Role of the overturning circulation in determining the potential vorticity over the abyssal ocean

Richard G. Williams, Kate Day, and Vassil Roussenov
Oceanography Laboratories, Department of Earth Sciences, University of Liverpool, Liverpool, UK

Richard Wood
Hadley Centre for Climate Prediction and Research, Bracknell, UK

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1. Introduction

[1] The evolution of the potential vorticity (PV) in the deep waters is diagnosed from output of the Hadley Centre Climate Model over a 1000 year integration. Extensive regions of low PV are formed whenever there is an inflow of dense fluid across the equator into a relatively isolated basin. Initially, the model reveals low contrasts in PV in the deep waters over the North Pacific and higher contrasts over the Southern Hemisphere and North Atlantic, which are consistent with climatological diagnostics. However, a drift in the climate model leads to the northward influx of bottom water weakening in the Pacific but strengthening in the Atlantic. This change in overturning circulation leads to the PV along deep neutral surfaces acquiring a weaker contrast over the North Atlantic compared with the Southern Hemisphere and the North Pacific. Transient tracer experiments suggest that these regions of low PV are useful in identifying abyssal layers that are strongly ventilated from across the equator.


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2. Potential vorticity (PV) has been used as a dynamical tracer to understand the circulation of the atmosphere [e.g., Hoskins et al., 1985] and ocean [e.g., McDowell et al., 1982]. Over the upper ocean, open contours of PV intersecting the mixed layer have been associated with gyre-scale ventilation [Layton et al., 1983], while nearly uniform PV has been associated with eddy homogenisation [Rhines and Young, 1982; Lozier, 1997] (Figure 1a). In the deep ocean, there is also a rich variety in the PV structure with contours inclined to latitude circles and even regions of nearly uniform PV [O'Dwyer and Williams, 1997] (herein-after referred to as OW) (Figure 1b). Interpreting the dynamical control of the PV distribution is difficult though, as the PV distribution does not uniquely identify the controlling process. For example, the nearly uniform PV in the upper thermocline of the North Atlantic might be formed through ventilation [Williams, 1991], rather than eddy stirring [Rhines and Young, 1982].

[3] In this study, we wish to investigate how the PV distribution for deep and bottom waters is controlled. On the basin scale, PV is effectively given by $f/h$ where $f$ is the planetary vorticity and $h$ is the layer thickness. As dense water masses spread from their source regions, mixing generally acts to decrease their stratification and increase their layer thickness. When the water masses spread poleward, the increase in the magnitude of $f$ partly compensates that of $h$ leading to weak meridional gradients in PV. Conversely, when the water mass moves equatorward, the changes in $f$ and $h$ reinforce each other leading to larger meridional gradients in PV. Extensive regions of low PV are expected to be formed when a single water mass enters a relatively isolated basin from across the equator (Figure 1b), since $f$ is zero there (OW). In practice, this process will only form extensive regions of low PV over the Northern Hemisphere, since all dense water masses are either formed or significantly modified at high latitudes in the Southern Ocean.

[4] Here, we investigate how these mechanisms control the distributions of abyssal PV using diagnostics from a global, coupled circulation model from the Hadley Centre Coupled Model (HadCM3) [Gordon et al., 2000]. The model has been integrated for a 1000 years at a horizontal resolution of 1.25 degrees, which is sufficiently long for the diagnostics to lose much of their memory of the climatological initial state. While the realism of the final state is debatable, we examine how the overturning circulation and PV evolve together over the Pacific and Atlantic basins. The climate model is too coarse though to assess any effects of mesoscale eddies. This study compliments a more idealised
2. Hadley Centre Model

2.1. Model Configuration

[6] The HadCM3 coupled climate model is described in detail by Gordon et al. [2000]. Here we repeat some of the features of the ocean component of the model that are most salient to this study. The ocean component follows the Cartesian formulation of Cox [1984] with a horizontal resolution of 1.25° and 20 vertical levels. A mixed layer model is included for tracers [Kraus and Turner, 1967; Gordon and Bottomley, 1985] and a K theory scheme for momentum [Pacanowski and Philander, 1981].

[7] Diapycnic mixing of tracers below the surface mixed layer is achieved through the Pacanowski and Philander [1981] scheme, where the diapycnic diffusivity, \( \kappa_d \), is given by

\[
\kappa_d(z) = \kappa_b(z) + \left( v_b + v_\alpha (1 + 5Ri)^{-2} \right) (1 + 5Ri)^{-1}.
\]

Here \( Ri \) is a local Richardson number, \( v_b = 1.0 \times 10^{-5} \) m²s⁻¹, \( v_\alpha = 5.5 \times 10^{-3} \) m²s⁻¹ and a linear profile is used for \( \kappa_d(z) \), increasing from \( 1.0 \times 10^{-5} \) m²s⁻¹ at the surface to \( 1.5 \times 10^{-4} \) m²s⁻¹ at 5000 m. In practice, \( Ri \) is large enough that \( \kappa_d \) is approximately equal to the background value \( \kappa_b + v_\alpha \) over most of the ocean, the exceptions being within and immediately below the surface mixed layer, and in the South Atlantic at the interface between the North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW). The Richardson number is calculated using potential density relative to the surface, and as a result the model mixing in the latter region is over excessive.

[8] Eddy transport and diffusion of tracers is represented by the Gent and McWilliams [1990] scheme with spatially varying transfer coefficients [Visbeck et al., 1997] in the range of 350 to 2000 m²s⁻¹, and by isopycnal diffusion of tracers with a constant diffusivity of 1000 m²s⁻¹. The ocean component is initialised from climatology [Levitus et al., 1994; Levitus et al., 1995] and the coupled global model is integrated for 1000 years without any artificial flux adjustment.

2.2. Model Drift of the Circulation

[9] The overturning circulations evolve in a contrasting manner over the Indian-Pacific and Atlantic basins during the 1000 years integration of HadCM3. Over the Indian-Pacific, there is a single overturning cell, initially reaching a maximum strength of 20 Sv at 25°S with 10 Sv crossing the equator (Figure 2a). This influx of dense fluid is returned as lighter intermediate fluid, rather than surface fluid; in accord with model studies based on silica distributions, such as Gnanadesikan [1998]. There is a striking collapse of this overturning cell in the North Pacific, which has less than 5 Sv of dense fluid fluxed northward to the equator by year 1000. This decrease in overturning circulation is reflected in a weakening of the meridional velocity in the western boundary at 4 km (Figure 2b).

[10] Over the Atlantic, there is the expected overturning cell importing light water and exporting NADW (Figure 3a). The cell extends from the surface to 3 km and reaches a maximum strength of 20 Sv at 40°N. There is 15 Sv of NADW exported across the equator at year 30, but this flux weakens to 10 Sv by year 1000. The northward flux of AABW is initially close to zero across the equator, but this flux strengthens to 5 Sv by year 1000. This change in the overturning circulation is reflected in the deep velocities changing from generally southwards in year 30 to northward in year 1000 along topographic boundaries; see velocity arrows in Figure 3b plotted along a neutral surface at a depth of typically 4 km.
In the coupled model integration, local errors in the region of AABW formation initially result in the AABW being formed too cold and salty. In the Indian-Pacific, this results in a strong northward flow of dense bottom water, as seen in Figure 2a at year 30. However, over the first few centuries of the run, an overstrong model freshwater cycle also results in NADW which is too warm and salty (A. K. Pardaens et al., Comparing HadCM3 against a range of datasets to evaluate the freshwater cycle, submitted to Climate Dynamics, 2001). From about year 100 onwards, this warm, salty bias contaminates the Circumpolar Deep Water, and the anomalous warming eventually reduces the rate of density increase in the deep Southern Ocean. This warming weakens the meridional density gradient driving the bottom water overturning cell in the Indian-Pacific (Figure 2b, 1000 years). In the Atlantic, the situation is more complicated because of the ‘competition’ between NADW and AABW. The intensification of the bottom cell there (Figure 3) is partly due to both the initial state underestimated the spreading of AABW and excessive mixing occurring between the bottom water and overlying deep water in the South Atlantic.

Over the whole domain, the sum of the overturning cells over the Pacific and Atlantic weakens from 39 Sv in year 30 to 29 Sv in year 1000. This reduction in overturning is due a decease in the northward flux of AABW over the Indian-Pacific. In the Atlantic, there appears though to be a compensating response with a strengthening in the influx of AABW (increasing by 5 Sv) and a similar weakening in the export of NADW. In comparison, Wang et al. [1999] using an idealised coupled climate model find that a change in freshwater forcing alters the relative formation rates of NADW and AABW, rather than their sum. However, this similar response to the HadCM3 evolution over the Atlantic may be fortuitous, since Wang et al. [1999] consider equilibrium solutions obtained after several millennia of integration, whereas the GCM run is still showing transient adjustment after 1000 years.

3. Evolution of Potential Vorticity Over the Deep Waters

3.1. Potential Vorticity Definition

The Ertel Potential vorticity is defined by

\[ PV = \frac{1}{\rho} (2\Omega + \nabla \times \mathbf{u}) \cdot \nabla \lambda, \]

where \( \Omega \) is the angular velocity vector, \( \mathbf{u} \) is the three-dimensional velocity vector, \( \lambda \) is a conserved tracer and \( \rho \) is
a reference density. In this study, we choose \( \lambda \) to be represented by the neutral density, \( \gamma \) [McDougall, 1988; Jackett and McDougall, 1997], which water parcels are believed to follow more closely, than potential density, \( \sigma \). The only exception is when the model forms artificially saline, deep waters with the definition of \( \gamma \) becoming unclear and then \( \lambda \) is represented by \( \sigma \) referenced to a depth of 4 km.

For completeness, the PV is evaluated with all terms retained in (2). However, due to the resolution of the model, the PV is nearly always dominated by contributions from the vertical component of planetary vorticity, \(-f \partial \gamma / \partial z\), and near the equator from the horizontal component of planetary vorticity, \(-2\Omega \cos \theta \partial \gamma / \partial y\); where \( \theta \) is the latitude and \( f = 2\Omega \sin \theta \). Hence, initially, the HadCM3 diagnostics are consistent with large-scale diagnostics based on climatology for the upper ocean [Keffer, 1985; You and McDougall, 1990] or deep ocean (OW).

### 3.2. Background State for Potential Vorticity

The undulations of the main thermocline control the meridional variation in neutral surfaces and PV in the upper waters as shown in Figure 4 (from model diagnostics for year 30 along 25\(^\circ\)W in the Atlantic and 160\(^\circ\)W in the Pacific). The thermocline is thicker over the subtropical gyres, centred at 30\(^\circ\)N and 30\(^\circ\)S. The neutral surfaces are pushed downward over the subtropical gyre and contain a thick layer of high PV. In contrast, the thermocline is thinner and shallower over the subpolar gyres, centred at typically 50\(^\circ\)N and 50\(^\circ\)–60\(^\circ\)S, where neutral surfaces are upwelled and only contain a thin layer of high PV.

Over intermediate depths, typically from 2–3 km, neutral surfaces are generally flat apart from south of 40\(^\circ\)S in the Atlantic and 50\(^\circ\)S in the Pacific where they shoal southwards. In the deep water below 3 km, neutral surfaces deepen northward from the Southern Ocean and eventually ground on the seafloor (Figure 4a). This structure leads to contours of low PV fanning out from the equator into the deep waters of both basins (Figure 4b). In the deep water, PV is generally higher in magnitude in the Southern Hemisphere, than in the Northern Hemisphere for a comparable latitude.

Over the 1000 year integration of HadCM3, the PV structure evolves, particularly for deep and bottom waters, and drifts away from the initial climatological state. Consequently, we examine the relationship between the overturning and changes in the deep PV structure.

### 3.3. PV Evolution Over the Deep Pacific

In the Pacific, the bottom waters become denser with neutral density increasing by 0.3 over 1000 years, while the surface waters become lighter by 0.1. Over the upper and
intermediate waters, there is a general increase in PV at a particular depth through the increased stratification over the model integration. Over the deep waters, the neutral surfaces gradually become flatter, while surfaces deepening northward become increasingly confined to a thinner, bottom layer (Figure 5a). This general flattening of the neutral surfaces suggests that the modelled mixing is insufficiently strong to convert all of the excessively dense, bottom water into deep or intermediate water in the Pacific. Consequently, the basin gradually fills up with denser bottom water, which in turn is associated with a weakening of the overturning circulation (Figure 2).

Initially, there are much stronger PV contrasts over the South Pacific, compared with the North Pacific, as diagnosed along a dense neutral surface and plotted in the map of Figure 5b. However, during the model integration, the gradual flattening of the neutral surfaces (around a depth of 4 km) leads to the PV contrast increasing, particularly over the northern basin. For example, at years 750 and 1000 for the North Pacific, the corresponding PV along a neutral or potential density surface reaches a magnitude of $6 \times 10^{-12} \text{m}^{-1} \text{s}^{-1}$ relative to an initial value of typically $3 \times 10^{-12} \text{m}^{-1} \text{s}^{-1}$. Note that at year 1000, PV is evaluated in terms of vertical spacing of sigma surfaces due to the drift in deep waters properties preventing neutral surfaces being evaluated.

Hence, the filling up of the Pacific basin with dense fluid and weakening in overturning is associated with an increase in the PV contrast within the northern basin. There are still some weak contrasts in PV along denser neutral surfaces, but as the integration progresses these low values in PV become increasingly confined to a thin layer above the seafloor.

3.4. PV Evolution Over the Deep Atlantic

In the Atlantic, there is a similar increase in stratification with the bottom waters also increasing by 0.3 in neutral density over 1000 years, and the surface waters decreasing by 0.1. However, unlike in the Pacific, the neutral surfaces in the bottom waters change during the integration from being relatively flat to deepening northward and eventually grounding (Figure 6a). This grounding of neutral surfaces, together with the overturning circulation in Figure 3, implies that the mixing is suffi-
Initially, over the Atlantic, there are relatively strong meridional contrasts in PV in both hemispheres (Figure 6b). As the inflow of bottom water increases and the neutral surfaces deepen northward, then the PV contrasts weaken over the Northern Hemisphere and strengthen over the Southern Hemisphere. Eventually, the bottom waters in

**Figure 5.** (a) Meridional section for neutral surfaces, $\gamma$, below 2 km along 170°W in the Pacific and (b) accompanying map of PV ($10^{-12}$m$^2$s$^{-1}$) along deep $\gamma$ surfaces for years 30, 100, 750, and 1000. The shading represents where the $\gamma$ surface has ground. The selected $\gamma$ surfaces for the PV maps are shown by the dashed line in Figure 5a. Note that for the deep Pacific in year 1000, $\sigma_4$ is plotted rather than $\gamma$. 

ciently strong here to convert the bottom waters to deep waters.
Figure 6. (a) Meridional section for γ surfaces below 2 km in the Atlantic and (b) accompanying map of PV (10^{-12} m^-1 s^-1) along deep γ surfaces for years 30, 100, 750, and 1000. The shading represents where the γ surface has ground. The section is along a variable track between 40° and 60°W, as shown by the dashed line in Figure 6b. The selected γ surfaces for the PV maps are shown by the dashed line in Figure 6a.
the North Atlantic acquire a relatively low PV of $2 \times 10^{-12} \text{m}^{-1}\text{s}^{-1}$, compared with a typical initial value of $4 \times 10^{-12} \text{m}^{-1}\text{s}^{-1}$. This evolution occurs rapidly over the first 100 years, perhaps linked to a transient influx or mixing event, then a relatively stable state is reached with extensive regions of low PV being maintained at years 750 and 1000.

3.5. Evolution of Meridional Gradients in PV

[23] The meridional gradient in PV is evaluated here in terms of the PV change relative to the planetary vorticity, $\partial PV/\partial f$, along a neutral surface for the Pacific and Atlantic for years 30 and 1000 (Figure 7). In each case, the reference PV gradient due to the change in $f$ is included, as marked by a horizontal line, assuming the mean stratification, $\partial \gamma/\partial z$, along the section.

[24] Over the Pacific, the meridional gradient in PV increases as the overturning collapses between years 30 and 1000 (Figure 7a). This increase in both hemispheres is due to the overall increase in stratification. However, over the Atlantic, the meridional gradient in PV generally decreases in the Northern Hemisphere and increases in the Southern Hemisphere between years 30 and 1000 (Figure 7b). This response is due to the strengthening in the northward flux of AABW over the Atlantic, which increases the input of more stratified, higher PV waters into the South Atlantic and increases the input of low PV waters into the North Atlantic.

[25] Only over the Northern Hemisphere does the meridional gradient in PV become much smaller than the reference PV gradient due to the change in planetary vorticity with constant stratification (compare with the horizontal lines in Figure 5). In particular, the PV only approaches a nearly uniform value for the North Pacific at year 30 (from 30$^\circ$S to 40$^\circ$N) and the North Atlantic at year 1000 (from 15$^\circ$S to 30$^\circ$N).

[26] In summary, there is a consistent response in how the PV is controlled by the overturning over the Northern Hemisphere: weak contrasts in PV are formed when the overturning circulation drives an influx of dense water from across the equator into the basin. Over the Southern Hemisphere, there is a general increase in the PV contrast, which in the Pacific is achieved by an increase in the stratification of the source waters and in the Atlantic is partly achieved through a strengthening in the northward flux of AABW.

4. Tracer Release Experiment

[27] In order to support the hypothesis that the overturning circulation controls the PV distribution, we now selectively release an artificial tracer and examine how its
spreading pattern evolves compared with the PV distribution for each basin.

4.1. Formulation of Tracer Experiment

[28] The tracer release experiments are conducted using offline, three-dimensional velocities taken from snapshots every 5 days over a 10 year period from the GCM. The velocities are from the sum of the instantaneous Eulerian velocity plus the eddy transport (or “bolus”) velocity using the Visbeck et al. [1997] closure. The tracer equation is solved using a flux limiter advection scheme [Stratford, 1999], which avoids spurious overshoots and limits artificial

Figure 8. Spreading of an idealized tracer released in the Equatorial Ocean for (a) Pacific and (b) Atlantic basins. The tracer source has a value of 1, and the tracer is advected for 100 years using velocities for a decade starting from years 30 or 1000. The tracer distributions are displayed as a zonally averaged section and a map of the tracer along the deep neutral surfaces (as shown by dashed lines in Figures 5a and 6a). Tracer concentrations greater than 0.2 and 0.05 are denoted by medium and light shading, respectively, and where the neutral surface has ground is denoted by dark shading.
Figure 9. Spreading of an idealized tracer released in the Southern Ocean for (a) Pacific and (b) Atlantic basins. The tracer source has a value of 1, and the tracer is advected for 100 years using velocities for a decade starting from years 30 or 1000. The tracer distributions are displayed as a zonally averaged section and a map of the tracer along the deep neutral surfaces (as shown by dashed lines in Figures 5a and 6a). Tracer concentrations greater than 0.2 and 0.05 are denoted by dark and light shading, respectively, and where the neutral surface has ground by dark shading.
diffusion. Since the tracer is used to identify the advective pathways, the tracer equation does not include any diapycnic or isopycnic diffusion (unlike in the original GCM). The spreading patterns of the idealised tracer are calculated up to 100 years using a repeated cycle of the 10 year velocities from the GCM.

4.2. Transient Tracer Distributions

[29] The tracer experiments are conducted with different source distributions. Firstly, the tracer is released along the equator to identify how the cross-equatorial transfer alters for different PV distributions. The tracer is injected in a 5° band along the equator into a particular model level at a depth of 4 km, which spreads into an initially untainted background. Over the 100 year integration, the tracer spreads over the whole of the North Pacific when advected by velocities from year 30 to 40, but is relatively confined to the tropics when the velocities are from the decade in year 1000; see the zonally integrated section and map of tracer concentration in Figure 8a. In contrast, over the Atlantic, there is a relatively restricted spreading using velocities from year 30, compared with an enhanced spreading into the northern Atlantic when the velocities are from year 1000 (Figure 8b).

[30] Secondly, the tracer is released in a band between 50°–60°S at a depth of 4 km in the Southern Ocean in order to mimic the spreading of AABW from the Southern Ocean. The tracer spreading patterns are broadly consistent with the more artificial equatorial case. Using velocities from year 30 to 40, the tracer spreads much further north from the Southern Ocean in the Pacific, than in the Atlantic (Figures 9a and 9b). However, using velocities for the decade in year 1000, there is an opposing pattern with more northward spreading in the Atlantic, than the Pacific. In addition, a tracer released in the North Atlantic at 50°N at a depth of 4 km reveals greater southwards spreading using velocities from year 30, than from year 1000 (shown by dashed contours in the section in Figure 9b). This result is consistent with the weakening in the export of NADW, compared with the strengthening in the influx of AABW between years 30 to 1000.

[31] These tracer diagnostics support the view that the increase in density of the bottom waters in the Atlantic and Pacific is achieved through an influx of bottom water from the Southern Ocean. This influx into the North Pacific has weakened and nearly ceased by year 1000 leading to nearly flat neutral surfaces and PV gradients determined by the planetary vorticity gradient. Conversely, in the Atlantic, there is still a significant influx of dense fluid into the northern basin with neutral surfaces deepening polewards, which leads to extensive regions of low PV.

[32] The weak contrasts in PV are coincident with where there is an influx of the tracer across the equator. For example, there are extensive regions of low PV and higher tracer concentrations coincident over the North Pacific at year 30 (Figures 5b, 8a, and 9a). As the overturning circulation changes, this correspondence between the low PV and influx of transient tracer instead occurs over the North Atlantic at year 1000 (Figures 6b, 8b, and 9b). In this model experiment, this relationship is due to the overturning circulation controlling the spreading of water masses and hence determining both tracer distributions. In turn, this correspondence suggests that weak contrasts in PV might provide a useful signal in identifying which layers have enhanced ventilation from across the equator.

5. Discussion

[33] Climatological diagnostics of potential vorticity (PV) suggest that there is a rich structure: PV contours deviate from latitude circles and there are even extensive regions of low PV (OW). In this study, we investigate how the PV distribution is controlled by diagnosing output from a coupled model integrated for 1000 years, which is sufficiently long for the model state to lose much of its dependence on the initial climatology. Whatever the realism (or lack of) in the climate model integration, our focus is to assess how the overturning circulation determines the PV contrasts formed in abyssal waters.

[34] In the coupled model integration, there is a drift in the deep water mass properties, probably in response to an overactive hydrological cycle, making the bottom water initially formed in the Southern Ocean too dense and salty. In the Pacific, the deep basin fills up with dense fluid, neutral surfaces flatten and the overturning weakens. The diabatic mixing is too weak to destroy the excessively dense bottom water in the North Pacific. In the Atlantic, the enhanced formation and spreading of AABW leads to an influx of bottom water and grounding of neutral surfaces over the northern basin. Over the integration, the AABW warms through mixing with excessively warm NADW, particularly over the South Atlantic. This additional mixing probably prevents the North Atlantic filling up with dense water and the overturning cell collapsing there.

[35] Whatever the causes of the coupled model drift, there is a consistent response for the PV structure. Weak contrasts in abyssal PV are indeed formed whenever there is a strong inflow across the equator into a northern basin, such as in the North Pacific and the North Atlantic at the start and end of the integrations respectively. This overturning-induced change might also be expected to provide an opposing response in each hemisphere. Assuming that the stratification is gradually eroded with mixing, then a stronger northward flux of bottom water should form a weaker meridional gradient in PV over the Northern Hemisphere, as well as a stronger gradient over the Southern Hemisphere. In this model integration, this response is seen over the Atlantic, but the dominant effect over the Pacific is a drift in the source bottom waters to higher stratification.

[36] In summary, this climate model study does suggest that the PV structure in the deep waters responds to the overturning circulation, although in reality spatial patterns in diabatic mixing are also likely to be important [see, e.g., Polzin et al., 1997]. The PV distribution might be useful as a proxy for identifying abyssal density layers which are strongly ventilated from across the equator. For example, enhanced ventilation of a layer from the Southern Ocean should eventually lead to higher concentrations in transient tracers and weaker contrasts in PV over the northern basins.

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