Decadal Response of Global Circulation to Southern Ocean Zonal Wind Stress Perturbation

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ABSTRACT

A substantial component of North Atlantic Deep Water formation may be driven by westerly wind stress over the Southern Ocean. Variability of this wind stress on decadal time scales may lead to circulation variability far from the forcing region. The Hybrid Coordinate Ocean Model (HYCOM), a numerical ocean model, is used to investigate the spatial patterns and the time scales associated with such wind variability. The evolution of circulation and density anomalies is observed by comparing one 80-yr simulation, forced in part by relatively strong Southern Hemisphere westerlies, with a simulation driven by climatological wind.

The volume transport anomaly takes about 10 yr to reach near-full strength in the entire Southern Hemisphere; however, in the Northern Hemisphere, it grows for the duration of the run. The Southern Hemisphere Indo-Pacific volume transport anomaly is about twice the strength of that found in the Atlantic. In the thermocline, water exits the southern westerlies belt in a broad flow that feeds a western boundary current (WBC) in both the Atlantic and Pacific Oceans. These WBCs in turn feed an Indonesian Throughflow from the Pacific and cyclonic gyres in the far north, which are broadly consistent with the Stommel–Arons theory. The deep return flow in each hemisphere is strongly affected by deep-sea ridges, which leads to a number of midocean “WBCs.” The wind perturbation causes isopycnals to sink over most of the basin. After about 20 yr, this sinking is very roughly uniform with latitude, though it varies by basin.

1. Introduction

The deep meridional overturning circulation (DMOC) of the ocean is intimately connected to global oceanic properties such as heat and chemical transports, temperature and salinity distributions, and western boundary currents. The DMOC is most prominent in the Atlantic Ocean, where approximately 15 Sv (1 Sv = 10^6 m^3 s^-1) of North Atlantic Deep Water (NADW) is produced in the subarctic and flows southward to the rest of the World Ocean (see, e.g., Ganachaud and Wunsch 2000).

Traditionally, the DMOC has been described as a thermohaline circulation, driven by pressure gradients produced by subsurface mixing and meridional surface density differences (Warren 1981; Bryan 1987; Marotzke and Willebrand 1991). The density-driven component of the DMOC appears to be augmented by a component driven by the subpolar westerly wind over the Southern Ocean (Toggweiler and Samuels 1995, 1998). Increasing the wind stress increases the amount of deep-water formation in the northern North Atlantic as well as the amount of deep water flowing out of the Atlantic at 30°S. The relative contributions of the wind and thermohaline components are not well constrained by observations; probably both components are important. Northward Ekman transport in the Southern Ocean feeds the upper limb of the Atlantic DMOC, while a deeper geostrophic return flow is fed by the lower limb. Subpolar wind in either hemisphere can affect the global DMOC (Tsujino and Suginohara 1999; Klinger et al. 2003, 2004), but the Southern Ocean wind dominates because of the absence of landmasses at the latitude of the Drake Passage, around 60°S.

The apparent importance of wind stress in the Southern Ocean for the steady DMOC implies that variability in the Southern Ocean can also influence the global circulation. The adjustment of the advective–diffusive balance in the thermocline is associated with time scales of centuries to millennia. However, an initial disturbance can...
reach all latitudes via coastal and equatorial Kelvin waves (Wajisowicz and Gill 1986; Kawase 1987) in a matter of months, assuming gravity wave speeds of $O(1$ $m^{-1})$ associated with the first baroclinic mode. The disturbances can spread into the interior oceans via Rossby waves that have time scales of about a decade at mid-latitudes for typical ocean basin widths (Wajisowicz 1986; Winton 1996). Therefore, it is possible that variability in the Southern Ocean wind stress on interannual-to-multidecadal time scales can have a significant effect on the global circulation.

Meteorological records over the subpolar Southern Hemisphere are sparse compared to Northern Hemisphere data. However, there is some indication of significant interannual-to-decadal variability in the Southern Hemisphere surface winds (Chen and Yen 1997), including changes in the intensity and latitude band of the subpolar westerlies (Van Loon et al. 1993, see their Figs. 8 and 9). Reanalysis data indicates a recent multidecadal strengthening trend in Southern Ocean wind speed (Renwick 2004), which appears to be related to ozone depletion (Thompson and Solomon 2002). The ocean–atmosphere model of Hall and Visbeck (2002) displays significant variability in the Southern Hemisphere annual mode, which controls the zonal wind stress in the Southern Ocean.

How would such wind variability affect the overturning circulation? McDermott (1996) examines the response of idealized ocean basins to a switched-on perturbation in the southern subpolar westerlies. In a single-basin ocean, the wind perturbation drives an overturning anomaly that fills the Southern Hemisphere after 6 yr and fills the entire basin by 50 yr. McDermott conducted further experiments with a two-basin idealization of the combined Atlantic and Pacific Oceans, but these experiments ran for a relatively short duration. Brix and Gerdes (2003) look at a similar switch-on experiment with a more realistic global configuration. In their experiments, the Atlantic overturning anomaly reaches the subpolar North Atlantic by 10 yr and strengthens by 20 yr.

When surface density is strongly constrained by restoring, we expect the steady overturning anomaly due to the Southern Ocean westerlies to be almost exclusively associated with the Atlantic Ocean, because in the steady state, thermostatic constraints force the wind-driven overturning to “feed” the northern deep-water production that exists in the Atlantic but not in the Pacific. This was Toggweiler and Samuel’s (1995) original explanation for the Drake Passage Effect—northward Ekman flow at the surface of the Southern Ocean must be balanced by a deep geostrophic southward flow that originates in the northern North Atlantic. In the evolving circulation, there is no such constraint, and it is possible for the southern westerlies to drive substantial overturning in both the Atlantic and Pacific. In fact, the global and Atlantic-only overturning anomalies shown by Brix and Gerdes (2003) imply significant volume transport into the Pacific on decadal time scales. If surface density is allowed to vary in response to the wind, then wind-induced density variations can in turn generate steady overturning anomalies in the Pacific (Toggweiler et al. 2006; de Boer et al. 2008; Rahmstorf and England 1997); however, here we are analyzing the response when the density effects are excluded.

Johnson and Marshall (2002, 2004) examine the behavior of the meridional overturning in a single-basin and multiple-basin ocean represented by a 1.5-layer stratification. They force the circulation with an imposed northern North Atlantic inflow to represent the switch on of the thermohaline circulation and study the evolution toward a final state in which there is no upwelling within the Atlantic itself; all the deep northern inflow exits from the southern boundary and is balanced by a southern inflow into the top layer. This model bears some resemblance to the wind switch-on problem discussed earlier. Johnson and Marshall find that by assuming that Kelvin wave and Rossby wave dynamics govern the evolution, analytic and quasi-analytic solutions of the resulting simplified equations actually show a range of time scales for the system to reach its final steady state. Close to the flow source, the system approaches equilibrium within a few years, whereas in the Southern Hemisphere (farther from the source), it takes a number of decades to approach equilibrium.

Hsieh and Bryan (1996) used a shallow-water model to study the pycnocline and sea level response to global warming imposed throughout the water column at high latitudes. The time scale of the signal was somewhat obscured by the continuous nature of the forcing. Huang et al. (2000) also used a shallow-water model to study the response of the global circulation to a change in the rate of deep-water formation. Cessi et al. (2004) showed that a shallow-water model gives response time scales similar to a general circulation model when simulating a periodic NADW signal. Cessi and Otteghuy (2003) show that the global response of a shallow-water model to periodic wind forcing is best described by basin-wide modes.

The goal of our paper is to take a closer look at the evolution of global circulation anomalies due to a change in Southern Ocean wind. We take a near-global ocean model with realistic topography and forcing and impose a switched-on increase in the Southern Hemisphere westerly wind stress. In the studies discussed earlier, McDermott (1996) uses an idealized configuration that
focuses on a single-basin ocean, and both McDermott (1996) and Brix and Gerdes (2003) show a relatively small number of snapshots to summarize the entire adjustment process. Johnson and Marshall (2002, 2004) offer a more comprehensive look at the propagation of overturning anomalies through the basins but do so in the context of a very idealized density structure and a somewhat different driving mechanism. We look at the evolution in both time and space of overturning in the Atlantic, Pacific, and Indian Oceans to better compare overturning anomalies in the different basins and to see how quickly the signal propagates into each basin. We also characterize the three-dimensional patterns of current anomalies and isopycnal displacements associated with the overturning anomalies. Because meridional overturning is difficult to measure in the real ocean, anomalies in these latter quantities are more likely to be observed.

2. Numerical model experimental configuration

The experiment was conducted with Hybrid Coordinate Ocean Model (HYCOM) version 2.1.03, a C-grid numerical ocean model with a hybrid vertical coordinate that can represent depth, density, or water column fraction in different parts of the same domain (Bleck 2002). It has seen increasing use for general circulation studies (Chassignet et al. 2003; Sun and Hansen 2003; Shaji et al. 2005; Sun and Bleck 2006). One advantage of the C grid in studying the time evolution of the flow is that it better represents the Kelvin wave propagation speed, unlike the B-grid, which greatly slows Kelvin waves at coarse resolution (Hsieh et al. 1983). For a full description of HYCOM, see Bleck et al. (2002) and Wallcraft et al. (2003).

In our experiment, each layer \( n \) is assigned a target density \( \rho_n \). The layer density \( \rho_n(x, y, t) \) is set to the target density unless that would make the layer thinner than a minimum thickness \( D_n \). This occurs in near-surface layers where the forcing makes the surface denser than some of the \( \rho_n \)’s, which in a purely isopycnal model would make those layers have zero thickness. In such regions, layer \( h_n \) is set to \( D_n \), and HYCOM behaves like a fixed-level model there. We set \( D_1 = 2 \) m and \( D_n = 1.198 D_{n-1} \). The model also calculates potential temperature \( \theta_n(x, y, t) \) and salinity \( S_n(x, y, t) \), which are related to density using \( \sigma_2 \); thermobaric compressibility is ignored. There are 25 layers, with target densities \( \rho_n \) for selected layers corresponding to the following \( \sigma_2 \) values: 27.8 \( (n = 1) \), 34.44 \( (n = 4) \), 34.64 \( (n = 8) \), 36.59 \( (n = 12) \), 37.09 \( (n = 20) \), and 37.24 \( (n = 25) \).

The horizontal grid spacing is 1.5° in longitude and (1.5°) \( \cos(\text{latitude}) \) in latitude. The domain is global between latitudes 71.82°S and 63.34°N, with solid boundaries in the north and south. Model topography is constructed from 5-minute gridded elevations—bathymetry for the world (ETOPO5) \( 1/12^\circ \) data. The ETOPO5 data are smoothed by taking a weighted average of neighboring points, so that the bottom depth \( B \) at latitude \( \phi \) and longitude \( \lambda \) is calculated from depths \( B_j \) of nearby locations (\( \phi_j, \lambda_j \)) using

\[
B(\phi, \lambda) = \frac{\sum_j B_j \exp[-(\phi - \phi_j)^2 + (\lambda - \lambda_j)^2/R^2]}{\sum_j \exp[-(\phi - \phi_j)^2 + (\lambda - \lambda_j)^2/R^2]},
\]

where \( R = 3.6° \).

In the model, ocean regions with depths of no more than 100 m are turned into land. This removes, for example, the shallow (<20 m) passage between Australia and New Guinea. Similarly, Borneo and Sumatra are attached to Southeast Asia. A significant region for interocean communication is the Indonesian waters between Australia and Borneo, which contain many islands and other complex topography. Any pathway between the Pacific and Indian Oceans through this region must encounter a ridge of about 1000 m or shallower. In the model’s relatively coarse representation of this “Indonesian passage,” the islands are eliminated and the minimum sill depth between the two basins in the Indonesian region is about 1500–1800 m.

We further adjust the topography after smoothing. The Mediterranean and some smaller marginal seas are eliminated and east Asian topography is simplified, with Japan fused to Asia. Smoothing opens passages between the Caribbean and the Pacific and submerges New Zealand and the Philippines. We restore more realistic features to these areas. Smoothing greatly lowers the South Sandwich Island ridge just to the east of the Drake Passage. We restore the ridge to a sill depth of 1–3 km.

The model is run with \( K \) profile parameterization (KPP; Large et al. 1994) vertical mixing, with background diffusivity of \( 10^{-5} \) m² s⁻¹ and background viscosity of \( 10^{-4} \) m² s⁻¹. “Horizontal” (within a model layer) viscosity includes both Laplacian and biharmonic terms. Background eddy viscosities, \( \nu_I \) \( (I = 2 \) for Laplacian and \( I = 4 \) for biharmonic), depend on grid spacing and are given by

\[
\nu_I = a_I \Delta x^{I-1}.
\]

As noted above, the grid spacing \( \Delta x \) varies with latitude; however, for a representative value of \( \Delta x = 10^5 \) m, our values of \( a_2 = 0.2 \) m s⁻¹ and \( a_4 = 0.05 \) m s⁻¹ yield \( \nu_2 = 2 \times 10^4 \) m² s⁻¹ and \( \nu_4 = 5 \times 10^{15} \) m⁴ s⁻¹. The Laplacian
eddy viscosity is the sum of the background term and a deformation-dependent term,

\[ \nu_{D_{ij}} = \Lambda \Delta x^2 \sqrt{(u_x - v_y)^2 + (v_x + u_y)^2} \]  

(Bleck et al. 2002), where subscripted \( x \) and \( y \) represent \( \partial/\partial x \) and \( \partial/\partial y \). We use \( \Lambda = 0.2 \). The model also uses a Laplacian diffusion for layer thickness. The thickness diffusivity \( a_2 \) is given by a formula analogous to (2), with \( a_2 = 0.01 \text{ m s}^{-1} \). Finally, there is horizontal Laplacian diffusion of temperature and salinity, with \( a_2 = 0.03 \text{ m s}^{-1} \). To reduce gridpoint-scale numerical oscillations in the horizontal velocity fields, a targeted Shapiro filter (Klinger et al. 2006) is applied, with updates every 10 time steps.

The model initial condition consists of temperature and salinity fields interpolated from World Ocean Atlas 1994 (Conkright et al. 1994; see http://www.cdc.noaa.gov/) monthly climatology. Surface forcing consists of surface heat fluxes based on bulk formula and Comprehensive Ocean–Atmosphere Data Set (COADS) data as well as restoring surface temperature and salinity to World Ocean Atlas monthly climatological values (30-day time scale, top 30 m). Wind stress (Fig. 1) is also based on COADS. Buffer layers near the northern and southern boundaries of the domain restore temperature and salinity toward the monthly climatology.

HYCOM uses a separate integration process for barotropic streamfunction and for all other fields. For each time step, the barotropic streamfunction is stepped forward 20 times (with time steps \( \frac{1}{20} \) of the standard time step). The standard time step for integration in these experiments is 0.5 h. For some parts of the initial integration (see next section), the time step is four hours for temperature, salinity, and interface depth, and 30 min for baroclinic components of velocity (Bryan 1984). Use of unequal time steps allows the model to approach a final state with fewer time steps, because the evolution is governed by the long density time steps, whereas the most stringent Courant–Friedrichs–Lewy (CFL) condition is satisfied by shorter momentum time steps.

3. Numerical experiments

We perform two simulations. In the base run, we time step the model toward a seasonally varying quasi-steady state using the climatological forcing. In the perturbation run, we use output from the base run as the initial condition for an experiment with a zonal wind stress multiplied by a factor of 1.5 south of 30°S (Fig. 1).

The base run is integrated for 260 model years. A model year is defined as 360 days to simplify diagnostics. Model years 31–70 of this period are integrated with the Bryan (1984) method. For the first 30 yr, the model uses a series of different parameter choices. Up to year 110, the model is integrated with biharmonic thickness diffusion, with \( a_4 = 0.05 \text{ m s}^{-1} \) in (2). This integration shows a surprising amount of time variability. The model generates “eddies” with length scales of 500–1000 km, time scales of decades, and interface excursions of tens of meters. These eddies are strongest in the Southern Ocean and near coastal boundaries. Dawe and Thompson (2005) argue that similar eddies in their numerical model occur because viscosity and diffusivity suppress baroclinic instability on appropriate scales of about 100 km and a few weeks while allowing eddies to develop at inappropriately large space and time scales. W. Large (2005, personal communication) also found a similar phenomenon.

We continued the base run with Laplacian rather than biharmonic diffusivity, as described in the previous section. This greatly reduces the decadal eddies. The model runs in this configuration from model year 110 to model year 260. It is preferable for the annually averaged model fields to reach a steady state to use as an initial condition for the perturbation experiment; however, as described below, drift in some model fields continued for the duration of the base run. Rather than use the computer resources necessary to continue the experiment for another half a millennium or more needed to reach steady state, we chose to do the perturbation experiment as a “twin.” Base run year \( t_0 = 180 \text{ yr} \) is the initial condition for the perturbation run. Each anomaly field at time \( t \) of the perturbation run is calculated by subtracting the corresponding field from the base run at time \( t_0 + t \). The perturbation experiment runs for 80 yr.
For the base run, the mean meridional overturning streamfunction $\Phi_z(y, z)$ (Fig. 2) shows the familiar features of the real world. In the Atlantic (Fig. 2a), sinking in the far north is associated with a NADW cell, and northward bottom flow is associated with an Antarctic Bottom Water cell. In the Indo-Pacific (Fig. 2b), there is not much northern deep-water forcing, though there is a relatively large inflow of bottom water from the Antarctic. Surface Ekman transport drives subtropical cells in both basins, particularly in the Indo-Pacific. In the global integral (Fig. 2c), the Ekman-driven Deacon cell in the Southern Ocean is also visible. The NADW cell has a peak overturning of about 12 Sv and a transport near 35°S (the latitude of the tip of South Africa) of about 7 Sv; both values are significantly weaker than over the last century of the run, with speeds ranging as in the real ocean.

Between layers are very small for deep layers, the bulk of the stratification occurs in the top half kilometer or so, as in the real ocean.

We measure the time evolution of $\Phi_z$ via three indices: NADW transport, Atlantic outflow, and Pacific outflow. The NADW outflow shows the clearest sign of nearing a steady state, with the rate of change decreasing from about 0.5 Sv decade$^{-1}$ (years 160–180) to about 0.05 Sv decade$^{-1}$ (years 240–260). The Atlantic outflow trend is similarly small by years 240–260, though a decrease in the rate of change is not apparent. The Pacific outflow has a relatively constant trend of about 0.3 Sv decade$^{-1}$ over the last 60 yr. Thus it is likely that the eventual model steady state would have a significantly weaker South Pacific overturning than the one seen in Fig. 2b.

Below about 500 m, layer interfaces continue to rise over the last century of the run, with speeds ranging from about 10 m decade$^{-1}$ at 1000 m to 30 m decade$^{-1}$ at 4000 m. There is some trend toward decrease in these rates with time, especially in the top half of the water column, but it is likely that, below the pycnocline, the model isopycnals are typically 200 m or more deeper than their final steady depths would be.

### 4. Meridional overturning anomaly

Snapshots of the meridional overturning streamfunction anomaly (relative to the unperturbed experiment) show key aspects of the evolution of the circulation after the westerlies are strengthened in the Southern Ocean. The global anomaly extends over most of the water column, with Ekman transport feeding a northward flow in the top kilometer and a southward flow below (Fig. 3). The anomaly has reached the equator by year 5, expanded well into the Northern Hemisphere by year 20, and continues to grow (year 79). These features are similar to those found by Brix and Gerdes (2003, see their Fig. 13). However, in their experiment the downwelling signal is much stronger north of about 30°N by 20 yr, and there is a northern recirculating cell at 10 yr that is absent in our experiment (not shown). It is interesting that our experiment shows these signs of a slower evolution because the resolution is higher and their B-grid model would be expected to have slower Kelvin wave propagation than our C-grid model.

We also calculate the streamfunction based on integration along isopycnals rather than along constant depths. The anomalies show very similar patterns and magnitudes (Fig. 3) to the $z$-coordinate overturning, though there is a tendency for the signal to be stronger at its northernmost region.

Comparison of Indo-Pacific and Atlantic $z$-coordinate overturning streamfunctions (Fig. 4) reveals a very interesting difference between the two basins. The Indo-Pacific signal is much stronger than the Atlantic signal, especially in the 20-yr snapshot. This is the opposite of the equilibrium response, which we expect to be dominated by the Atlantic. The initial response may be related to the fact that the Ekman transport into the Indo-Pacific is larger than that of the Atlantic, simply because the Atlantic is so much narrower. The vertical structures of the anomalies are broadly similar between the two basins. The density-coordinate overturning anomaly has similar features and magnitudes as the $z$-coordinate anomaly.

For a more detailed look at the time evolution of the overturning anomaly, we plot contours of the maximum $z$-coordinate streamfunction anomaly for each latitude and time (Fig. 5). Features are similar in the $\rho$-coordinate overturning (not shown). The annual average overturning has a significant amount of small-scale “noise” (Fig. 5a), which we remove with a moderate amount of temporal and spatial smoothing (Fig. 5b). The global overturning anomaly has, broadly speaking, different behavior in three different latitude bands (Fig. 5b). From about 60°S to 30°S, the wind stress directly forces the overturning and the annually averaged volume transport adjusts immediately to a value which then stays constant over the 80-yr run. The rest of the Southern Hemisphere also adjusts quickly to a quasi-steady state but with a period of around eight years. The adjustment period for
the Northern Hemisphere, in contrast, appears to be a number of decades. There are some similarities to the evolution calculated in Johnson and Marshall (2002), who also found fast time scales in the hemisphere in which the overturning was initiated and slower time scales in the other hemisphere. In their model, volume transport in the fast hemisphere reached its equilibrium value on the time scale of a Kelvin wave transit (a few months), then decreased 5%–25% over about 10 yr, and increased to the final value on the time scale of decades. The other hemisphere reached a small value on the Kelvin wave time scale and then increased to near the final value over a few decades.

The time evolution of the Indo-Pacific (Fig. 5c) and Atlantic (Fig. 5d) reflect the Southern Hemisphere differences that were visible in the snapshots (Fig. 4). In the Indo-Pacific, the Southern Hemisphere transport quickly (in about one decade) attains its near-maximum value, after which it becomes relatively steady with some sign of decline after several decades. We expect the values to decline eventually to an equilibrium state with a relatively small Indo-Pacific anomaly. The Indo-Pacific streamfunction is stronger at about 20°S than it is immediately to the south, indicating upwelling just north of the wind perturbation. In contrast to the Indo-Pacific, the strengthening of the Southern Hemisphere’s Atlantic transport anomaly continues over the entire run, though it too strengthens the most over the first decade. The two basins have similar volume transport behaviors in the Northern Hemisphere. Both the Atlantic and Indo-Pacific have volume transports that are relatively uniform throughout most of the southern latitude (at any given time), but the Indo-Pacific has a strong gradient from about 5°S to 10°N that is lacking in the Atlantic. It is puzzling why this discrepancy exists. It may be due to the unique geometry of the Indo-Pacific, where the Pacific and Indian are separated by Australia but communicate through the equatorial channel at Indonesia. The Northern Hemisphere does not reach steady state during the experiment, but it is interesting that even after 80 yr the Atlantic transport is only slightly larger than the Pacific.

5. Density structure anomaly

The evolution of density is interesting because in the real ocean, it is somewhat more straightforward to measure density than to measure overturning. Numerically, HYCOM uses hybrid layers in which density can vary. However, below the surface, density is generally constant within a layer, so the evolution of the subsurface layer interface depths is a good measure of the density evolution.

As with overturning, the density field is eventually modified in regions far from the wind perturbation. In the first five years, changes in isopycnal depths are largely confined to the latitudes of the wind perturbation (Fig. 6, top panels), except near the northern boundary of the Pacific Ocean, where deep isopycnals undergo significant excursions. The northern boundary evolution appears to be a manifestation of intrinsic variability of the model. In the ocean directly below the wind stress perturbation, the isopycnals move in a direction consistent with Ekman pumping, with shallowing in the upwelling region around 60°S and deepening in the downwelling region around 40°S. As the downwelling reaches farther northward (Figs. 3 and 4), the isopycnals also deepen in regions north of the wind stress perturbation. By 20 years, the deepening extends throughout the Southern Hemisphere and past the equator (Fig. 6, middle), and by 79 yr virtually the entire domain is affected (Fig. 6, bottom panels).
The overall pattern is similar in all three basins, but there are differences as well. The magnitudes of the anomalies are different in all three basins, with the Indian Basin tending to be the strongest. The isopycnal perturbations show a surprisingly rich vertical structure. In the Pacific, the largest isopycnal excursions are in layer 20 and deeper at 20 yr and at layers 11 to 15 at 79 yr. The Indian and Atlantic basins show two or even three simultaneous local maxima in isopycnal excursion at 20 and 79 yr.

For a closer look at the evolution of isopycnal depth in time and latitude, we examine the interface at the bottom of layer 12 (Fig. 7). Typical depths of this interface are about 1000 m (see Fig. 2 for interface locations). The latitude dependence of the depth anomaly north of 30°S is established within about 20 yr, and the rest of the experiment can be roughly characterized as a fixed profile uniformly sinking over time (Figs. 7a–c). There is some further adjustment near the northern boundary of each basin over the last 40 yr, however. The shape of the depth anomaly south of 30°S is less constant. Its evolution can best be viewed as a function of time in Fig. 7d. The maximum shallowing of the interface in the far south, which is at different latitudes for different basins (54°–60°S), shows a fast rise for the first decade or two followed by a leveling off and a slight decline. The maximum deepening, at about 40°S in each basin, shows a fast increase over the first 5–10 yr followed by different behavior in different basins. In the Atlantic, the depth anomaly becomes somewhat smaller.
and then levels off, while in the Indian and Pacific the depth anomaly continues deepening at a rate comparable to the deepening at the equator and 30°N.

6. Horizontal velocity patterns

a. Methodology and significance test

The meridional overturning streamfunction obscures the horizontal structure of the velocity anomaly field. The overturning anomaly shown in Figs. 3 and 4 has a relatively simple vertical structure, with generally northward flow anomaly over the top kilometer or top 13 layers of the model (Fig. 2) and generally southward flow anomaly over the rest of the water column. Thus there is some justification for looking at the horizontal flow in two layers: a “top-layer velocity” integrated over the top 13 model layers and a “bottom-layer velocity” integrated over the bottom 12 model layers. In this section we refer to these as “velocities” for brevity, even though they are actually vertically integrated velocity anomalies. Our investigation of velocities focuses on the area north of 30°S—that is, north of the area of direct forcing by the wind stress perturbation.

Velocities time averaged over individual years show significant year-to-year differences in both direction and magnitude, especially in the tropics. Here we want to focus on decadal-scale variability through a few “snapshots” consisting of 3–4-yr averages. Even averaging over a few years, it is not immediately clear whether a given velocity vector is simply a residual of the interannual variability or a genuine indicator of the decadal trend. We therefore provide a crude test of the significance of each vector.

We assume that our “measurement” is of velocity on roughly decadal time scales, whereas the “error” is due to variability on time scales of up to a few years. We calculate top- and bottom-layer velocities for years 3–5, 16–19, and 39–41 to estimate time-average velocities in each of those intervals (Figs. 8 and 11). In the figures, velocity arrows are put into three categories based on an estimate of significance, where speed $U = \sqrt{u^2 + v^2}$ and

![Fig. 4. $\Phi_z$ Anomaly for (left) Indo-Pacific and (right) Atlantic basins, for 3-yr average centered at years (top) 5, (middle) 20, and (bottom) 79. Contour interval is 0.5 Sv, with contours thickened at 1-Sv intervals.](http://journals.ametsoc.org/jpo/article-pdf/39/8/1888/4501661/2009jpo4070_1.pdf)
error estimate $\sigma_U$: 1) “least significant” if $U < \sigma_U$; 2) “somewhat significant” if $\sigma_U < U < 2\sigma_U$; and 3) “most significant” if $U > 2\sigma_U$.

We estimate $\sigma_U$ as follows: The error in annually averaged velocity components $u$ and $v$ at some model grid point is $\sigma_x$ and $\sigma_y$, respectively. To estimate this error, we take the standard deviations of $u$ and $v$ over each of the three time intervals (years 3–5, etc.) and take the average of the three estimates of $\sigma_x$ and $\sigma_y$. Velocity averaged over $N$ years then has an expected error of $(\sigma_u, \sigma_v) = (\sigma_x, \sigma_y)/\sqrt{N}$. The RMS error associated with $U$ is then given by

$$\sigma_U^2 = \left(\frac{\partial U}{\partial u}\right)^2 \sigma_u^2 + \left(\frac{\partial U}{\partial v}\right)^2 \sigma_v^2,$$

which gives us

$$\sigma_U = \sqrt{u^2 \sigma_u^2 + v^2 \sigma_v^2}.$$

One could assign a measure of significance to individual components of velocity instead, but the errors are typically of similar magnitude in the zonal and meridional directions. Given the relatively crude nature of this error analysis, the extra detail does not seem to be warranted.

b. Results

We focus on circulation north of $30^\circ$S, where flow anomalies are caused by remote wind perturbations south of $30^\circ$S. The local wind-driven velocity anomalies
south of 30°S are generally stronger than the remotely forced anomalies.

In his idealized, single-basin simulation, McDermott (1996, his Fig. 16) displayed near-surface velocity anomalies at 4 and 50 yr. These snapshots showed a northward-flowing western boundary current (WBC) that reached to about 25°N at 4 yr and to the northern boundary by 50 yr. North of about 10°N, an interior path also flowed northeastward. At all latitudes shown, there tended to be northeastward flow. As described below, some similarities can be seen in our model in the Atlantic, which is analogous to the single-basin McDermott showed.

Our first snapshot, centered on four years (Fig. 8a), shows the top-layer velocity field at a time when the overturning perturbation is largely confined to the Southern Hemisphere (Figs. 3 and 4). Note that the figure exaggerates weaker speeds relative to stronger speeds so that both intense jets and broad flows can be seen. It should also be remembered that the circulation anomalies are perturbations to an upper-ocean circulation dominated by the wind-driven gyres, and they don’t generally change the direction of the flow. The most prominent currents consist of westward flow just north of 30°S in all basins and northward WBCs in the Southern
Hemisphere. The eastward-flowing Antarctic Circumpolar Current also strengthens (not shown). The South Pacific WBC feeds a current flowing into the Indian Ocean through the Indonesian Passage. This flow continues as a zonal current across the Indian Ocean. It is also interesting that even on this short time scale, there is a velocity signal throughout the Northern Hemisphere, though it clearly stands out from the interannual variability only in a limited region of the North Atlantic.

In the 16–19-yr average (Fig. 8b), the Southern Hemisphere circulation is similar to the earlier snapshot, but the zonal continuation of the Indonesian Throughflow into the Indian Ocean is no longer apparent. The Atlantic WBC continues into the Northern Hemisphere but separates from the continent near Cape Hatteras and continues northward to feed a cyclonic gyre in the subpolar North Atlantic. In the Pacific, a broad northward current also feeds a cyclonic subpolar gyre. These cyclones are in marked contrast to the steady-state response to strengthened Southern Hemisphere westerlies (Klinger et al. 2003) and to McDermott (1996), in which the WBC does not separate. There is also eastward flow over much of the Atlantic and Pacific between 20°S and 40°N. In the tropics the velocities are generally not significant according to the criteria discussed above; however, in the Pacific, the consistency of the eastward flow (seen in most tropical grid points in all three panels of the figure) suggests that the eastward tendency is a real feature.

In the 39–41-yr average (Fig. 8c), the flow pattern is similar to, but stronger than, the 16–19-yr average. The westward jet in the Indian Ocean has also reestablished itself. An 80-yr snapshot (not shown) also has similar flow patterns.

Figure 8 shows that in the top layer, some latitude bands are dominated by WBCs. It is expected that WBCs would have higher velocities than the flow spanning the rest of the ocean. However, the broad, weak flows can make a large contribution to the volume transport, as in the case of the Sverdrup theory. What is the relative contribution of the two flow regions in our top-layer velocity anomalies?

To answer this question, we calculate the volume transport in a “western” region and an “eastern” region. The border between the two regions is 1°–2° to the east of each ocean’s western boundary. The border is moved eastward near the equator in the Pacific, where the WBC is far east of Southeast Asia as it flows around the northern coast of New Guinea (see Fig. 8) and in the Caribbean, where the model resolution does not allow the WBC to flow into and out of the Gulf of Mexico.

The volume transports calculated in this way can be divided into three latitude bands. In all three basins (Fig. 9), there is a broad interior northward flow and a southward WBC flow south of 20°S. In the Atlantic and the Pacific, the WBC does account for most of the northward volume transport from about 20°S to about 20°N (Figs. 9b,c). North of that region, the interior contribution is increasingly important.

The qualitative flow patterns—northward-flowing western boundary currents in the Southern Hemisphere and cyclonic gyres in the north of the Northern Hemisphere—are consistent with the Stommel–Arons theory (Stommel and Arons 1960). There must be a northward flow to feed the Northern Hemisphere downwelling. Downwelling is associated with an interior flow that is southward in the Southern Hemisphere and northward in the Northern Hemisphere. Also, downwelling is small in the Southern Hemisphere. For these reasons the northward transport must be carried by the WBC in the Southern Hemisphere, whereas in the Northern Hemisphere, both the WBC and interior flow can carry the northern transport. As one looks at a latitude far enough to the north, the interior transport is actually larger than is needed to feed the downward transport to the north of the given latitude and the WBC reverses flow. The northward interior flow and southward WBC form the gyres described earlier.
We now test whether the Stommel–Arons theory is quantitatively applicable to the flow anomalies. In the region where the Ekman pumping perturbation is zero, the linear vorticity equation integrated from depth $z_B$ to the surface can be written as
\[ \beta V_I = -fw_B, \]  
where $f$ is the Coriolis parameter, $\beta$ is its gradient with respect to meridional distance, $V_I$ is the "interior" meridional velocity anomaly (neglecting WBCs) integrated from $z_B$ to the surface, and $w_B$ is vertical velocity anomaly at $z = z_B$. If we integrate the equation from the western boundary to the eastern boundary, use the definitions of $f$ and $\beta$, and use the meridional streamfunction to calculate $w_B$, we can rewrite (6) as
\[ T_I = \tan(\phi) \frac{\partial \Phi}{\partial \phi}, \]  
where $T_I$ is the northward volume transport anomaly associated with interior flow, $\phi$ is latitude, and $\Phi$ is evaluated at $z = z_B$ (anomaly only). We compare the estimate of $T_I$ [given by (7) and based on $w_B$] with the value measured from the top 13 layers. In the Pacific,

**Fig. 8.** Velocity anomaly fields vertically integrated over the top 13 layers of the model, averaged over years (a) 3–5, (b) 16–19, and (c) 39–41. Vector shade and weight represents significance of vectors based on size of interannual variability, with thick black representing most significant category, thin gray least significant, and thin black representing intermediate significance (see text for detail). To improve clarity, arrow length is proportional to the square root of vector magnitude, and the field has been smoothed and subsampled to a coarser grid than that used by the model.
the combined thickness of these layers is typically within about 100 m of \( z_B = 2000 \)-m depth, so this value is chosen for evaluating \( \Phi_z \). Choosing year 16–19 as a representative time slice, there is broad agreement between the Stommel–Arons estimate and the actual interior volume transport (Fig. 10). A good deal of the noise in the Stommel–Arons estimate may be due to the large high-frequency WBC variability shown by gray arrows in Fig. 8b. This noise is amplified by taking \( \partial \Phi_z / \partial f \). The Atlantic (not shown) also has broad agreement with strong noise in the Stommel–Arons estimate.

For a purely baroclinic signal, the deep flow should simply be opposite to the relatively shallow flow seen in Fig. 8. In reality, there are some interesting differences, apparently due to bottom topography (Fig. 11). In all three time averages, the South Pacific “WBC” occurs next to the Kermadec Ridge, which reaches northward from New Zealand, rather than against the Australian coast. Observations indicate that the time-mean WBC in the real Pacific Ocean is also located next to this ridge (Hogg 2001). In the model, this WBC is not connected to the Indian Ocean by a strong current because topography blocks the Indonesian Throughflow at these depths.

Other deeper bathymetric features also seem to produce “midbasin” WBCs. In the South Atlantic, there is a WBC next to South America; however, there is also an opposite (northward) flow just to the east of the Mid-Atlantic Ridge, which reaches up to about 3500–4000 m below sea level in the model. In the North Pacific the deeper (typically 4000-m ridge crest in the model) ridge associated with the Hawaii, Marshall, and Gilbert Islands also seems to guide a relatively strong current from the subpolar region to the equator. This current is most prominent in the 16–19-yr average, though there is some sign of it at the other times as well. The bottom layer, as defined in this section, reaches from the bottom up to 1–2-km depth, but the detailed vertical structure of these deep WBCs (not shown) demonstrates that the currents are primarily below the ridge crests. It should also be noted that the smooth model topography is a very crude representation of the actual island arcs in the region. Similarly, because the Ninety East Ridge in the Indian Ocean is a very narrow feature, it is much less prominent in the smoothed model topography. In general, Indian Ocean currents do not show obvious influences from the bathymetry.

The midbasin meridional currents induced by bottom topography obscures zonal currents in the Atlantic and western half of the Pacific. The eastern Pacific does show some counterparts to the top-layer velocities, with
eastward flow near 30°S and some sign of westward flow (largely of the least significant category according to the test described earlier) at other latitudes.

7. Discussion and conclusions

a. Adjustment time scale

Our plots of volume transport anomaly as a function of latitude and time (Fig. 5) invite comparison to a similar plot of overturning strength, Fig. 4 of Johnson and Marshall (2002). Both results concern the initiation of an overturning circulation forced from high latitudes. It is interesting to see to what extent Johnson and Marshall’s simple model (1.5 layers, no diffusivity, no background flow) reproduces the behavior of our model, which has more complete dynamics. However, we must be cautious because of the geometrical differences between the models, with Johnson and Marshall (2002) examining a single basin with volume transport imposed at the northern boundary and with all cross-isopycnal flow occurring outside the model domain, whereas our model forces flow into several basins, all from the south. Johnson and Marshall viewed their model as representing the imposition of a thermohaline flow, but the model is sufficiently abstract that we can also consider it to represent a wind-driven circulation, as in our experiment.

Both models feature marked differences between hemispheres, with a relatively rapid adjustment in the “source” hemisphere in which the flow is imposed and a slower adjustment in the “sink” hemisphere on the
other side of the equator. The adjustment differs somewhat between the two models.

In the sink hemisphere, the volume transport in Johnson and Marshall (2002) approaches the source value \( \Phi_S \), within 50 yr. Initially, their model has a large meridional gradient in volume transport, but this gradient steadily decreases over time. In our experiment, the meridional gradient persists for the entire 80 yr of the run in both the Atlantic and Indo-Pacific. Because the evolution on long time scales must involve diabatic processes, Johnson and Marshall’s adiabatic model is not adequate for estimating the time scales for evolution in the sink hemisphere. (Diabatic processes are assumed to occur outside the open boundaries of the model domain.)

In Johnson and Marshall’s model, the volume transport in the entire source hemisphere reaches the source value in less than three months. This volume transport decreases to a minimum value \( \Phi_M(\phi) \) in a time \( t_M(\phi) \) and then increases toward the source value again. Here \( \Phi_M \) decreases with distance from the source, down to about 0.5 of the source value at the equator, and \( t_M \) ranges from about 10 yr near the source to less than 5 yr at the equator. Our global volume transport (Fig. 5, upper-right panel) reaches a near-constant value in less than a year south of 35°S; however, in the rest of the source hemisphere, it takes 5–10 yr to approach its final value. Unlike Johnson and Marshall’s volume transport, it does not vary much with latitude within this region (say, 30°–5°S) for a few years, though gradients start to appear after about eight years. Similar behavior occurs in the individual Indo-Pacific and Atlantic basins.

The difference in time scales for the response of the source hemisphere between the more idealized model and the more realistic model is an interesting topic for future research. One possibility is that in our model, it takes the imposed Ekman transport roughly a decade to generate a meridional geostrophic flow just to the north of the wind-perturbation latitudes. This contrasts with Johnson and Marshall’s model, in which the meridional flow is directly imposed.

In summary, our experiment confirms Johnson and Marshall’s “equatorial buffer” separating a fast source hemisphere response and a slow sink hemisphere response in a more realistic configuration forced by Southern Ocean wind anomalies. However, the laws governing the time scales within each hemisphere differ from Johnson and Marshall’s case.

**b. Overturning magnitudes**

A key qualitative result of this study is the relative magnitude of the Atlantic and Indo-Pacific basins. In the steady state, a perturbation to the southern west-erlies is expected to have the greatest influence on Atlantic overturning, if we suppress changes to surface density induced by the wind perturbation. On the decadal time scales observed here, however, the overturning anomaly in the Northern Hemisphere Pacific is about the same size as in the Atlantic, and in the Southern Hemisphere, the Indo-Pacific is twice as large as in the Atlantic. This transient behavior may be related to the relatively large width of the Indo-Pacific sector, hence the relatively large zonally integrated Ekman transport there. Even after 79 yr, the overturning anomalies in the two basins are of similar magnitude, suggesting that a large Pacific anomaly may persist in steady state.

An important observational consequence of the decadal results is that measurements of deep meridional overturning, especially in the Pacific, may be complicated by time variability in the Southern Ocean wind field.

In this study, we choose an arbitrary wind perturbation strength, so the overall amplitude of the model anomalies are only meaningful as a measure of the sensitivity of the system to wind. Even within that context, there is great uncertainty. The strength of the wind-driven component of the meridional overturning depends on a number of poorly quantified parameters, notably the strength of eddy mixing in the latitude band of the Drake Passage. The base run used here tends to generate a somewhat weak North Atlantic Deep Water cell compared to observations (Ganachaud and Wunsch 2000; Schmitz and McCartney 1993; Talley et al. 2003). The Miami Isopycnic Coordinate Ocean Model (MICOM), a similar model, tended to have a weaker large-scale steady-state response to Southern Ocean wind perturbations than another model, the Modular Ocean Model (MOM), when the two models were compared in an idealized configuration (Klinger et al. 2003). In the Hallberg Isopycncal Model (HIM), another hybrid-coordinate model, the NADW response to a wind perturbation is significantly larger than in our experiment for similar forcing and geometry (Hallberg and Gnanadesikan 2006, personal communication). The wind climatology used here is also thought to be about 40% too weak in the Southern Ocean compared to other climatologies, such as Quick Scatterometer (QuikSCAT; PO.DAAC 2003). In our experiment, the global overturning anomaly due to a 0.055 N m\(^{-2}\) increase in winds is around 3 Sv over much of the experiment, which is about the same size as the anomaly found by McDermott (1996) on decadal scales for a similar wind perturbation applied to MOM with a “Southern Ocean” only \( \frac{1}{6} \) as wide. The sensitivity of the real-world amplitude to the wind stress of the transient and mean overturning remains an open question.
c. Density response

The overturning anomalies are accompanied by isopycnal displacement as well. In the vicinity of the wind perturbation, there is a fast [O(10 yr)] adjustment in isopycnal depth of 50–100 m, with isopycnals rising in the Ekman upwelling region and sinking in the Ekman downwelling region. Over the rest of the domain, isopycnals sink more steadily at a rate of a few meters per decade over the entire 80-yr simulation. This sinking is probably part of the adjustment toward a state with a thicker thermocline, as would be expected for a stronger overturning (Gnanadesikan 1999).

The deepening of the isopycnals in response to enhanced overturning contrasts with the isopycnal shallowing found by Huang et al. (2000). Their result is probably due to the artificial thermal structure of their model, in which equilibrium isopycnal depth is restored to a predetermined value that is not based on the advective–diffusive balance.

Sinking isopycnals correspond to warming deep water. Therefore, some component of deep decadal and multidecadal warming (seen, for instance, by Levitus et al. 2005; Kawano et al. 2006; Johnson et al. 2007) may be due to strengthening Southern Ocean wind stress.

d. Global velocity distribution

Dividing the water column into two layers (roughly the top kilometer and the abyss), we displayed the horizontal velocity anomalies produced north of the wind perturbation. The surface layer in the Atlantic and Pacific Oceans are each characterized by a broad northward flow at 30°S, feeding a WBC that in turn feeds a broad cyclonic flow in the Northern Hemisphere, including a southward WBC in the north. This contrasts with the result of McDermott (1996), in which the northward-flowing WBC reaches the northern boundary. In the framework of Stomme and Arons (1960), the direction of the WBC depends on the distribution of vertical velocity with latitude. Our results suggest that relatively subtle differences in this distribution between our experiment and McDermott’s can make a qualitative difference in the horizontal circulation.

Part of the Southern Hemisphere Pacific WBC feeds a strong flow into the Indian Ocean. The model Indonesian Passage is broader and deeper than in the real ocean, but it is likely that the real ocean could also support large flow anomalies in the passage, so the effect on the Indian may be realistic.

Throughout the ocean, the deep circulation anomalies are strongly affected by topography. Deep flow in the Indonesian Passage is blocked, and WBCs flow far from the actual coastlines at the Kermadec Ridge, Mid-Atlantic Ridge, and other topographic features. Because many of these ridge systems tend to be aligned north–south, broad zonal flows are less prominent in the abyss than in the top layer.

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