GEOPHYSICAL RESEARCH LETTERS, VOL. 27, NO. 6, PAGES 759-762, MARCH 15, 2000

Global adjustment of the thermocline in response to deepwater formation

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Abstract. The global adjustment of the thermocline in response to deepwater formation is studied in a single mode model on a beta-plane. The signal is carried from ocean to ocean by Kelvin waves, which travel equatorward along western boundaries, eastward across the equator, poleward at the eastern boundaries, and then eastward around the southern tip of continents into the next ocean basin. The interior is filled by Rossby waves emanating from eastern boundaries. Stronger (weaker) deepwater formation induces an upward (downward) motion of the main thermocline in the world oceans. The adjustment is completed on centennial time scales.

1. Introduction

Paleoproxy evidence indicates that production of North Atlantic Deep Water (NADW) was much reduced or even shut down at times in the past [*Oppo and Fairbanks*, 1987] which coincide with rapid changes in global climate. However, *Cane* [1998], citing both theory and general circulation model (GCM) results, argues that the global extent of these rapid changes is evidence against NADW shutdown as the prime cause. The atmospheric response is confined to the Northern Hemisphere. Advection through the deep ocean is too slow and too small in magnitude to have the necessary global impact. *Cane* [1998] overlooked a mechanism [previously suggested by *Döscher et al.*, 1994] that could effect rapid global changes: planetary wave propagation through the world ocean. This letter seeks to elucidate the global adjustment process by studying a reduced gravity model of the global ocean.

2. Model Formulation

We begin with a linear shallow water equation for the upper ocean on an equatorial beta plane [*Kawase*, 1987]:

$$u_t - \beta y v = -g h_x - K u \tag{1}$$

$$v_t + \beta y u = -g h_y - K v \tag{2}$$

$$h_t + H(u_x + v_y) = -\lambda h + Q \tag{3}$$

The continuity equation includes a deepwater source distribution Q and a simple Rayleigh damping $-\lambda h$. These equations can be non-dimensionalized by introducing the following scales for velocity, length, depth, and time:

$$c = (KgH/\lambda)^{1/2}, \quad L = (c/\beta)^{1/2}, \quad H, \quad T = (c\beta)^{-1/2} \tag{4}$$

For $\lambda = K$, this equation set is reduced to

$$u_t - yv = -h_x - ru\ , \quad v_t + yu = -h_y - rv\ , \tag{5}$$

$$h_t + (u_x + v_y) = -rh + Q' \tag{6}$$

where $r = KT$ is the new non-dimensional friction parameter, $Q' = TQ/H$.

Apart from the western boundary, the solution consists of equatorial Kelvin waves leaving the western boundary. At the eastern boundary the Kelvin waves are reflected as westward Rossby waves, which establish the circulation in the interior. Along the wave pathways, the amplitude gradually declines due to dissipation. As the circulation approaches equilibrium, the time-dependent terms drop off, and the steady state solution in an ocean bounded at $X = 0$ at the west and $X = L_B$ at the east is [*Cane*, 1989]:

$$h = AF, \quad F = (\cosh 2r\xi)^{1/2} e^{-y^2/2 \cdot \tanh 2r\xi} \tag{7}$$

where A is a constant for each basin to be determined by mass balance in the model. $F \approx e^{-r\xi y^2}$ for very small r and $F \equiv 1$ along the eastern boundary and the equator, and $\xi = (L_B - X)/L$ is the non-dimensional zonal coordinate.

When r is very small, the mass communication M between two adjacent basins is primarily controlled by the semi-geostrophic current around the southern tip of the continent separating the basins. Denoting its latitude

Paper number 1999GL002365.
0094-8276/00/1999GL002365$05.00

by y^s, $M = \Delta h / y^s$, where Δh is the layer thickness across the boundary current. Thus the mass balance in, say the Indian Ocean, is the influx being balanced by the upwelling inside the basin and the flux out:

$$(A_A - A_I F_{Iw})/y^s_{A,I} = r \iint_{S_I} h\, dxdy + (A_I - A_P F_{Pw})/y^s_{I,P} \tag{8}$$

The subscripts A, I, and P indicate the Atlantic, the Indian, and the Pacific Oceans; S_I is the surface area of the Indian Ocean; $F_{Iw} \approx e^{-r\xi_{Iw} y^s_{I,P} y^s_{I,P}}$, ξ_{Iw} is the western boundary of the Indian Ocean; $y^s_{A,I}$ ($y^s_{I,P}$) is the southern tip of the continent separating the Atlantic and Indian (Indian and Pacific) Oceans. For small friction, $r \iint_{S_I} h\, dxdy \approx r A_I S_I$ to the first order in r. There is a similar relation for the Pacific:

$$(A_I - A_P F_{Pw})/y^s_{I,P} = r \iint_{S_P} h\, dxdy + A_P / y^s_{P,A} \tag{9}$$

In addition, we have the total mass balance in the world ocean:

$$Q' = r(A_A S_A + A_I S_I + A_P S_P) \tag{10}$$

The solution can be found by combining equations (8, 9, 10). When r is very small, the first term on the right-hand side of (8, 9) can be neglected.

3. Numerical Experiments

The numerical model, which uses the INC scheme (Israeli et al., 2000), is integrated using $\Delta x = 1°$ and $\Delta Y = 0.5°$, and a time step of $\Delta t = 10$ hours. The equivalent depth scale is $H = 0.92$ m, and the Rayleigh friction and damping coefficients are $K = \lambda = 4 \times 10^{-10} \mathrm{s}^{-1}$, which corresponds to a non-dimensional friction coefficient $r = 4.8 \times 10^5$. For a mean thermocline depth of 300 m, this corresponds to a diapycnal mixing of 0.36 cm^2 s^{-1}, somewhat larger than the low open ocean mixing rate of 0.1 cm^2 s^{-1}, inferred from tracer release experiments. We choose this larger value for the mixing rate because the basin mean diapycnal mixing inferred from global tracer budget is O(1 cm^2 s^{-1}).

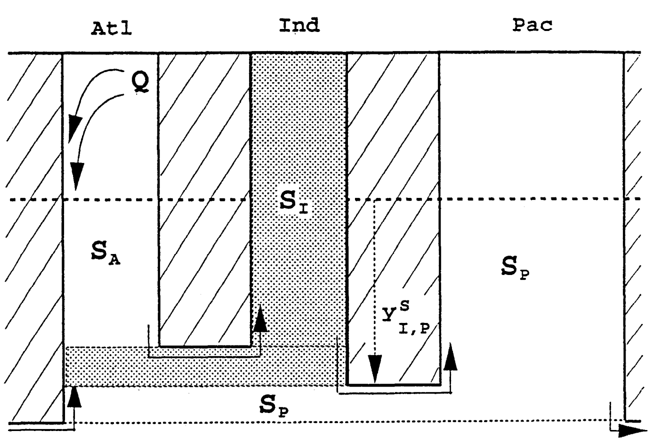

Figure 1. Circulation in the idealized world oceans driven by a point source of deep water Q. S_A, S_I, and S_P are the area of each ocean, and $y^S_{I,P}$ is the southern tip of the continent separating the Indian and Pacific.

A source of 10 Sv, approximately the rate at which NADW currently crosses the equator from the North Atlantic, is uniformly distributed within a band north of 50°N in the North Atlantic. After 500 years the solution seems to approach a quasi-steady state; however, a slow trend of adjustment can be seen from the time evolution of the solution. Note that a slowdown in deepwater formation is equivalent to a source of upper layer water and a sink of lower layer water. Thus, the h field discussed below should be interpreted as the downward motion of the thermocline in response to a slowdown of deepwater formation, or its upward motion in response to increased deepwater formation.

We have run many numerical experiments, including cases in which the world oceans are represented by the highly idealized rectangular basins shown in Figure 1. All results from these numerical experiments fit our simple formulas very closely. Here we show only experiments in which realistic coastlines are used, Figure 2.

First, for the control case, the Indonesian Passage is closed, and corresponding southern tips of the continents are 34°S, 44°S, and 55°S, respectively, and the amplitude of the thermocline depth perturbation in the world oceans is $A_A : A_I : A_P = 105 : 87 : 46$ (m), Figure 2a. Thus, the two-layer model predicts that a deepwater formation rate of 10 Sv induces an upward motion of the thermocline on the order of 50–100 meters. Assuming $r \approx 0$, equations (8, 9, 10) are in error by less than 10%. Errors are due to the approximation in neglecting friction, additional dissipation in the numerical model for realistic, jagged coastlines, and the fact that the solution has not completely reached the steady state. It is interesting to note that there is a current of 3 Sv going through the Drake Passage. From (10), it is readily seen that if the friction/damping coefficient is reduced 10 times, this recirculation will increase 10 times because the amplitude of the perturbation is inversely proportional to r.

In the second case, the Indonesian Passage is open, the thermocline perturbation ratio is $A_A : A_I : A_P = 106 : 67 : 54$, Figure 2b. The effective southern tip of the boundary between the Indian and Pacific is moved up to 9.1°S, still far enough from the equator so that the boundary current is still controlled by the geostrophic relation. The deepwater flow takes a shortcut through the Indonesian Passage. This direct path gives rise to a slightly larger thermocline depth perturbation in the Pacific. The mass flux through the Indonesian Passage is 3.3 Sv. This deepwater flux is inversely proportional to the friction parameter r; thus, if r is reduced 10 times, this deepwater flux should increase 10 times.

In the third case both the Indonesian and Drake Passages are closed; the thermocline perturbation ratio is $A_A : A_I : A_P = 98 : 81 : 43$, Figure 2c. With the Drake

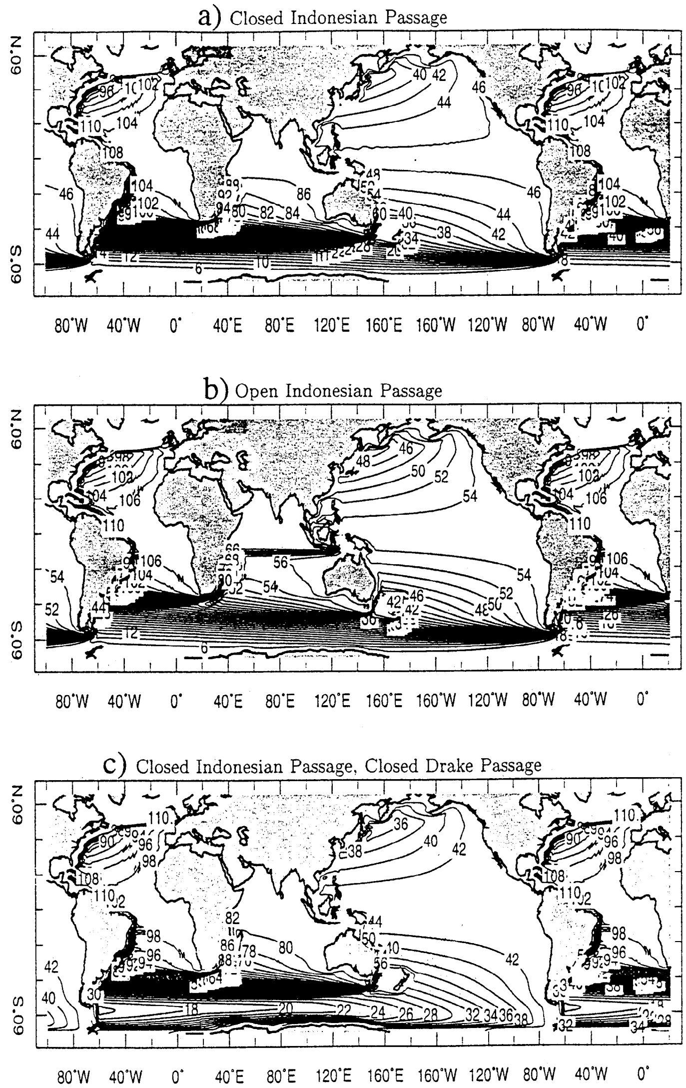

Figure 2. The displacement of the thermocline (m) induced by NADW source of 10 Sv.

Passage closed, equations (8) and (9) need modification: the last term on the right-hand side of (9) is eliminated and the area of the Pacific basin is expanded to the Antarctic coast. It is interesting that in this simple model the thermocline adjustment in the global oceans is not sensitive to whether the Drake Passage is open or closed. However, because the Drake Passage plays such a vitally important role in controlling the thermohaline and thermocline circulation in the world oceans, the meaning of this result should be interpreted with caution.

To test the sensitivity of the model to the friction/damping parameter r, we carried out several numerical experiments. As suggested above, decreasing r from its already small value will increase the thermocline displacement everywhere approximately inversely with r. If r is increased by a factor of 5 and the Indonesian Passage is closed, the depth perturbation ratio is $A_A : A_I : A_P = 49 : 25 : 4.5$. Thermocline perturbations are now mostly confined to the Atlantic Ocean, with much smaller amplitude in the Indian, and extremely low amplitude in the Pacific. For such a large r, the friction terms dominate the right-hand sides of (8, 9). In each basin, the layer depth perturbation along the equatorial boundary, the western boundary, and the eastern boundary is larger than in the ocean interior. As r is further increased, layer depth perturbations will be further confined to these boundary layers where Kelvin waves carry the signal through the world oceans.

Both Kelvin waves and Rossby waves establish the steady solutions discussed above. When NADW is formed, Kelvin waves start to carry the deep water to the world ocean via the deep western boundary current. Kelvin waves travel very fast: 3 m/s for the first baroclinic mode. Within a month a signal will reach the western equatorial Atlantic. From here the Kelvin waves travel along the equator. Reaching the eastern boundary, the signal propagates poleward along the coast into both hemispheres, generating Rossby waves that propagate westward, carrying the signal to the ocean interior.

The southern branch of the coastal Kelvin waves reaches the southern tip of the African continent and turns eastward into the Indian Ocean, where it travels toward the equator. The first signal will reach the equatorial Indian Ocean within a year, but it takes about 5 years for the amplitude there to become appreciable. From there it takes another 5 years to generate a significant signal in the equatorial Pacific, though the first sign of change will reach the equatorial Pacific within a year.

The eastern coast of South America consists of three regimes, Figure 1, controlled by Rossby waves generated along the southern coast of Africa (S_A), along the coast of Australia (S_I, the shaded area), and along the western coast of South America (S_P). Given the much longer arrival times of the Rossby waves at high latitudes and over longer distances, the response of the thermocline at these locations to changes in deepwater formation in the North Atlantic is much delayed.

4. Discussion

In contrast to an explanation relying on advection in the deep ocean, our model explains how a signal may be transmitted globally from the North Atlantic in times short compared to the O(1500-year) time scale of the Dansgaard–Oescher cycles. The signal is transmitted rapidly from basin to basin by coastal and equatorial Kelvin waves, and is carried into the interior by Rossby waves. In our model the amplitude is substantial: a 10-Sv change in the rate of NADW formation gave a 100-m thermocline depth change in the Atlantic, and a 50-m change in the Pacific. Note that a slowdown or shutoff of NADW formation, which will cause the North Atlantic SSTs to cool, will deepen the thermocline elsewhere, which tends to make SSTs warm. This is not consistent with observations, e.g., for the Younger Dryas, when cooling was prevalent globally. Note that the GCM response to cooling in the North Atlantic typically shows a mild warming over much of the globe in response to a shutdown of NADW, e.g., *Manabe and Stouffer* [1988].

While the adjustment mechanism we describe surely operates in the real ocean, our model is very simple and inferences for the real ocean must be drawn cautiously. Our results apply to a two-layer ocean, or, equivalently, to the first baroclinic mode. The low friction limit would not be expected to hold for higher baroclinic modes. Thus the signal in higher modes will be strongly diminished outside the Atlantic. To take the results in Figure 2 literally is to put the signal all into the first baroclinic mode; this surely overstates the change in thermocline depth. If the signal is carried out of the North Atlantic by perturbations with the scale of the mean deepwater flows, then the response to a change in NADW involves many modes and the distant response would be far weaker than suggested by our first baroclinic mode results. Results from the GCM do show a weak global signal.

Perhaps the most unrealistic feature of our model is its treatment of the Southern Ocean. As noted above, it made little difference whether the Drake Passage was open or closed, a finding in disagreement with a substantial literature. In our flat-bottomed model with no Scotia Arc there is no way to transmit a signal across the latitudes of the Drake Passage. There is no baroclinic eddy or other frictional mechanism to move properties across the Antarctic Circumpolar Current. Perhaps most important in view of the literature on the subject, there is no interaction between North Atlantic Deep Water and water formation in the Southern Ocean. It remains for future work to consider the rapid global adjustment mechanism discussed here in a more realistic context.

Acknowledgments. This study was carried out during RXH's visit to Lamont. RXH is supported through a grant from NSF OCE-9616950 (WHOI Contribution Number 9968), MAC and NK are supported through a grant from NSF OCE-9819538.

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(Received June 25, 1999; revised September 14, 1999; accepted January 14, 2000.)