THE OCEAN'S OVERTURNING CIRCULATION

Each second millions of cubic meters of water circulate between the ocean's surface and its depths. Despite the crucial role this 'overturning' may play in regulating Earth's climate, we still do not know how vulnerable it is to anthropogenic environmental changes.

J. Robert Toggweiler

The ocean is a mysterious place. As terrestrial creatures, we are familiar with a small volume of relatively warm ocean water near the surface. The deep ocean is not so familiar. Isolated from the thermal energy of the Sun and the mechanical energy of atmospheric winds, the deep ocean is completely dark and filled with near-freezing water. Yet the deep ocean is far from stagnant. Radioactive tracer measurements show that the ocean's circulation brings its interior water into contact with the atmosphere every 600 years or so.

The circulation that brings cold water into the interior takes the form of an "overturning" in which cold, dense water sinks near the poles and is replaced by warmer water flowing poleward from low latitudes. In the relatively small areas in the North Atlantic and around Antarctica where new deep water forms, the ocean gives up large amounts of heat to the atmosphere—on the order of 50 W/m^2, an amount comparable to the solar energy reaching the lowest layers of the atmosphere in those regions, especially during the winter months. This localization of the ocean's heat loss has a major impact on temperate- and high-latitude climates, as indicated by evidence in the geologic record pointing to abrupt changes in the ocean's overturning as a possible cause for large swings in high-latitude climates, such as the ice ages.[1]

Oceanographers know a great deal about the paths by which the cold, dense water enters the deep ocean. They routinely monitor the strength of the flow in regions close to the sources of new deep water. However, the completion of the overturning cycle—the return of old, deep water to the surface—is very difficult to measure directly and more difficult to explain, as is the vigor of the overturning circulation.

Computer models of the circulation built by oceanographers and climatologists give a wide range of circulatory solutions consistent with our knowledge of the forces acting on the system. We still do not know which of these circulation models is correct.

The consequences of our ignorance could be enormous. Changes in the ocean's overturning are one of the most important feedbacks in greenhouse warming scenarios. Using a coupled ocean–atmosphere general circulation model, Syukuro Manabe and Ronald Stouffer[2] showed that after a fourfold increase in atmospheric CO_2, the ocean's overturning slowly dies out within 200 years. A total collapse of the overturning could drastically alter the Earth's carbon cycle, possibly even causing the deep ocean to outgas additional CO_2 to the atmosphere. It also would disrupt the cycling of nutrients to the ocean's biota. While coupled global circulation models are still too young to permit us to view this result as a prediction, "wake-up calls" like this alert us to the kinds of changes we might have in store and have stimulated new interest in improving global circulation models and in conducting experiments that constrain these models. (See figure 1.)

A full understanding of the ocean's overturning may require some sort of hybridization of two theoretical frameworks. One framework contends that the overturning is driven by buoyancy forces, that is, the creation of dense and not-so-dense water masses by the addition or removal of heat and fresh water. The other sees the overturning as a largely mechanical process driven by the wind stress in the region of the Antarctic circumpolar current.

Formation of new deep water

Two main varieties of deep water, distinguishable by their salinities and temperatures, enter the deep ocean. The type from the North Atlantic, called the North Atlantic deep water, is warmer and saltier (2.5 °C, 35 grams of salt per kilogram of seawater). Deep water forming in the Weddell Sea of the Antarctic is fresher, colder and denser (−1.0 °C, 34.6 grams of salt per kilogram of seawater), and so flows along the ocean's floor. It is called, appropriately, Weddell Sea bottom water. (See figure 2.) More than half the ocean's volume is filled with cold water from these two sources. Deep water forms and flows away from its sources at rates on the order of a million cubic meters per second. The deep boundary current transporting North Atlantic deep water southward along the coasts of North and South America carries[3] about 15×10^6 m^3/sec, equivalent to 15 times the flow of all the world's rivers. The bottom current flowing out of the Weddell Sea transports[4] $2–3 \times 10^6$ m^3/sec.

One can gain some idea of the rates of production and flow of deep water from figure 3, which shows the north–south distributions of tritium (^3H) in the deep North Atlantic found during surveys in 1972 and 1981, 10 and 20 years after atmospheric nuclear weapons tests dumped large amounts of tritium into the northern North Atlantic and Arctic Oceans.[5] New deep water sinking north of Iceland (65° N) and in the Labrador Sea (60° N) has carried bomb-test tritium into the ocean's interior. Within

J. Robert Toggweiler is on the research staff at National Oceanic and Atmospheric Administration's Geophysical Fluid Dynamics Laboratory, in Princeton, New Jersey.

Experiment to measure vertical mixing of ocean waters was conducted by injecting SF_6 tracer into the Atlantic along a constant-density surface about 310 m deep. After six months the tracer had spread laterally over hundreds of kilometers, but vertically only a few tens of meters—much less than would be necessary to account for the vigor of the ocean's overturning. In this picture the SF_6-injection sled is being retrieved aboard the R/V Oceanus, operated by Woods Hole Oceanographic Institution. (Courtesy of James Ledwell, Woods Hole.) **Figure 1**

20 years of the bomb tests virtually all the deep water north of 40° N in the North Atlantic had measurable levels of tritium. Penetration of bomb tracers into the southern ocean was much more limited.[6]

Salt content and temperature together determine the density of seawater. At low temperatures a 1-part-per-thousand increase in salinity increases water's density as much as would a 7 °C temperature decrease. Salinity plays a very important role in determining where new deep waters form.

The low and middle latitudes of the ocean typically lose water vapor to the atmosphere (evaporation exceeds precipitation), while the high latitudes gain fresh water (precipitation exceeds evaporation). The excess precipitation in high latitudes tends to make surface waters less dense, thereby countering the effect of high-latitude cooling. No deep water forms today in the North Pacific, primarily because North Pacific surface waters are too fresh even at the freezing point to sink below a few hundred meters.

In the North Atlantic, by contrast, surface salinities remain very high even in regions of net precipitation. Wallace S. Broecker and George H. Denton attribute this increased Atlantic salinity to a net export of water vapor from the Atlantic and its drainage basin. The largest outflow occurs by means of moist, low-level winds blowing across Central America into the Pacific. As a result the North Atlantic is the ocean's largest source of new deep water.

Many researchers think that the formation of North Atlantic deep water is especially vulnerable to changes in high-latitude precipitation and runoff of the type predicted by most atmospheric global circulation models of our greenhouse future. The lowering of Atlantic salinities by excess precipitation is what eventually kills the overturning in Manabe and Stouffer's coupled ocean–atmosphere model.

Removal of old deep water

According to a long-standing theory, old deep water is removed from the ocean's interior by widespread upwelling into the ocean's thermocline—a low- and midlatitude feature that divides the warm shallow layers of the ocean from its cold deep layers. (See figure 2.) The theory contends that upwelling occurs because a downward mixing of heat across the thermocline decreases the density of old deep waters and allows them to be displaced upward by new deep waters flowing toward the equator from the poles. Enough heat must be added that the upper ocean can absorb the old deep waters. Once absorbed by the upper ocean, water can flow poleward to complete an overturning cycle.[7]

Given the average age of the ocean's deep water, it is fairly straightforward to determine the amount of mixing necessary to remove all the ocean's deep water. If the mixing process is described as simple diffusion, one needs a level of mixing commensurate with a diffusion coefficient of about 1 cm^2/sec. Because vertical mixing in a stratified ocean works to raise the ocean's center of mass, a substantial source of turbulent energy is needed to sustain a global mixing rate of that size. Oceanographers have been trying for over 20 years to determine whether vertical mixing rates in the ocean can actually be so large.

Detailed measurements of the ocean's vertical temperature gradient suggest that mixing rates are much smaller. Such measurements routinely encounter a microstructure caused by turbulent mixing. The temperature variance associated with the microstructure is thought to be a direct reflection of vertical mixing. If vertical mixing rates were as large as 1 cm^2/sec, the temperature variance should be much larger than observed.[8] The best measurements to date of mixing rates were performed by James Ledwell of the Woods Hole Oceanographic Institution and his colleagues Andrew Watson and Cliff Law, who recently injected a small amount of a man-made tracer, SF_6, into a constant-density layer in the Atlantic thermocline and observed its vertical spread over time.[9] (See figure 1.) They found that the tracer patch thickened over six months from about 20 m to 55 m, equivalent to a mixing coefficient of only 0.11 cm^2/sec and consistent with the level of mixing inferred from microstructure measurements.

Assuming that the efforts of Ledwell and coworkers and others to assess vertical mixing rates are definitive in showing that mixing levels in the thermocline are too weak to sustain the type of overturning envisioned by the standard model, some new mechanism is needed to account for the removal of old deep water, one that is also consistent with the preponderance of deep-water formation in the North Atlantic over that in Antarctica. One candidate mechanism is the removal of old deep water by the wind-driven overturning that drives the Antarctic circumpolar current.

The Antarctic circumpolar current is a deep zonal current that flows eastward around the earth much like the idealized zonal current described in the box on page 49. At its southern extreme the current passes through a gap between the southern tip of South America (56° S) and the northern tip of Antarctica (62° S), known as Drake Passage, in which there are no continental obstructions around entire latitude circles. The Southern Hemisphere westerly winds drive a northward surface flow, with the maximum wind stress located about 6° of latitude north of Drake Passage.

The real ocean bottom, unlike that in the idealized circumpolar current, is not flat. The Antarctic current cell encounters a series of north–south ridges that traverse the region below depths of 1500–2000 m. As described in the box, no net poleward flow can occur above the ridges in the latitude band of Drake Passage, because there is no boundary against which the ocean can build a pressure gradient toward the west. Below the ridges, pressure-gradient forces remove the momentum put into the Antarctic current by the wind,[10] much like the frictional boundary layer in the idealized current.

Drake Passage is located in the region of the westerly wind band where water upwells from below to feed the diverging surface flow. Because net poleward flow above the ridges is prohibited, the upwelled water must come from below the ridges, that is, from the depths below 1500–2000 m, where old deep water resides.[11]

Because near-surface water in the vicinity of Antarctica cools to nearly the same temperature as water in the deep ocean, water south of the current is dense but weakly stratified. Very little mixing energy is required to overcome the stratification. Thus old deep water can be removed relatively easily if it is transported southward to the latitude band of Drake Passage.

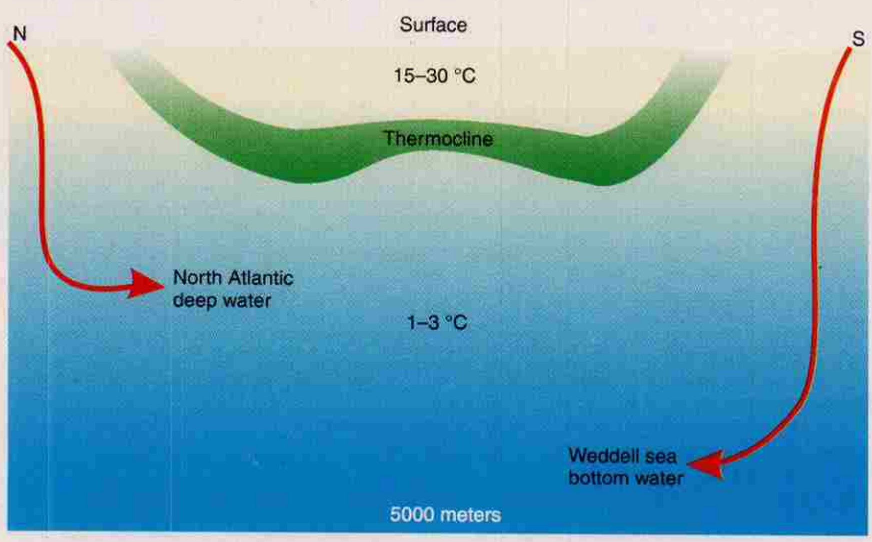
Where the deep water comes from is more difficult to understand. Even though the region north of Drake Passage corresponds to the area of convergence and downwelling shown in the box, the water column north of the Antarctic current is too strongly stratified for the relatively low-density, near-surface water to downwell to the depth of the ridges where it can flow poleward and feed the upwelling associated with Drake Passage. A global circulation model experiment I carried out with Bonnie Samuels shows how the ocean may solve this problem.[12] When we forced a global ocean model with stronger Southern Hemisphere westerlies, the model, somewhat paradoxically, produced more deep water in the North Atlantic. This result suggests that the extra downwelled water is advected all the way up to the North Atlantic, where it becomes dense, sinks in the usual way and then flows back to the south as a deep current. In other words, the ocean orchestrates a large-scale overturning across both hemispheres to accommodate the overturning near Drake Passage.

This assymetrical role of the Northern and Southern Hemispheres in the global overturning suggests that a marked asymmetry also should exist between the climates in the Northern and Southern Hemispheres. Because Drake Passage has no near-surface obstructions against which the ocean can build a pressure gradient, warm water north of the passage cannot flow directly down to Antarctica. Hence the atmosphere must transport the lion's share of the heat that reaches Antarctica. Because midlatitude atmospheric heat is transported mainly by storms, it stands to reason that the middle latitudes of the Southern Hemisphere are much stormier than they would be if Drake Passage did not exist. Stronger storms imply a stronger westerly wind stress on the ocean and, according to the arguments made here, more overturning in the North Atlantic. It is somewhat ironic that Drake Passage, by inhibiting the poleward transport of heat in the Southern Hemisphere, may greatly augment heat transport in the Northern Hemisphere.

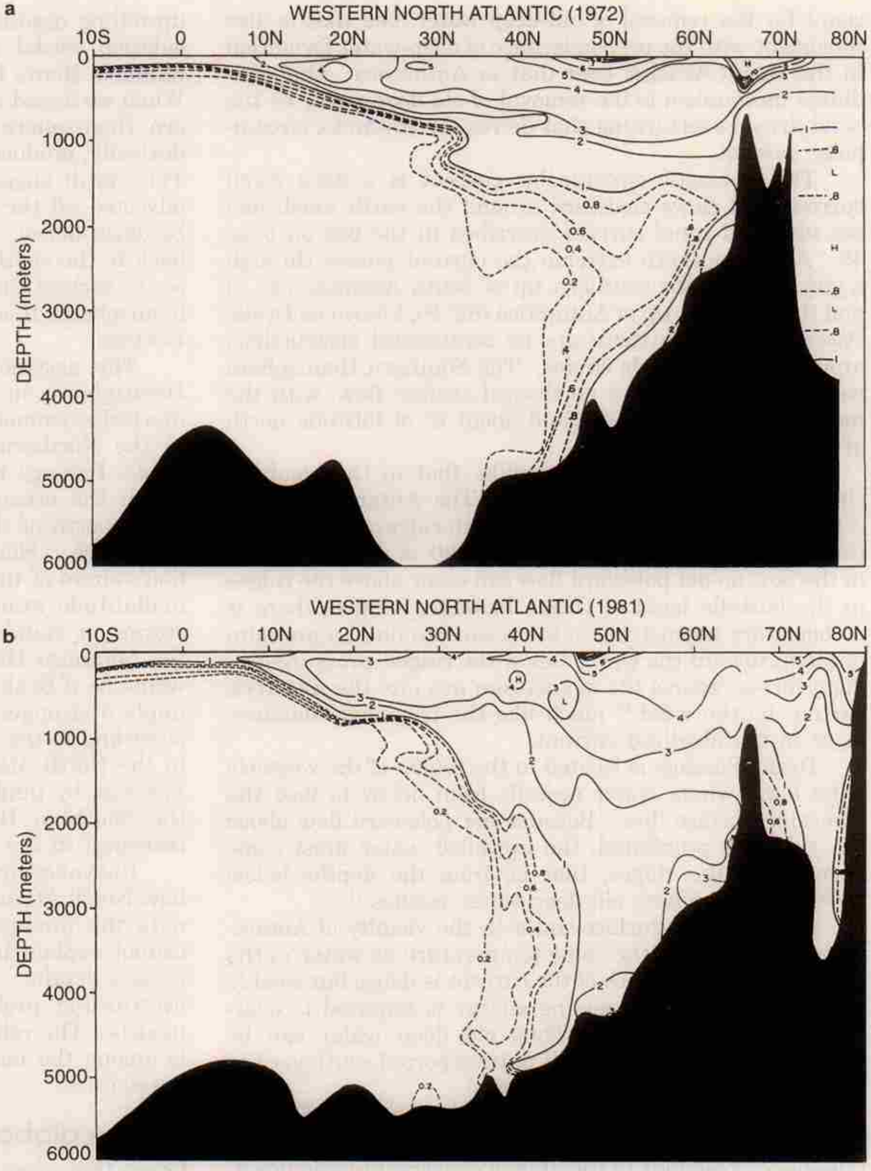
Buoyancy-driven models do a good job of explaining how North Atlantic waters remain saline enough to dominate the production of new deep water. Such models cannot explain how old deep water is removed from the ocean's depths. A complete understanding of the ocean's overturning probably requires a synthesis of the two models. The relative importance of the two mechanisms is among the many questions that future research must answer.

Ocean global circulation models

Given that most of the support for a mechanical model of the ocean's overturning comes from simulations done with global circulation models, it is important to understand how they work and what their limitations are. (See the article by William R. Holland and James C. McWilliams in



Ocean waters are stratified with respect to temperature and density. As depth increases, a relatively thin layer of warm water gives way to the thermocline, where temperatures decrease rapidly. Below the thermocline, the ocean has a fairly constant temperature (1–3 °C). New deep water enters the deep ocean primarily in the North Atlantic and around Antarctica, where relatively high salinities and low temperatures increase the water's density enough that it sinks into the denser bottom waters. **Figure 2**



Tritium contamination of North Atlantic surface waters by 1960s atmospheric nuclear tests provides a way of following the progress of the ocean's overturning in the North Atlantic. Tritium levels mark the southward progress of new deep water. **a:** Data taken in 1972 show that the ocean below 3000 m and south of about 43° N latitude was essentially free of tritium. **b:** Nine years later, regardless of depth, only water south of 35° N remained uncontaminated. The contours in the figure give the ratio of tritium to hydrogen in so-called tritium units. One tritium unit reflects a tritium to hydrogen ratio of 10^{-18} corrected to the activity levels that would have been observed on 1 January 1981. **Figure 3**

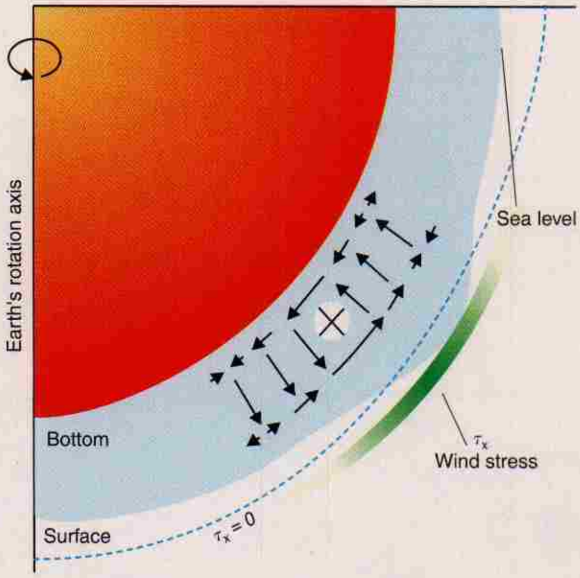
PHYSICS TODAY, October 1987, page 51.) Global circulation models determine a circulation by dividing the ocean into a three-dimensional grid of boxes and integrating an equation of motion forward in time for each box, subject to wind stresses imposed at the surface and the constraints imposed by an irregular ocean-floor topography. Separate equations, each subject to boundary conditions at the surface, govern the internal distribution of temperature, salinity and density. (In effect, the boundary equation in the North Atlantic incorporates the salinity mechanism of Broecker and Denton.) Horizontal variations in density produce an internal pressure field, which is added to the pressure field forced by the wind. Temperatures, salinities and velocities evolve in time subject to the total (internal plus external) pressure at every grid point.[13]

Unfortunately global circulation calculations consume large amounts of computer time, and this limits the fineness of the grid and therefore the precision of the calculations. Moreover, since many of the eddies responsible for mixing in the ocean take place on a scale significantly smaller than the grid size, global circulation models must introduce so-called diffusion or mixing terms—terms proportional to second derivatives of quantities like salinity and temperature—to simulate the mixing effects of eddies. In practice, however, the values of the diffusion terms are dictated by the need to keep the model numerically stable and are often larger than the values determined experimentally. As a result, it can be quite difficult to tell whether an effect observed in a global circulation model is real or an artifact of the diffusion terms or the coarseness of the grid.

One can perceive some of the difficulties of working with global circulation models by examining the overturning patterns (integrated around latitude circles) in figure 4. Figure 4a shows the global overturning produced by the global circulation model Samuels and I used[12] to illustrate the link between wind-driven overturning in the Antarctic current and deep-water formation in the North Atlantic. For this calculation the ocean was divided into 12 vertical levels and grid boxes 4° on a side. The model was integrated subject to steady annual-mean boundary conditions for 2000 years, during which time the model's interior temperatures and salinities evolve toward final states that are consistent with the model's dynamics and surface forcing. Figure 4b shows the ocean's meridional overturning as realized in a global circulation model in which the ocean was divided into 1° squares and 60 vertical layers and integrated for only two years; in this case, the model was initialized to observed temperature and salinity distributions. Within the first two years the overturning is strongly constrained by the ocean's observed density field.

In figure 4a the three tight, closed wind-driven overturning cells in the low to middle latitudes (between 60° N and 30° S) are quite shallow, penetrating only a few hundred meters below the surface. A fourth cell, associated with the Antarctic circumpolar current and southern hemisphere westerlies from 30° S and 60° S, extends much deeper into the interior. The large-scale overturning associated with the North Atlantic deep water dominates the interior between 1000 and 3000 meters and links up to some extent with the deep wind-driven cell in the south. This aspect of the model's overturning is consistent with

An Idealized Circumpolar Current



One can begin to understand water flow in the Antarctic circumpolar current by considering an idealized model of the current driven by midlatitude westerly winds that vary with latitude. We assume there are no continents and ignore density stratification of the water. The figure at right shows a cross section of the current looking east; darker shades of green correspond to higher wind velocities.

Forces acting on a differential volume of water in the figure include the pressure gradient $-\nabla P$, boundary forces $\boldsymbol{\tau}$ (wind stress at the surface and frictional stress along the bottom) and the Coriolis force, which is an effect of the Earth's rotation equal to $-2\rho\boldsymbol{\omega}\times\mathbf{v}$, where $\boldsymbol{\omega}$ is Earth's angular velocity and $\mathbf{v}$ is the velocity of the water. In the local Cartesian frame at a latitude θ (positive x is to the east, y is to the north, and z is radial), Newton's second law yields the following coupled equations for the x and y velocities of a unit volume of water:

$$ma_x = 0 = -2\rho\omega v_y \sin\theta + \partial P/\partial x - \partial\tau_x/\partial z$$

$$ma_y = 0 = 2\rho\omega v_x \sin\theta + \partial P/\partial y - \partial\tau_y/\partial z$$

At the surface, $\partial\tau_x/\partial z$ and the Coriolis terms dominate. The wind stress τ_x sets in motion the uppermost 20–50 meters of water. The Coriolis term deflects this water layer to the right in the Northern Hemisphere and to the left in the Southern Hemisphere until the x component of the Coriolis force exactly compensates the wind stress. Thus an initial east–west, or zonal, wind stress gives rise to a north–south, or meridional, surface current.

Because the wind stress varies with latitude, surface flow diverges south of the latitude of maximum wind stress and converges to the north, leading to a small north–south slope $\partial h/\partial y$ in the sea surface. The resulting higher sea levels in the north ($\Delta h \approx 40$ cm over the width of the westerly band) generate a pressure difference $\partial P/\partial y = \rho g\, \partial h/\partial y$ that, along with the Coriolis term, puts the entire column of water in motion toward the east. Such ocean currents, isolated from boundary layers, are said to be in geostrophic balance. Because geostrophic currents are essentially frictionless a relatively small pressure head at the surface acting over a vertical extent of several thousand meters can drive a current that transports 100×10^6 m^3/sec at a speed of 2 cm/sec. The convergence and divergence of water also drive vertical motion as shown in the figure. Without continents, $\partial P/\partial x = 0$, and there can be no net north–south flow in the interior.

Along the bottom of the channel, friction acts on the water in the x direction, generating a meridional flow that transports exactly the same volume of water as the layer at the top (but in the opposite direction) and removes the momentum added to the ocean by the wind.

the hypothesis that most of the upwelling of the deep water is associated with the Antarctic current. In one important respect, these results are at variance with the mechanically driven model. Several streamlines in figure 4a extend upward across 1000 m in low and middle latitudes, indicating a fairly strong upwelling into the thermocline. The total volume of this upwelling is about 15×10^6 m^3/sec, of which half occurs in the Atlantic and half in the Pacific. It is fairly easy to show that this upwelling is sustained by mixing of one form or another.

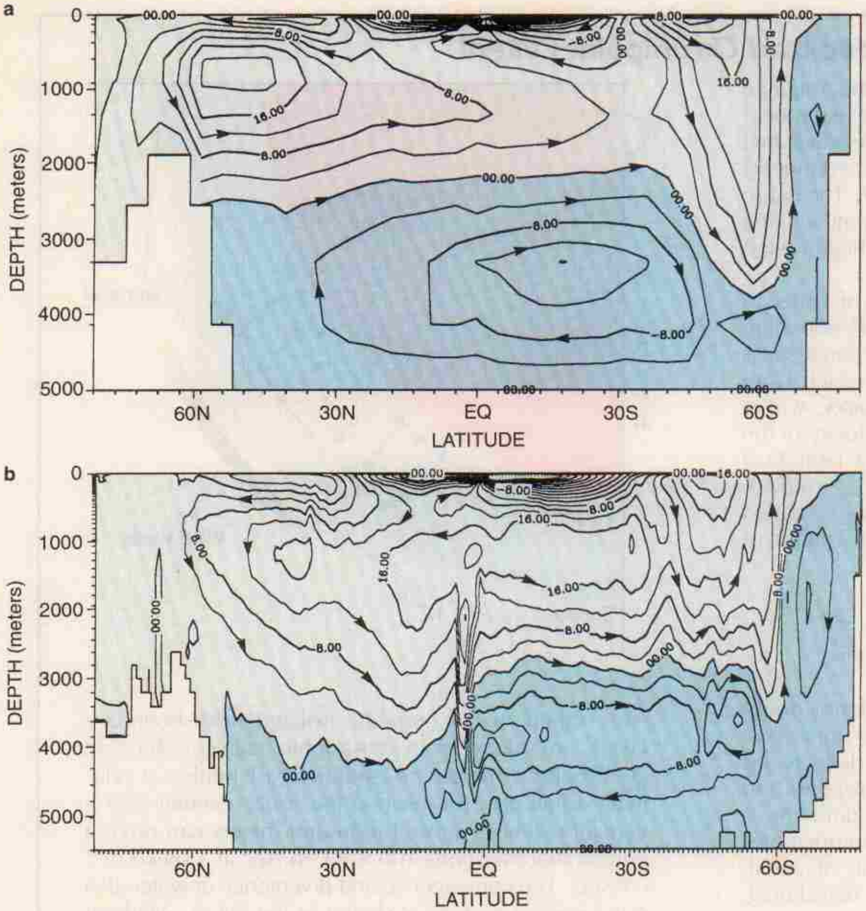
The global overturning of the 1° model in figure 4b shows essentially no net upwelling across 1000 m except in the far south (around 60° S). In the Southern Hemisphere the bottom-water cell and the North Atlantic cell combine to produce a total poleward flow of nearly 28×10^6 m^3/sec. Most of the streamlines tracking the 16×10^6 m^3/sec of water sinking in the North Atlantic can be traced directly to 60° S, where 16×10^6 m^3/sec of deep water upwells from below 2000 m in the latitude band of Drake Passage.

The difference between the overturning in figures 4a and 4b is striking. Water that upwelled across 1000 meters from the middle depths in the 4° model continues to flow south in the 1° model. The 1° model's internal pressure field is dictated by observed horizontal density variations, which maintain a strong southward flow.

According to radiocarbon measurements, the deep water between 1000 and 3500 m in the Indian and Pacific Oceans is the oldest deep water in the ocean. If one looks at flow only in the Indian and Pacific Ocean sectors of the 1° model, one finds no upwelling across 1000 m. Old deep water in the Indian and Pacific is removed from the interior of the ocean not by upwelling into the thermocline but by advection to the south.

The overturning in the 1° model looks very much like the description of the overturning in the mechanical model described above. The predicted overturning in the coarser-resolution model, while sharing some of the attributes of the 1° model's overturning, seems to be profoundly distorted by having so much deep water upwelling in low and middle latitudes.

Unfortunately the overturning in the 1° solution weakens considerably when the model is allowed to run beyond two years. Experience with lower-resolution models suggests that the overturning will begin to recover after 100 years or so, but only as model temperatures and salinities evolve away from the observed state and mixing effects begin to dominate the circulation. The big challenge for ocean modelers is to find internally consistent



General circulation models give strikingly different predictions for the ocean's overturning when integrated with different grid sizes and over different lengths of time. **a:** A model with 4°×4° grid boxes and 12 vertical divisions integrated over 2000 years reproduces the gross features of the ocean's overturning. However, spurious mixing effects lead to too much water upwelling in low and middle latitudes. **b:** When a 1°×1°, 60-vertical-level model is integrated for two years, no net upwelling across 1000 m is seen except at the latitude of Drake Passage. This result agrees with the mechanical model of upwelling. Contours in the diagrams are streamlines at intervals of 4×10^6 m^3/sec. **Figure 4**

solutions in which spurious mixing effects do not dominate the predicted overturning.

Consequences

The high-resolution calculation discussed above seems to support a model of the ocean's overturning in which the wind-driven overturning associated with the Antarctic circumpolar current and Drake Passage plays a very important role. In this model, the wind-driven upwelling in Drake Passage and the formation of new deep water in the North Atlantic are directly linked, despite the thousands of miles between them. This direct linkage has many important consequences.

Several studies with global circulation models have shown that buoyancy-driven overturning circulations are vulnerable to shutdown by relatively small increases in high-latitude precipitation. An overturning that includes a significant degree of mechanical forcing by Southern Hemisphere winds over Drake Passage may be less vulnerable to such changes. The studies by Manabe and Stouffer[2] show that while the overturning dies out as a result of a fourfold CO_2 increase, it survives a twofold increase. The ocean model used by Manabe and Stouffer is very similar to the low-resolution model used to generate figure 4a. I suspect that linkages between deep-water formation in the north and wind-driven overturning in the south contribute to the model's resilience. A model with less low- and midlatitude upwelling may be even more resilient. Answering exactly how much more resilient will require more detailed calculations with more realistic mixing effects.

The linkage also shows that the global overturning involves the transport of both deep and surface waters over thousands of miles. Under such circumstances, ocean–atmosphere interactions are too complex to be simulated by the simple boundary conditions used in ocean-only models. Only coupled ocean–atmosphere calculations can incorporate the subtle interactions that drive the overturning. Coupled models will be needed to understand exactly how the cycling of freshwater in one hemisphere and the action of winds in the other work together via the deep ocean to maintain the Earth's climate over long periods of time.

It is clear from the above discussion that much work remains to be done before we can say that we understand the ocean's overturning. Work done to date only serves to emphasize that the ocean and the atmosphere form a highly integrated system of energy transport, in which local changes can have global consequences.

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