decrease in downward diffusive fluxes through weakened stratification is relatively modest.

A third process through which heat can be transported from the abyss is vertical advection. The advective heat gain is proportional to the following expression:

$$\int wTdA = \int_{\text{downwelling}} w_D T_D + \int_{\text{upwelling}} w_U T_U \qquad (2)$$

where A is the area of the bottom layer and the U and D subscripts refer to upwelling and downwelling, respectively (so that w is the

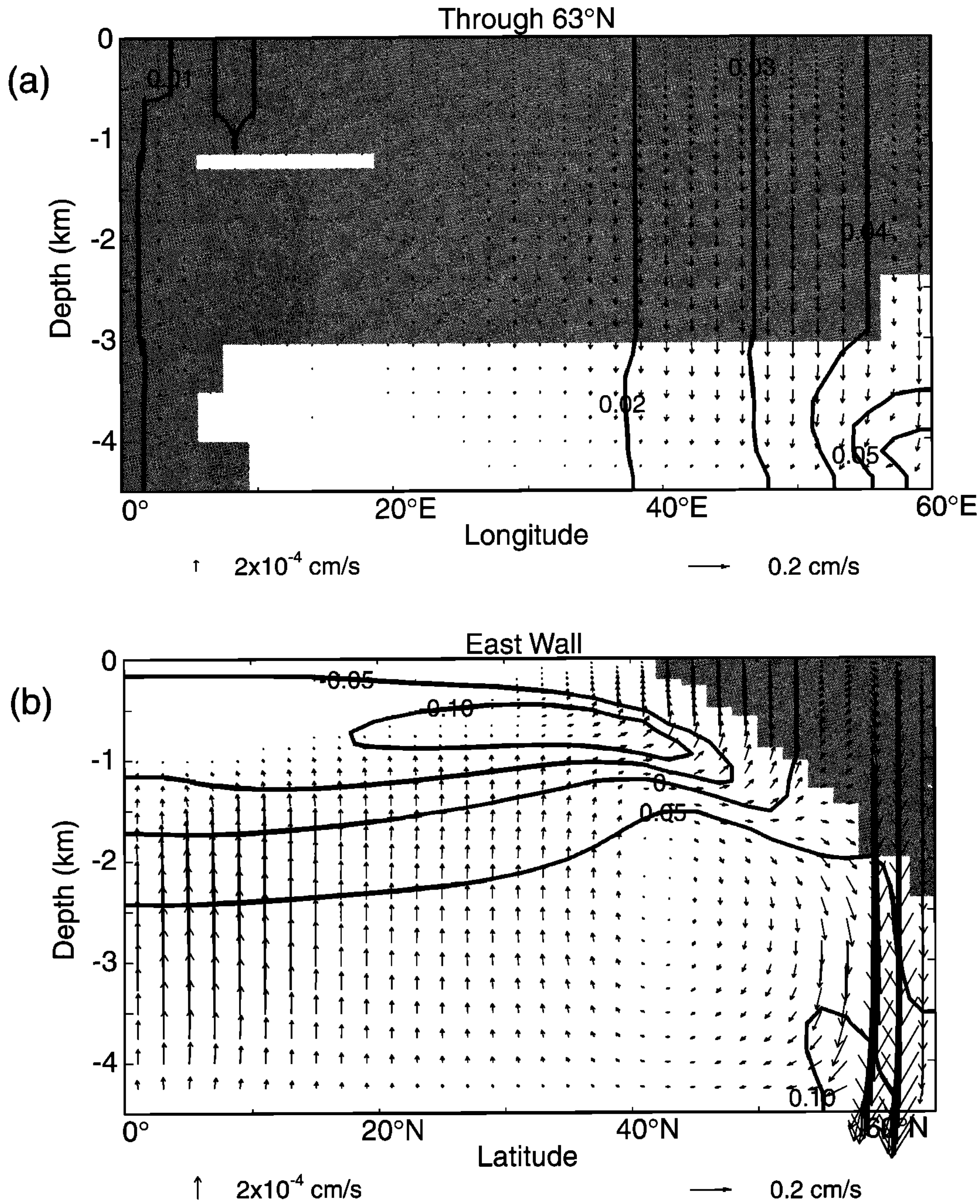

Figure 4. Plot of anomalous circulation and anomalous temperature (contours) in (a) the zonal plane against the model's back wall and (b) the meridional plane adjacent to the eastern boundary. Shading represents grid cells with active convective mixing.

magnitude of the vertical velocity). Prior to geothermal heating, (2) is negative (i.e., downward) because relatively warm water is injected into the bottom layer but upwelling occurs at colder temperatures (SM). With geothermal heating the advective response in the upper ocean leads to greater downwelling in the northeast corner, conveying water that is anomalously warm into the abyss (i.e., the $\overline{w}_D T'_D$ and $w'_D \overline{T}_D$ terms in the heat budget are both negative quantities, with the overbars and primes indicating background and anomalous quantities, respectively). Since the anomalous temperature is a greater at the downwelling site than in areas of upwelling, changes in (2) do not directly account for heat loss required for balance. Rather, the advective change results in a larger temperature difference between the properties of deep downwelling and deep convection (i.e., the zonal gradient at high latitudes), so the mesoscale eddy parameterization is more effective at spreading cold water from the northwest corner, which also serves to transport enthalpy out of the bottom layer.

The two arguments presented above, namely that water flows around the deep convection site and that background diapycnal heat fluxes are weak, also apply to the lower abyss, implying that an advective response across middepth must occur. Above middepth the component of the anomalous circulation through the high latitudes is critical in communicating the geothermal heating to the surface. Anomalous flow into the northeast is not limited to the

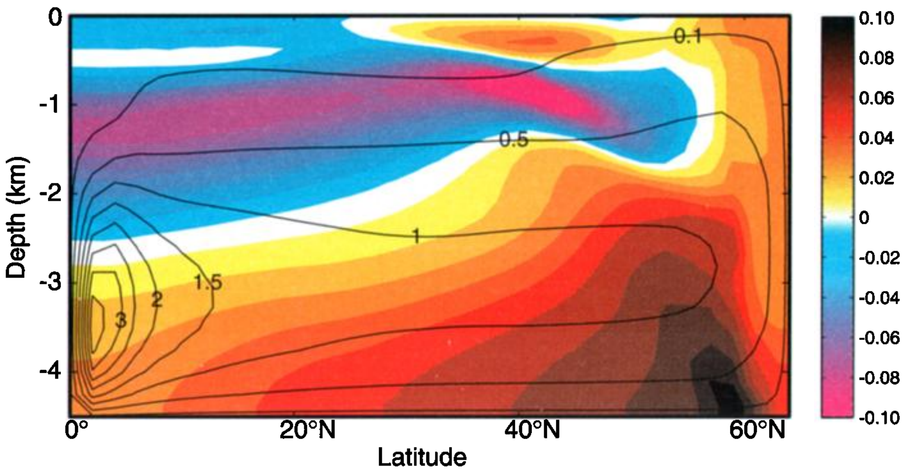

Plate 2. Contour plot of $\Delta T'$, the anomalous temperature difference between the eastern and western boundaries, overlaid by the anomalous meridional overturning circulation. Circulation contours are labeled in sverdrups, with solid contours indicating clockwise flow.

surface layer but occurs throughout the top 2500 m. Because deep mixed boundary layers occur along the northern and eastern walls, as shown in Figures 4a and 4b, respectively, anomalous heat fluxes into these convecting columns can be communicated quickly to the surface. Without the efficient mixing process in these deep mixed layers a more significant advective surface perturbation would be required.

3.5. Dynamics of Anomalous Meridional Overturning Circulation

In thermal wind balance the strength of the meridional overturning circulation is a function of the density difference between the eastern and western walls [*Marotzke*, 1997]. We will use this balance to obtain an order of magnitude estimate of the perturbation deep cell, based on the increase in temperature as a function of the zonal transit time across the tropics, which is in turn a function of the background large-scale circulation. Given an approximate deep zonal velocity of 0.1 cm s^{-1}, it takes water roughly 200 years to cross the basin. Our applied geothermal heat flux can heat a 3000 m column of water by 0.03°C during this transit time [*Joyce et al.*, 1986]. This value roughly agrees with the model's anomalous temperature difference between the east and west walls in the tropics, as shown in Plate 2 (it is more difficult to predict the boundary density difference at high latitudes a priori). The density difference is related to an overturning rate through thermal wind:

$$\frac{\Delta v'}{\Delta z/2} = \frac{g}{f\rho_0}\frac{\Delta\rho'}{\Delta x}, \tag{3}$$

where $\Delta\rho'$ is the anomalous east-west density difference and Δz is the depth of the abyss (i.e., so that $\Delta z/2$ is approximately the depth of the southward branch of the anomalous cell). If we assume that the meridional velocity increases linearly from maximum southward flow at the ocean bottom and vanishes at the center of the

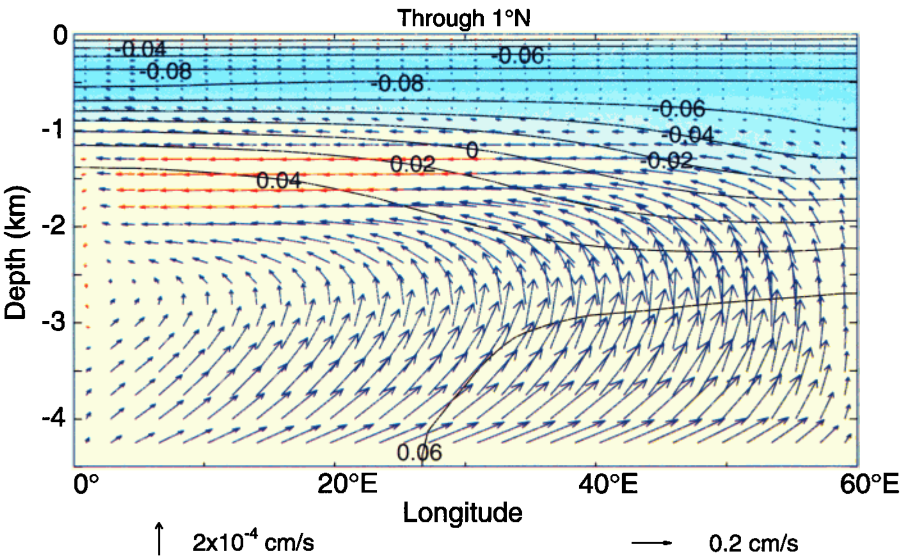

Plate 3. Plot of anomalous circulation and anomalous temperature in the zonal plane adjacent to the equatorial boundary. Blue vectors indicate anomalous upwelling, and red arrows indicate anomalous downwelling.

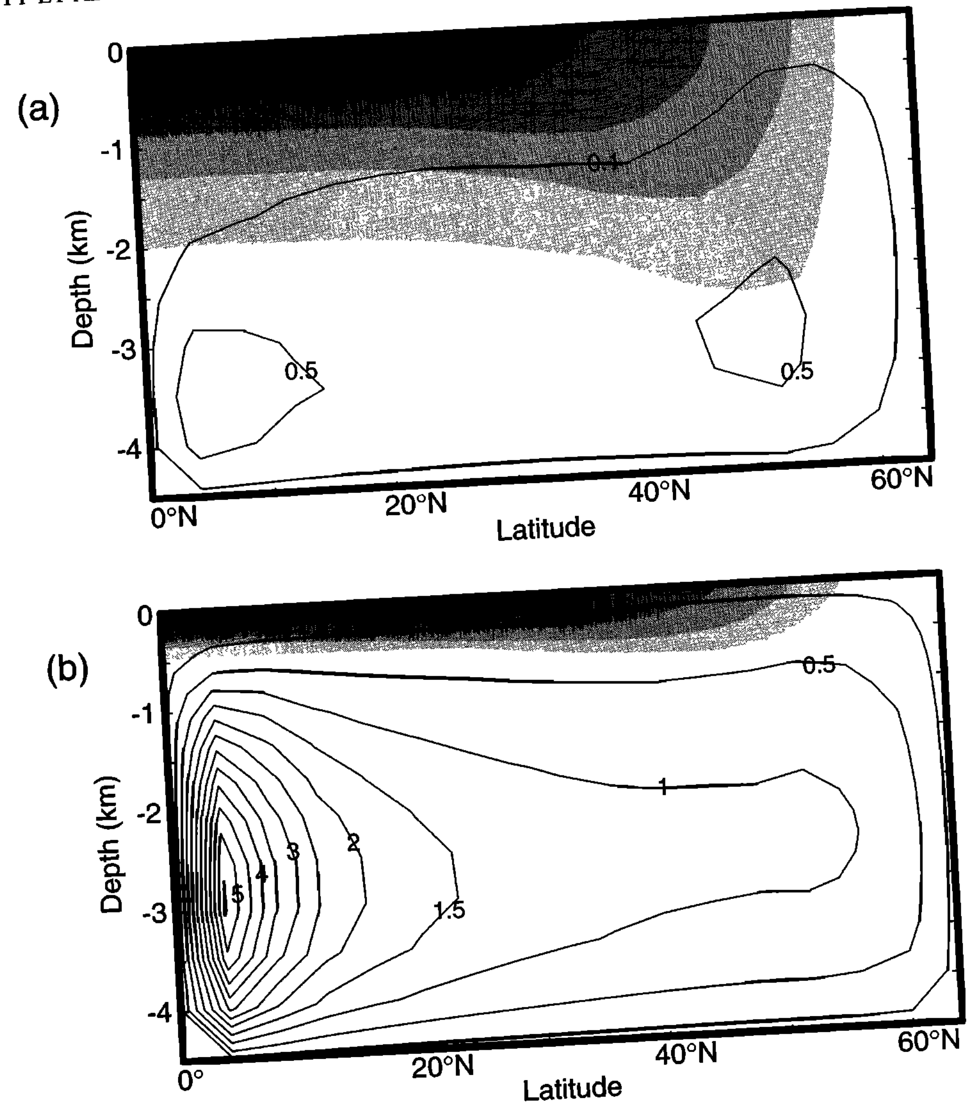

Figure 5. Anomalous meridional overturning circulation and zonally averaged temperature (shading) given (a) a 400% increase in boundary mixing strength and (b) an 80% decrease in boundary mixing strength. Contours are labeled in sverdrups, with solid contours indicating clockwise flow; isotherms are 1.35°, 2.7°, 5.4°, 10.8°, and 21.6°C.

anomalous cell (i.e., so that the average southward velocity is $\Delta v'/2$), the estimate of overturning strength is

$$\Psi' \approx \frac{\Delta v'}{2}\Delta x\frac{\Delta z}{2} = \frac{\alpha g}{8f}\Delta T'\Delta z^2, \qquad (4)$$

where $\alpha = 1.8 \times 10^{-4}$ K^{-1}, the thermal expansion coefficient. If we take $f \sim 10^{-5}$ s^{-1}, the value of the Coriolis parameter at 4°N (corresponding with the peak in anomalous overturning, as observed in Figure 2), our estimate for the anomalous overturning is 6 Sv. This estimate is somewhat too large but is the correct order of magnitude. The estimate is more accurate at higher latitudes. For example, at 45°N, $\Delta T' \sim 0.07$°C and $f \sim 10^{-4}$ s^{-1}; hence the estimated overturning strength is slightly over 1 Sv, in general agreement with Figure 2b. At the equator, use of (4) is problematic as the geostrophic relation breaks down.

We have been careful to obtain our a priori estimate of anomalous overturning maximum based on the southward branch of the deep cell. As seen in Plate 2, $\Delta T'$ varies considerably in the northward branch of the cell, i.e., between 1500 and 3000 m depth. In fact, $\Delta T'$ changes sign near the top of the anomalous cell, providing southward shear. This reversal is necessary or else the anomalous northward velocities would continue through the thermocline. As shown in Plate 3, this reversal is accomplished by cooling at the top of the anomalous cell where upwelling occurs, primarily the eastern side of the basin.

In geophysical fluid systems both stratification and rotation can influence the dynamical behavior by restricting vertical motions [e.g., *Cushman-Roisin*, 1994]. Given our estimate for the background deep zonal flow, a typical Rossby number U/fL in the model solution is $O(10^{-5})$. This number is quite small even within a few degrees of the equator, suggesting that rotational terms dominate the momentum equation throughout the basin. Using a representative model value of N (2×10^{-3} s^{-1}), we estimate the internal Froude number U/NH as $O(10^{-3})$. Hence, although the stratification is weak in the abyss, it is also sufficient to restrict vertical motion. Although both processes are clearly important, the ratio of the Rossby number to Froude number suggests that flow is mostly influenced by rotation. This analysis is consistent with the observed geothermal response in that limited vertical motion does occur, but all upwelling occurs adjacent to the equator, where f is as

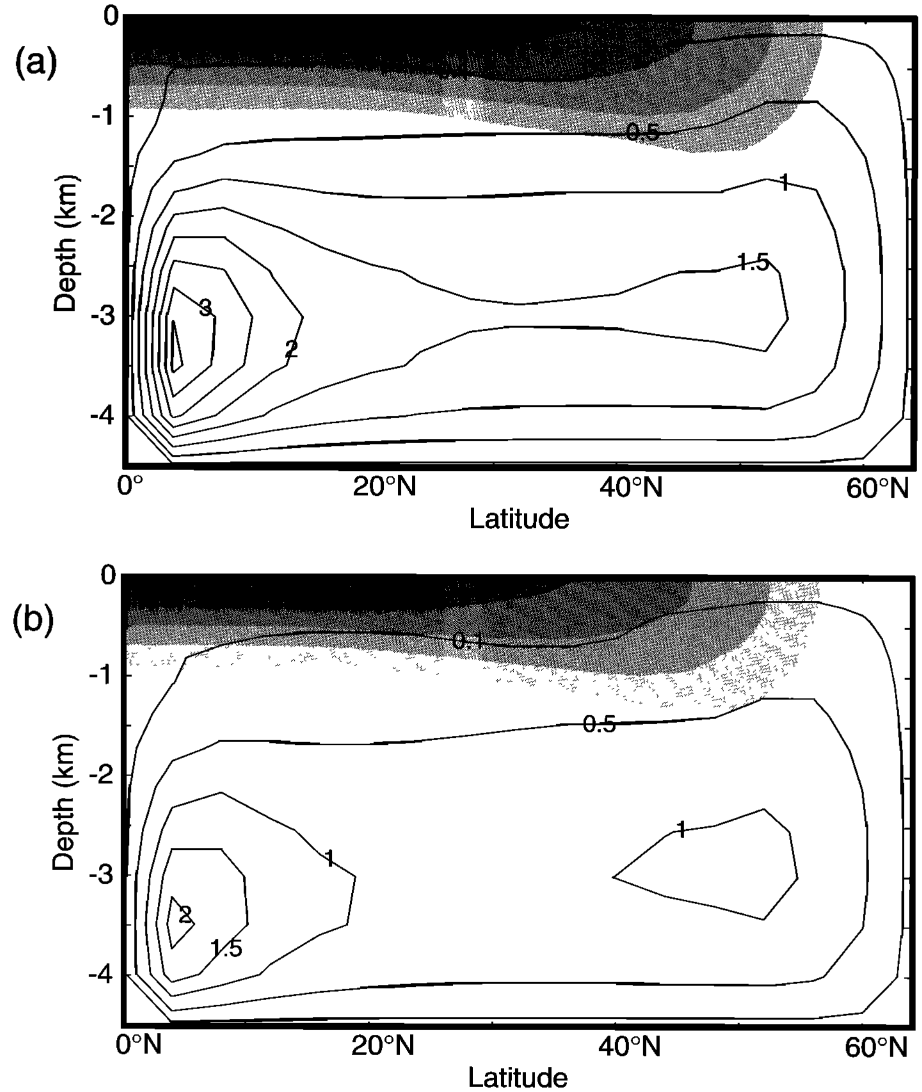

Figure 6. Anomalous meridional overturning circulation and zonally averaged temperature (shading) given (a) an increase in geothermal heating along a 12° meridional strip through the middle of the basin, maintaining heating elsewhere, and (b) a more significant increase in heating along this mid-ocean ridge with zero geothermal heat input elsewhere, so as to maintain the total heat input of the uniform geothermal heating run. Contours are labeled in sverdrups, with solid contours indicating clockwise flow; isotherms are 1.35°, 2.7°, 5.4°, 10.8°, and 21.6°C.

small as permitted by our crude resolution. Away from the equator, weak anomalous downwelling occurs in the ocean interior, as observed in meridional cross sections (not shown). The vorticity equation $\beta v = f\partial_z w$ implies meridional flow if vertical convergence is nonzero; in the southward branch of the anomalous cell, $\partial_z w < 0$, whereas $\partial_z w > 0$ in the northward branch. The vorticity equation could seemingly also permit a reversed anomalous circulation, i.e., with narrow sinking at the equator and weak upwelling in the interior. However, this circulation would flow counter to the background overturning and therefore would be ineffective at removing the surplus heating. We ran an additional experiment (not shown) with an inverted equator-pole temperature gradient, which produced a background state with deep sinking at the equator. Given this surface forcing, we obtained an anomalous geothermal response of narrow sinking at the equator and broad upwelling in the ocean interior, consistent with this argument.

In contrast with the interior flow, anomalous upwelling occurs along the tropical eastern and western boundaries (as shown for the eastern boundary in Figure 4b). At the eastern boundary, upwelling occurs when anomalous eastward zonal flow reaches the boundary. At the western boundary it is not immediately clear why the anomalous upwelling occurs. As discussed by *Spall* [2000], viscous terms are important in the potential vorticity budget at the boundaries. Anomalous eastern and western boundary upwelling occurs regardless of whether diapycnal mixing is concentrated at the boundaries, as there was no change using our model run with uniform diffusivity.

Interestingly, the perturbation in the meridional overturning circulation is similar as forced by either geothermal heating or enhanced deep diapycnal mixing (see SM's Figure 12b for the overturning stream function with enhanced deep mixing, using an otherwise similarly configured model). Not coincidentally, both geothermal heating and increased deep diffusion lead to greater heat fluxes into the model's bottom layer. There is a fundamental difference in the dynamics of the response, however. With enhanced deep mixing the rising motion of the deep cell occurs

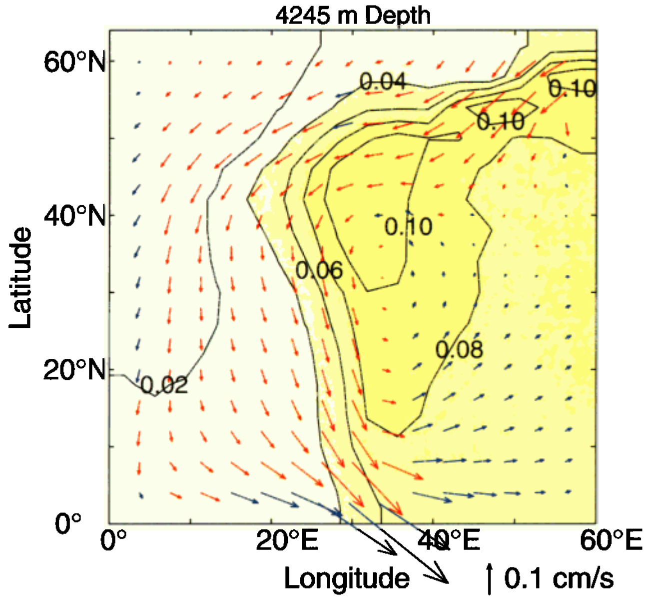

Plate 4. Plan view of the anomalous circulation and anomalous temperature of the bottom model layer in the ridge-only geothermal heating run. Blue vectors indicate anomalous upwelling, and red arrows indicate anomalous downwelling.

where forced by mixing as required by advective-diffusive balance (in SM's experiments this resulted in more "boundary" upwelling, including along the equator). In contrast, the geothermal heating response can be interpreted as a free response constrained by rotational dynamics.

To test whether the anomalous overturning maximum in our solution is dependent upon the equatorial grid spacing, given the localization of anomalous upwelling in these grid cells, we reran the geothermal heating simulation with resolution decreased by 50%. We found only a slight decrease in the magnitude of the anomalous overturning cell (not shown).

4. Alternate Forcing Scenarios

4.1. Alternate Background States

Our analyses in section 3 suggest that the magnitude of the geothermal response is a function of the background state flow and density structure. To test this hypothesis, we reran our geothermal simulation with both stronger and weaker background meridional overturning circulations.

A stronger background overturning can be achieved through increased diapycnal mixing [*Bryan*, 1987; *Marotzke*, 1997]. Here we increased the boundary mixing strength fourfold, which produced a maximum overturning of 31 Sv at steady state (as compared to 12.7 Sv for the standard run). The anomalous geothermal response in overturning is shown in Figure 5a. There are several reasons why this response is considerably weaker than our standard geothermal response. First, increased mixing thickens the thermocline, thus decreasing the depth of the abyss while increasing its stratification. Consequently, the buildup of meridional shear resulting from east-west boundary temperature differences occurs across a reduced depth, producing a weaker anomalous overturning cell. Second, the background flow's transit time through the abyss (primarily a function of overturning strength) is decreased, resulting in weaker anomalous zonal temperature gradients, again reducing the buildup of meridional shear. Finally, the increased effectiveness of diffusive processes implies that a weakening in stratification can lead to a more significant reduction in deep diffusive heat fluxes. In other words, the system is better able to offset the geothermal heating through a diffusive response, making the advective response less critical as a means to reestablish a steady state heat balance.

The anomalous geothermal response with boundary diapycnal mixing strength reduced by 80% is nearly double that of the standard experiment (Figure 5b). Here the background state overturning maximum is 4.6 Sv, the same order of magnitude as the geothermal response, indicating that nonlinear interactions are likely to be significant. The opposite of the arguments employed in the stronger mixing case apply; namely, the abyss is deeper and less stratified, the transit time through the abyss is longer, and the diffusive fluxes are extremely weak (although we suggest the latter is of little consequence given that these fluxes are already quite small in the standard run). Note that the increased response occurs primarily at low latitudes, as compared to Figure 2b. The anomalous circulation through high latitudes, whereby heat is effectively communicated to the atmosphere through advection into deep mixed layers, is little changed as the applied geothermal heating is unchanged from our standard run. Given the sluggish deep circulation, here convection along the bottom boundary is maintained even as the system has reached a steady state response to geothermal heating. Accordingly, the warmest waters in the bottom layer are now found in the tropical eastern boundary rather than in the northeastern corner (not shown). Anomalous cooling in the thermocline peaks at 0.3°C, double that of the control experiment.

4.2. Variations in Distribution and Magnitude of Geothermal Heating

As discussed in section 1, the flux of geothermal heat is enhanced in the vicinity of the mid-ocean ridges. To examine how localization affects the anomalous response, we ran two variations of our geothermal heating experiment. In the first we specified stairstep fluxes of 100–200–100 mW m^2 across a 12° longitude span through the center of the ocean basin, maintaining the background 50 mW m^2 at other locales. This new forcing represented a 30% increase in total geothermal heating. Note that this is only a partial representation of the possible effects from ridge heating as we maintain a flat bottom in all runs. The low-latitude response (Figure 6a) shows only a modest increase in maximum overturning as compared to the uniform heating run at this resolution. More significant is the degree to which the anomalous overturning extends to higher latitudes. In the second variation we located all anomalous heating along a 200–400–200 stairstep, which maintained the total heat flux of the uniform forcing run. In this run the maximum overturning at low latitudes is reduced considerably (Figure 6b), but the anomalous downwelling through high latitudes is similar to the uniform heating run.

These two additional runs, taken together, uphold the arguments presented heretofore. When the total geothermal heating is increased, the anomalous high-latitude circulation must increase as this component represents the only means by which excess heat is transported out of the abyss. The strength of the anomalous low-latitude circulation is largely a function of the east-west density difference. Increased heating along the "ridge" produces a local maximum in temperature (as shown in Plate 4), which leads to anomalous cyclonic circulation; as such, the ridge heating is only partially communicated to the eastern boundary. This behavior is analogous to that of localized interior mixing, which is also less effective at driving a meridional overturning circulation (see SM for a discussion). On the western side of the ridge there is a fairly sharp anomalous temperature gradient. As required by geostrophic balance, a significant portion of the anomalous circulation flows along this gradient, with less flow through the deep western boundary current as observed in the uniform heating run.

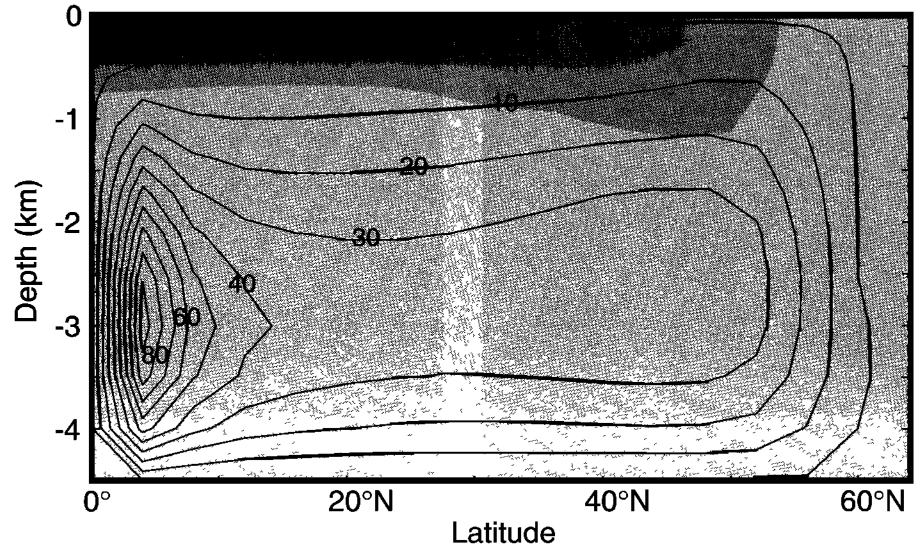

Figure 7. Anomalous meridional overturning circulation and zonally averaged temperature (shading) produced by a uniform geothermal heating of 5 W m^{-2}. Contours are labeled in sverdrups, with solid contours indicating clockwise flow; isotherms are 2.7°, 5.4°, 10.8°, and 21.6°C.

To test the sensitivity of the response to the magnitude of geothermal heating, in as much as the heating rate has not been constant through geologic time (see *Kadko et al.* [1995] for a review) we ran an experiment with an arbitrarily chosen uniform 5 W m^{-2} geothermal heating, a 2 order increase from our control run. The choice was not meant to be realistic but simply to represent an extreme case whereby geothermal fluxes are closer in magnitude to surface heat fluxes. As shown in Figure 7, the resulting anomalous circulation is quite remarkable. The response is essentially an exaggerated version of our results given a weaker background circulation. Here the anomalous response constitutes the dominant flow except in upper thermocline waters. The bottom portion of the thermocline has been eroded in the tropics, and the anomalous overturning maximum is 94 Sv. There is also a secondary maximum in the anomalous stream function at high latitudes (as before, about one third of the overall maximum). Even in this extreme, with the predominant ocean circulation now primarily driven by geothermal heating rather than by diapycnal mixing, the resulting maximum meridional heat transport is only 0.1 PW greater than the nongeothermal run. This result suggests that any realistic changes in geothermal heating rates are not likely to affect climate directly through the oceans' heat transport.

5. Geothermal Heating in a Double Hemisphere Model

We wish to test whether the equatorial upwelling observed in the single-hemisphere runs is a robust feature or due to the limited geographical extent of the model. In addition, introducing a second hemisphere adds two notable complexities to the system: a second competing source for deep water and the possibility of cross-hemisphere flow. These experiments also serve as a conceptual link between the geothermal heating results in a realistic global configuration, as described by *Adcroft et al.* [2001], and our single hemisphere results. In this section we extend our results by adding

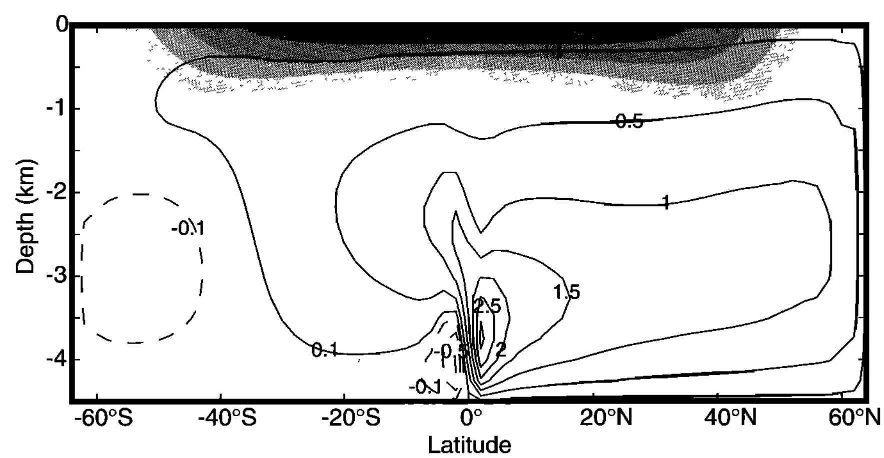

Figure 8. Anomalous meridional overturning circulation and zonally averaged temperature (shading) in a double-hemisphere basin given uniform geothermal heating of 50 mW m^{-2}. Contours are labeled in sverdrups, with solid contours indicating clockwise flow; isotherms are 1.35°, 2.7°, 5.4°, 10.8°, and 21.6°C.

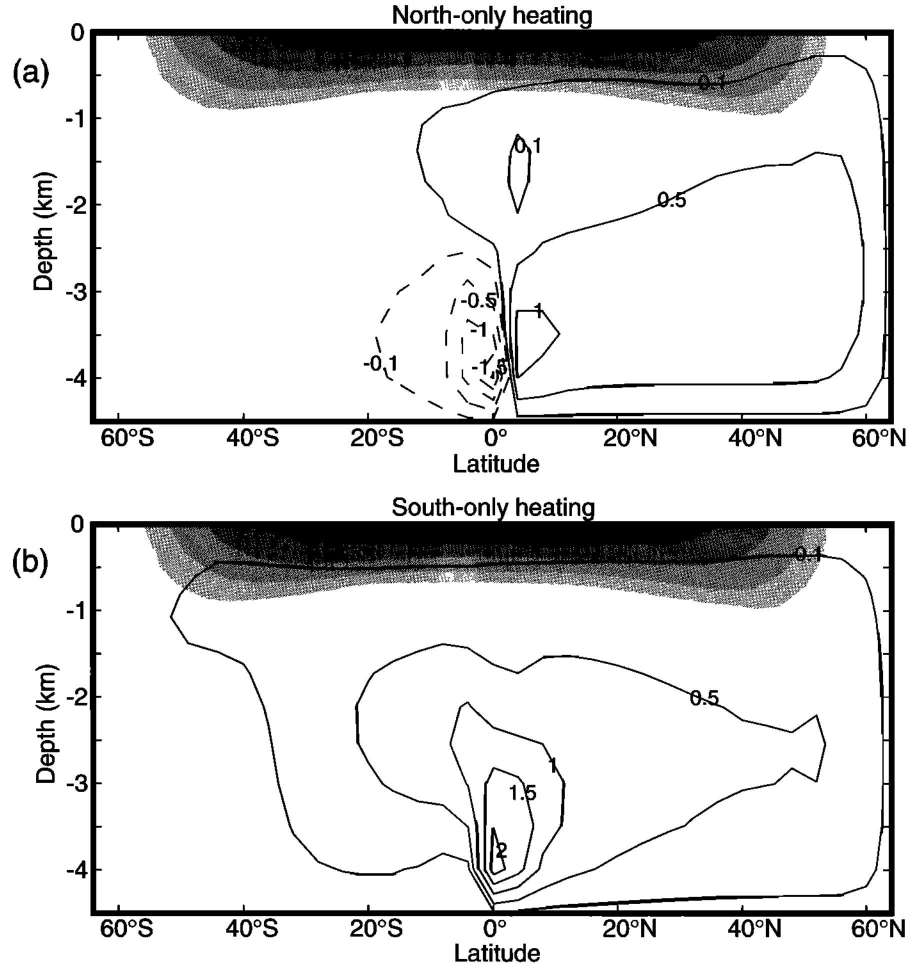

Figure 9. Anomalous meridional overturning circulation and zonally averaged temperature (shading) in a double-hemisphere basin given (a) geothermal heating of 50 mW m^{-2} in the Northern Hemisphere, with no geothermal heat input in the Southern Hemisphere, and (b) geothermal heating of 50 mW m^{-2} in the Southern Hemisphere, with no geothermal heat input in the Northern Hemisphere. Contours are labeled in sverdrups, with solid contours indicating clockwise flow; isotherms are 1.35°, 2.7°, 5.4°, 10.8°, and 21.6°C.

a second hemisphere to our ocean basin, setting up an asymmetric background flow pattern through a 1°C pole-to-pole difference in surface-restoring temperature profiles. The dynamics of the two-hemisphere flow is discussed in detail by *Marotzke and Klinger* [2000], but a salient feature is that northern originated water fills much of the abyss, with downwelling water in the Southern Hemisphere only able to penetrate to middepth.

Figure 8 shows the anomalous meridional overturning in the double-hemisphere system. Anomalous upwelling occurs all along the equator, although there is no longer any boundary mixing here, as was parameterized in the single-hemisphere model. Total equatorial upwelling is about the same as in the single-hemisphere run, although some of the tropical downwelling occurs in the Southern Hemisphere, as evidenced by the counterclockwise cell abutting the equator to the south. As before, the geothermal heat flux is lost to the atmosphere locally in the northeast corner (not shown). Since the area-integrated geothermal heat flux is now twice as large, the anomalous surface warming is greater in magnitude and covers a larger area than in the single-hemisphere run. The anomalous overturning through the Northern Hemisphere high latitudes is only slightly enhanced from the single-hemisphere results (namely, Figure 2b versus Figure 8). As we have argued that this component of the circulation is critical in communicating the geothermal heat fluxes to the surface, the lack of a bigger response is surprising and suggests that the anomalous circulation is a more efficient heat conveyor in the double-hemisphere system. In contrast with the single-hemisphere results, the enhanced downwelling in the northeast is to large extent supported by anomalous upwelling in the Southern Hemisphere (as implied by Figure 8). As such, the cold tongue that results from anomalous upwelling into the upper layers (as observed in Plate 1b) is now present in the Southern Hemisphere. The Northern Hemisphere thermocline is less anomalously cold in the double-hemisphere run, so advection into the Northern Hemisphere's deep mixed boundary layers conveys a greater heat content. Thus, by diverting anomalous flow through the Southern Hemisphere the double-hemisphere system is able to transport effectively geothermal heat fluxes to the surface through a weaker advective response. This "global" anomalous meridional overturning cell is similar to what is observed in *Adcroft et al.*'s [2001] simulation, where realistic topography presumably limits the magnitude of the anomalous response at the equator.

As compared with the single-hemisphere system, there is little change in the flow pattern through the Northern Hemisphere. In the Southern Hemisphere the anomalous warming is greater but is nearly zonally uniform (not shown). The lack of zonal asymmetry means that a strong anomalous meridional overturning circulation cannot be supported by thermal wind balance south of the tropics. There is anomalous flow in the Southern Hemisphere that circulates zonally, however, rising in the eastern boundary and sinking in the western boundary. Along both boundaries, anomalous flow augments the background flow (sinking along the western boundary is necessary for thermal wind balance of the cross-hemisphere meridional overturning circulation, as noted by *Marotzke and Klinger* [2000]).

We ran two additional double-hemisphere geothermal heating runs: in the first we applied the geothermal heating only in the Northern Hemisphere, and in the second we applied the flux only in the Southern Hemisphere. To lowest order the response to heating across the full basin can be viewed as the superposition of these separate runs. In the northern-only run (Figure 9a), the counterclockwise cell in the Southern Hemisphere is more pronounced; here the deep heating in the Northern Hemisphere results in a weak anomalous warming in the Southern Hemisphere abyss, albeit with more zonal structure than when heated directly. Also, note that the extension of the anomalous overturning into the Southern Hemisphere high latitudes is less pronounced than in Figure 8. Therefore it is not surprising that stronger anomalous flow into northern high latitudes is observed as compared to the southern-only run (Figure 9b), which exhibits stronger intrahemispheric flow (the anomalous circulation through the northern high latitudes is weaker than the full-basin heating run in both cases as the area-integrated heat flux is decreased by half). The southern-only run is also notable in that all anomalous downwelling occurs in the Northern Hemisphere, producing a single overturning cell.

6. Summary and Discussion

Our experiments illustrate the effect of geothermal heating on the large-scale circulation in an idealized ocean basin. Prior work has focused on the dynamical response to high-temperature venting, which can penetrate to greater heights than either low-temperature venting or warming through a weak conductive flux. Hence only the high-temperature forcing has previously been thought to be dynamically significant, even though the integrated heat flux through the low-temperature sources is much greater than that of local hot spots [*Stein et al.*, 1995]. Both effects have previously been ignored in general circulation models, and only the high-temperature forcing has been studied in regional models. Our results suggest that the weaker conductive heat flux can have a significant impact in the general circulation: we find that the integrated effect of uniform weak heating through the ocean bottom produces a significant change in the large-scale overturning circulation, on the order of several sverdrups. Although the meridional heat transport of the anomalous circulation is insignificant as compared to that of the background surface-forced circulation, the deep circulation may have an important indirect effect on climate through the transport of carbon and other nutrients, which have their concentration maxima at depth.

We find that the large-scale response to geothermal heating is largely advective, as suggested by the change in the meridional overturning circulation. A simple diffusive response is not realized, which would require a significant perturbation in deep-water properties. In the presence of a background flow field, heating of the abyss does not occur uniformly; in a rotating system these anomalous temperature gradients are then balanced geostrophically through anomalous flow. The change in deep circulation is governed by the requirement that the flow communicate the geothermal heating to the surface in order that steady state be achieved. Anomalous advection into convecting boundary layers can effectively pipe anomalous heat to the surface, precluding the need for strong anomalous flow at the surface. To lowest order the geothermal response serves to augment the preexisting background circulation. A more stratified abyss and/or a stronger background overturning reduce the magnitude of the geothermal response. When geothermal heating is applied to a double-hemisphere ocean model, our findings are largely consistent with those in a single-hemisphere model.

In low latitudes a maximum in the anomalous overturning circulation results from a zonal temperature gradient that is established as deep eastward flow is warmed. All anomalous upwelling (in the zonal mean) occurs along the equator, although some of this anomalous flow simply recirculates in the tropics. The thermal wind relation can be used to estimate the magnitude of this local maximum, presuming that the background flow and stratification are known. As might be expected in a rotating system, the thermal wind relation also can be used to relate the anomalous overturning strength to zonal temperature gradients at middle and high latitudes.

From a thermodynamic standpoint, weak geothermal heating at depth is able to drive a substantial anomalous overturning circulation because the heating occurs at low geopotential, thus increasing potential energy. Although geothermal fluxes are much smaller in magnitude than surface heat fluxes, the latter occur at high geopotential and thus require diapycnal mixing in order to drive a meridional overturning circulation.

Without the context of a background circulation, however, uniform geothermal heating is unable to generate a coherent large-scale response. Consider the model's response to geothermal heating in a temperature-only "box" ocean governed by uniform surface forcing. Without an imposed meridional temperature gradient or applied wind stress the resulting ocean is isothermal, and no background circulation is established. If geothermal heating were to be applied uniformly along the ocean floor, it would generate unstable columns of water everywhere. Convective adjustment redistributes this heat throughout the full depth of the water columns so that at the surface the additional heat is released to the atmosphere. With steady heating from below and restoring at the surface (cooling), every water column is in a constant state of convective adjustment. This configuration closely resembles the classical Rayleigh-Bénard convection problem with rotation [see *Emanuel*, 1994, chap. 3]; presumably, with a sufficiently high-resolution model the steady state solution would consist of an array of Rayleigh-Bénard convective overturning cells (the scale of these cells would likely be on the order of the ocean depth, although the realized form of Rayleigh-Bénard convection in our configuration is not entirely clear; *Krishnamurti* [1980] found that in certain laboratory configurations with very high Rayleigh number, domain-filling circulations were possible because of up-gradient momentum transport by convection). Given our limited resolution, however, the convective adjustment parameterization is the only active process in the model, conveying the deep heating directly to the surface. At no time would any large-scale motion occur. Even though geothermal heating embodies the necessary characteristic of heating at low geopotential, in the absence of a forced circulation or maintained stratification it is insufficient to produce a large-scale circulation.

The connection between diffusive heating and geothermal heating as "driving" mechanisms can be extended further. Both produce heating at depth, which can be conveyed back to the surface through convective mixing. Thus convection acts as a means by which the ocean can remove heating necessary to sustain motion, whether the heating is diffusive or geothermal.

Our model is highly idealized and is therefore at best only a crude representation of the real ocean, but it has proven helpful in understanding the geothermal response in a more realistic global configuration [*Adcroft et al.*, 2001]; it is tempting to speculate further that geothermal heating might similarly affect our oceans' circulation. Although we have not made a direct comparison between our single-hemisphere model and the

North Atlantic, it is implicit that our model represents a simplified depiction of North Atlantic Deep Water circulation. However, we would be remiss in not pointing out several limitations that obscure how these results might apply to the real ocean. First, our model has no topography; as such, all geothermal heating occurs at a single depth, in contrast with the real ocean where the heating is effectively spread across a wide range of depths. Second, our model has a single source of deep water (even in our double-hemisphere configuration), whereas the real ocean is characterized by deep-water formation in both the Northern Hemisphere and the Southern Hemisphere. Given that much of the deep ocean floor is in contact with the denser Antarctic Bottom Water rather than the North Atlantic Deep Water, we suggest that our model is actually a better analog of how geothermal heating affects the circulation in the South Pacific, where Antarctic Bottom Water is formed. Accordingly, the global model results presented by *Adcroft et al.* [2001] show a stronger, more coherent response in the Pacific Ocean.

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