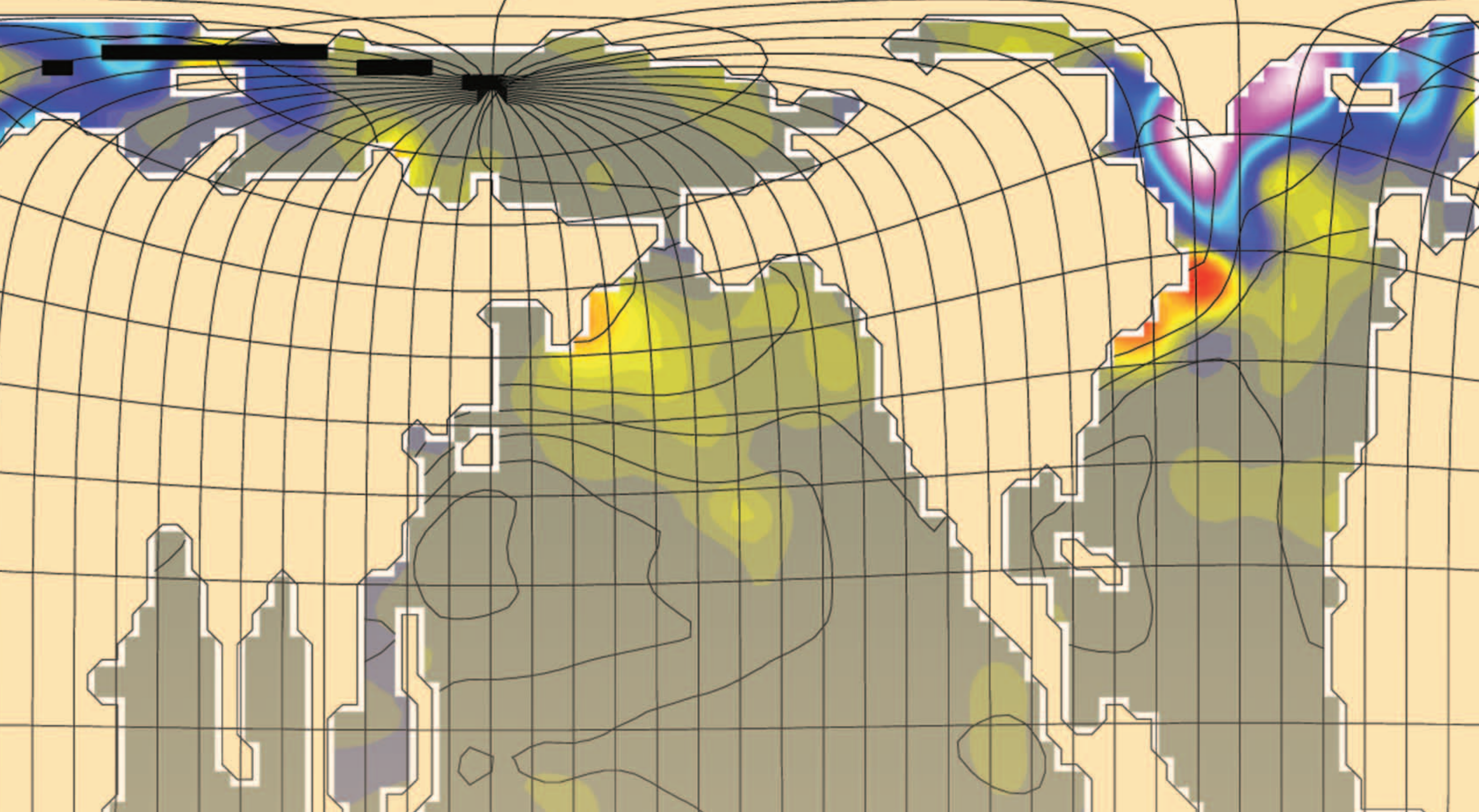


The Ocean Perspective

Uncertainties in Climate Prediction

Rainer Bleck

The ocean is but a thin coating on our planet. Ocean circulation, therefore, appears predominantly two-dimensional; however, ocean depth, the third dimension, cannot be neglected in ocean models. Surprising as it may be, the premier numerical challenge posed for ocean models used for climate prediction is keeping the warm poleward-flowing surface water thermally insulated from the cold abyssal return flow—as insulated as it is in nature. Los Alamos supports several approaches to ocean simulations, whose results give a hint about the uncertainties involved in climate prediction. The model designed to come closest to preserving the warm poleward and cold return flows of the ocean "conveyor" is the layer model, which pictures the ocean as a stack of immiscible layers. Compared with other models, the layer model also produces more stable oceanic circulation in the face of climate changes. Yet the jury is still out on whether "more stable" is the same as "more realistic."

Is it preposterous to predict Earth's climate 50 or 100 years ahead if we cannot reliably forecast the weather two or three days into the future? Fortunately, the situation is not as hopeless as one may think. There are fundamental differences between the two tasks.

Mathematicians classify weather prediction as an "initial value" problem because the accuracy of a weather forecast depends crucially on how well the initial state of the atmosphere is known. Climate prediction, on the other hand, is primarily a "boundary value" problem. In this case, the main task is to reproduce the time-averaged flow of solar energy through the nooks and crannies of the land-ocean-atmosphere system. To do so well, one needs to know those nooks and crannies, and one needs to know how much energy arrives at the top of the atmosphere as a function of latitude and time of year. But the exact locations of the transient disturbances that determine the oceanic and atmospheric "weather" need not be known, either initially or at a later time. In essence, when we predict future climates, we try to assess whether modifying certain parameters, such as the ellipticity of the earth's orbit or the chemical composition of the atmosphere, will change the way energy flows through the earth system. This task does not critically depend upon our ability to predict tomorrow's weather or the onset of the next El Niño—even though a forecast model that does well in these respects will increase our confidence in the correctness of the climate forecast.

Simulating systems that are as complex as Earth's climate is hard. Two types of errors may affect the simulation: errors in the physics of the model and errors in the mathematical approximations needed to simulate climate processes on a computer. Being able to distinguish between these two error types may help us develop more-accurate climate models. But to separate errors, scientists need tools, and model diversity is among the few available ones. In the realm of ocean modeling, Los Alamos has been supporting model diversity for over a decade. Several ocean-circulation models have been brought to or developed at the Laboratory, and they are designed to solve the same physical problem while being numerically dissimilar. By comparing their results, scientists get a feel for the size of the uncertainties. This article will use three examples related to Elf Niño, the heat-carrying ocean conveyor, and oceanic carbon sequestration to illustrate this approach.

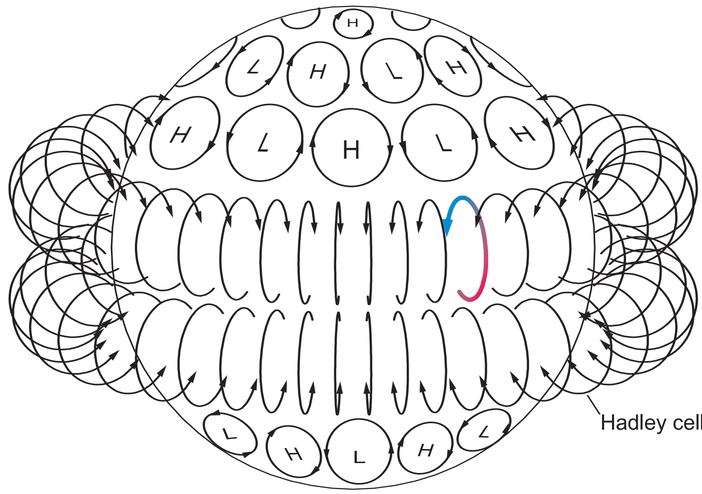


Figure 1. Heat Transport in the Atmosphere
This schematic view shows two atmospheric circulation modes important for poleward heat transport. A vertical-meridional overturning circulation (Hadley cell) dominates near the equator. Horizontally rotating eddies (the highs and lows on weather maps) dominate at mid to high latitudes.

Modes of Poleward Heat Transport

Our planet absorbs solar energy at low latitudes and radiates energy back into space at high latitudes. This is so because the earth is a sphere and its axis of rotation is more or less perpendicular to its orbital plane around the sun. For this system to remain in a steady state, heat on earth must continually flow poleward in both hemispheres. Transporting this heat is the job of the atmosphere and ocean because, in contrast to the solid earth, they can move heat efficiently by setting up warm currents flowing poleward and cold ones flowing back to the equator.

From here on things get complicated. The earth's rotation greatly inhibits meridional displacement of water or air because a northward- or southward-moving fluid parcel away from the equator also changes its distance from the earth's axis. In fact, the angular-momentum balance constraints resulting from the earth's rotation are so severe that the atmosphere can maintain a meridional overturning circulation (a closed loop consisting of air rising at low latitudes and sinking at high latitudes) only near the equator in the so-called Hadley cell (Figure 1). At mid-to-high latitudes, the earth's rotation forces the atmosphere to resort to a different mode of heat transport, namely, transient eddies, popularly known as highs and lows, which intermittently push warm air poleward and cold air equatorward over distances too small for the angu-

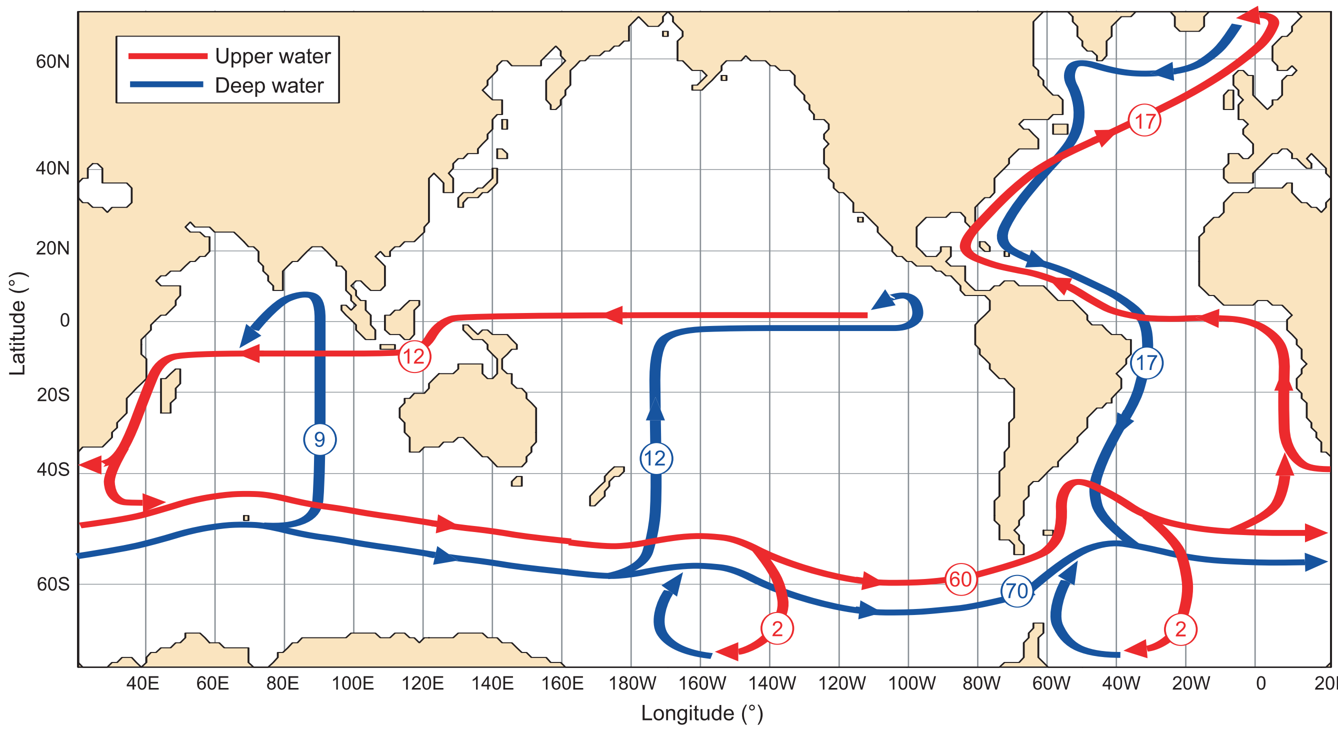


Figure 2. Heat Transport in the Ocean
The thermally forced ocean circulation spans ocean basins, as shown in this figure. Vertical and horizontal details are simplified but less so than in Broecker (1991). Wind-driven currents are omitted except for the Antarctic Circumpolar Current. Circled numbers represent transport in sverdrups (1 Sv = 106 m^3 s^{-1}, corresponding to roughly the volume transport of five Amazon Rivers). The schematic does not reflect the fact that downwelling takes place in geographically confined regions (Greenland/Norwegian Sea, Weddell Sea, and Ross Sea) while upwelling is a much more widespread process. Thus, not all the water entering the Indo-Pacific basins from the south up-wells in the specific locations indicated in the drawing. (Adapted from Sun and Bleck 2001a and Schmitz 1996).

lar momentum constraint to kick in.[1] The two modes of heat transport in the atmosphere and their respective geographic domains are depicted schematically in Figure 1.

In the ocean, in contrast to the atmosphere, steady meridional motion can be sustained over long distances when a current can "rub" against a continental margin and thereby shed momentum. This is why meridional ocean currents, such as the Gulf Stream, must always flow along the edge of an ocean basin, never in the middle. (Emphasis here is on the word "meridional." East-west currents can cross ocean basins in an unrestricted manner. Otherwise, the warm waters of the Gulf Stream would not be able to reach Europe.) Eddies, analogous to those in the atmosphere, do exist in the ocean, but their contribution to heat transport tends to be overshadowed by the contribution of the boundary currents. The Southern Ocean, being devoid of meridional land barriers, is the obvious exception; there, as in the atmosphere, ocean eddies play a primary role in heat transport.

The ability of the ocean to maintain steady meridional motion over considerable distances actually allows the ocean to develop two types of heat transport mechanisms not found in the atmosphere: a Hadley cell–like meridional overturning circulation extending all the way to the subpolar seas—dubbed the ocean conveyor (Broecker 1991)—and a basin-spanning horizontal gyrating motion. The former, depicted schematically in Figure 2, is primarily maintained by differential heating and cooling; the latter, by the torque exerted on the ocean by the prevailing pattern of tropical easterlies and extratropical westerlies.

[1] Even though extratropical eddies are as flat as pancakes, their flow field is not entirely two dimensional; in fact, they draw their energy from the rise/descent of warm/cold air masses. Their residual effect, if analyzed in a proper entropy-oriented framework, therefore, is to extend the Hadley cell to higher latitudes.

Implications for Ocean and Climate Modeling

To faithfully replicate the relevant heat-transport mechanisms on our planet, a climate model must be able to reproduce the action of atmospheric lows and highs without which there would be hardly any heat transport in the atmospheric submodel. In other words, the atmospheric submodel must be what ocean modelers refer to as "eddy resolving." In the oceanic submodel, on the other hand, the first order of business is to correctly simulate the major current systems, both those associated with the wind-driven horizontal gyre circulation and those associated with the thermally driven meridional overturning circulation.

This is not to say that the effect of ocean eddies can safely be neglected. Wherever they are in the ocean (and they are almost everywhere), eddies will transport some heat. However, in most oceans, except the Southern Ocean, the contribution of the eddies is overshadowed by the contribution of meridional current systems. As a result, the penalty for "parameterizing" the eddies' role, instead of explicitly resolving the eddies, is minor. Turning this argument around, one should expect the Southern Ocean to emerge as a major Achilles' heel in noneddy-resolving ocean modeling.

Eddy resolution in the ocean is a major problem. According to hydrodynamic instability theory, tailored to fluid motion on a rotating sphere, eddy size depends on the vertical density contrast in the fluid. Because this contrast is much smaller in the ocean than in the atmosphere, ocean eddies turn out to be roughly 10 times smaller in diameter (that is, 100 times smaller in area) than their atmospheric counterparts. Hence, the number of eddies to be tracked by an eddy-resolving ocean model through their individual life cycles exceeds by two orders of magnitude the number of eddies in a global weather model. Furthermore, in the context of climate, individual eddies would have to be simulated not only for the duration of a 5- or 10-day weather forecast, but also for decades or possibly centuries. This task is beyond the capabilities of even our biggest and fastest computers.

Contrary to common perception, the ocean is quite shallow, a thin coating on our planet, and oceanic circulation appears, therefore, predominantly two-dimensional. However, the presence of meridional overturning circulations and the concomitant reversal of current direction with depth mean that the third (vertical) dimension cannot totally be neglected when modeling the ocean. Surprising as it may sound, the premier numerical challenge posed by the third dimension in ocean models used for climate prediction is to keep the warm poleward-flowing surface water thermally insulated from the cold abyssal return flow—as insulated as it is in nature. Given the long time scales involved (decades to centuries) and the relative proximity of the two circulation branches (a few kilometers), this is indeed a major challenge. It has motivated the development of a class of ocean models that, instead of carrying ocean state variables on a rigid, crystal-like lattice, picture the ocean as a stack of immiscible layers whose thicknesses are allowed to evolve freely in space and time. By allowing grid cell interfaces (and the state variables riding on them) to bob up and down with the vertical component of motion, these so-called layer models control vertical mixing processes much better than models based on a rigid spatial grid. (The dispersive effect of an oscillating vertical motion field on such properties as temperature in a fixed-grid ocean model is illustrated in Figure 3). As a result, warm surface currents in a layer model are less likely to lose heat through contact with the cold return flow than those in a traditional fixed-grid ("level") model. In theory, at least, this difference translates into a more robust heat-delivery system and a more accurately simulated climate.

Potential density, defined as density corrected for compressibility effects, is a proxy for entropy in seawater and hence is conserved in the absence of heat-transferring, or diabatic, processes. Because oceanic flow below the surface layer generally comes close to being adiabatic, the layers in a layer ocean model are typically chosen to coincide with constant potential-density, or isopycnic, layers. The resulting impermeability of layer interfaces under adiabatic flow conditions allows vertical property exchange by diabatic mixing, to the extent that it occurs, to be modeled explicitly before a background of zero numerical mixing.

Replacing the traditional Eulerian vertical coordinate by a Lagrangian one, tied to the oceanic potential density field, sounds easier than it is. Given the small but persistent background mixing in the ocean, maintenance of a steady climate state requires that each parcel of seawater communicate with the atmosphere at least intermittently to replenish its temperature and salinity—the two ingredients that set the density of seawater. This is to say that each layer in a layer model must be allowed to "outcrop," or rise to the surface. Picturing the world ocean as a lens of light, warm water centered on the equator and floating on a body of dense, cold water, one readily sees that the densest layers outcrop closest to the poles, layers of intermediate density outcrop at mid-latitudes, and so forth (refer to Figure 4). To avoid having to deal with time-dependent lateral boundaries for individual coordinate layers, today's isopycnic models extend ocean layers, regardless of the actual extent of the water, over the

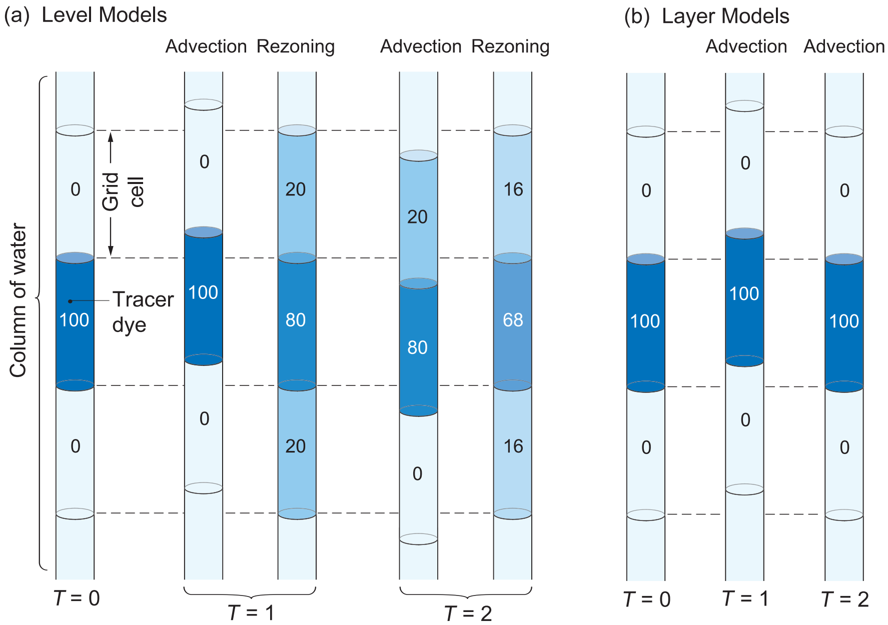


Figure 3. Anomalous Vertical Mixing in Fixed-Grid Models

(a) This schematic illustrates numerical dispersion in a water column, resulting from oscillatory vertical motion typically associated with passing gravity waves. Time increases from left to right. Shown is a vertical stack of three grid cells. The initial state, *T* = 0, is chosen to coincide with the wave trough, at which time the center grid cell is assumed to be filled with a tracer of concentration 100. In the advection step at *T* = 1, the approaching wave crest causes the water in all three cells to rise by a distance corresponding to one-fifth of the vertical cell size. The clock is stopped momentarily to allow the tracer to be reapportioned, or rezoned, among the original grid cells, which in contrast to the water column, stay fixed in a level ocean model. Because of rezoning, the tracer is split between two cells (*T* = 1). Next, the clock is running again. The approaching next wave trough causes the water column to return to its initial position during advection, at *T* = 2. With the clock stopped again, the tracer is being rezoned a second time. Tracer concentration in the center cell is now down to 68, with the remainder spread over the two adjacent cells. Note that this is an extreme example. Dispersion can be reduced by use of more sophisticated rezoning schemes. Also, gravity waves, while ubiquitous, usually have smaller amplitude than assumed here. (b) It is important to note that layer models skip the rezoning steps and thereby maintain a concentration of 100 in the center cell.

whole model domain as empty or massless layers. All these conditions translate into tricky numerical issues, making layer models inherently more complex than traditional level models.

Because of these tradeoffs, neither model class can be regarded as superior in every respect in simulating the global ocean circulation. However, two models that start from the same physics—including the "closure" model that approximates the effect of turbulent exchange processes at the small, unresolved scales—but express that physics in different mathematical form provide important insight into the inevitable degradation inherent in solving differential equations by computer. This comparative approach, therefore, affords some measure of the overall uncertainties in climate prediction.

It is important to note that the two ocean-model classes differ not only in their numerical representation of a given set of differential equations but also in the differential equations themselves. This begs the question, "how can there be two sets of equations for a single, uniquely defined physical problem?" The answer is that the underlying physical principles (Newton's law, conservation of mass, and others) can be cast in different forms, depending on which variables in the set consisting of depth, temperature, salinity, density, and velocity are treated as dependent variables. In level models, depth is an independent variable, whereas water density is a dependent variable, stepped forward in time as one solves prognostic equations for temperature and salinity. The equations governing layer models, on the other hand, treat density as an independent variable and, in the spirit of maintaining consistency between the number of unknowns and equations, they treat depth (in the form of layer thickness) as a dependent variable. It is this switch, rather than variations in the way differential equations are translated into algebraic ones, that gives different properties to the solutions obtained from level and layer models.

Los Alamos Contributions

In the early 1990s, the Department of Energy (DOE) Office of Science joined other federal agencies in funding the development of layer ocean models for climate prediction. The main reason was the perceived need to enrich the ocean model "gene pool," which at that time was rather sparse and showed signs of model inbreeding. Today, both level and layer models are firmly established at Los Alamos. The level model class is represented by the Los Alamos–developed Parallel Ocean Program (POP). For a detailed account of ocean-modeling advances achieved through development of POP, refer to Malone et al. (1993, 2003). The layer model class is represented by the Miami Isopycnic Coordinate Ocean Model (MICOM) by Bleck et al. (1992) and its hybrid-coordinate offshoot HYCOM by Bleck (2002).

Hybrid-coordinate models are designed to combine the advantages of layer and level models. Starting at the surface, one assigns progressively larger "target" potential-density values to coordinate layers in hybrid models. Each coordinate layer is expected to track its assigned isopycnic layer in the model domain in space and time but may deviate from it to form a conventional constant-depth layer if (and only if) the target density is too low to exist in a given water column. Layers assigned to relatively warm, or low-density, water, which in traditional isopycnic models would only exist at low latitudes, thereby are allowed to molt into constant-depth layers poleward of their outcrop latitude. These redefined layers provide a framework for solving the model equations in subpolar oceans, where the lack of vertical density contrast makes it hard to represent vertical structure in terms of density classes.

Judging from the willingness of such federal agencies as the Naval Research Laboratory and the National Weather Service to adopt HYCOM (see, for example, http://www7320.nrlssc.navy.mil/ATLhycom1-12/skill.html), the hybrid model concept is widely being regarded as a significant step toward creating a flexible, multipurpose, next-generation ocean model. The COSIM (for Climate, Ocean, and Sea Ice Modeling) group at Los Alamos is under contract with the DOE Office of Science to produce a hybrid-coordinate version of POP as well.

The algorithm in HYCOM that determines whether a given coordinate layer can retain its isopycnic character at a given location or whether it must be assigned a constant thickness and be "frozen" in space has elements in common with the Los Alamos–developed arbitrary Lagrangian-Eulerian (ALE) technique (Hirt et al. 1974). However, whereas traditional ALE applications focus on maintaining a nonzero mesh size, the HYCOM algorithm addresses the more vexing problem of moving coordinate layers through the fluid to realign them with their respective target isopycnals after they have become separated. An illustration of how hybrid-coordinate models work in practice is given in Figure 5.

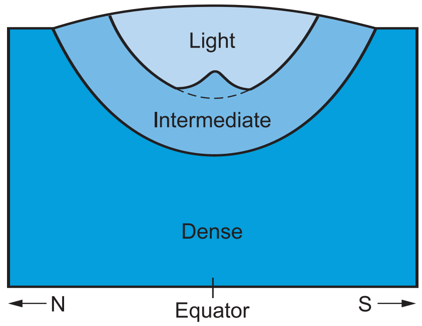


Figure 4. The Ocean as a Lens of Light Water Floating on Dense Water
As their densities increase, ocean layers outcrop progressively closer to the poles. Coordinate layers in MICOM follow the same general pattern.

Examples of Multimodel Climate Sensitivity Experiments

The vagaries of weather forecasts are the butt of jokes. Yet the meteorological community has rather precise information about the "skill" of numerical models used in daily forecasting and about the associated uncertainties. This information is precise because weather models are intended to duplicate the behavior of a readily observable system and because gathering statistical information about model skill is made easy by the large and ever-growing ensemble size.

The situation is quite different in decadal to centennial climate prediction because of the lack of verification data, the sheer number of natural processes contributing to the steadiness of climate (or its change, as the case may be), and the need to either treat in cursory fashion (parameterize) or totally omit from the model those processes that are deemed less central to the climate problem than others. Uncertainty quantification in long-range climate prediction, therefore, is a science that arguably is not even in its infancy.

Not much needs to be said about the lack of verification data. Important climate-relevant aspects of the earth system, such as atmospheric greenhouse-gas concentrations and the oceanic abyssal circulation, have been observed only in the last half century in sufficient detail to validate three-dimensional climate models. This observational record is vitally important as it provides a glimpse at the performance strengths and limitations of today's climate models, but it cannot serve as a database for rigorously assessing model skill. Stated differently, the 50-year observational record allows us to check the appropriateness of certain parameterizations (also referred to as physical closure assumptions) in climate models, but as an "ensemble" of one, it is insufficient for quantitatively evaluating prediction uncertainty.

At present, the focus in the climate research community is on the number (and ranking) of climate-contributing natural processes and on the need to parameterize. One can argue that, given the complexity of the climate problem and the finite nature of computing resources, there is not a single process that is not, in one way or another, parameterized in a climate model. The omission of possibly relevant detail begins with the transfor-

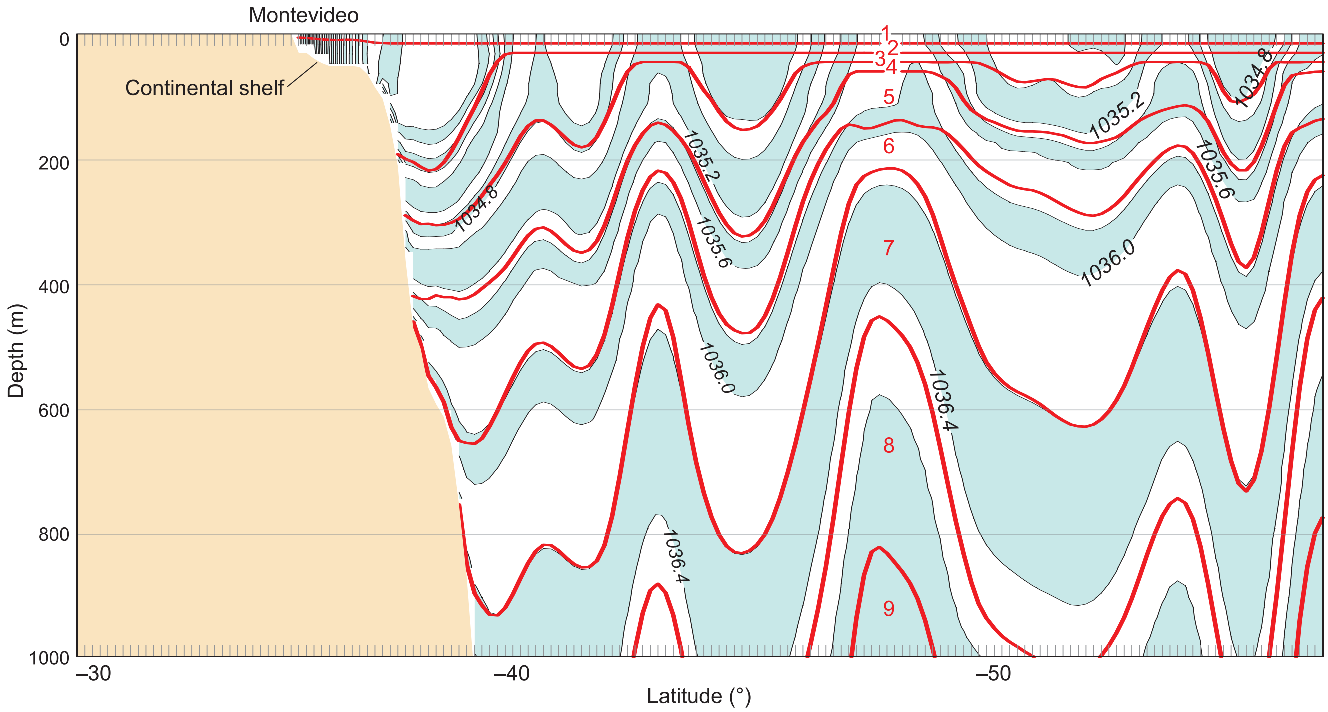


Figure 5. Ocean Density and Hybrid Coordinates Near the Falkland Islands

This is a sample vertical section through a HYCOM solution extending south from Montevideo into the eddy-rich confluence region of the Brazil and Falkland currents. The red numbers from 1 to 9 are the hybrid layers. The South American continent is shown at left. The latitude (°) is marked as negative numbers along the bottom. The heavy red lines represent HYCOM's coordinate surfaces; the shaded contours, outlined by light black lines, represent potential density in kilograms per cubic meter. Tick marks along the abscissa indicate grid resolution (approximately 15 km). The ordinate shows depth in meters. Note that coordinate surfaces follow isopycnals at depth but turn horizontal near the surface whereas the associated isopycnals outcrop. Density undulations indicate the presence of "cold-core" and "warm-core" eddies (which in the southern hemisphere spin clockwise and counterclockwise, respectively). Crowded isopycnals on the continental shelf indicate the presence of low-salinity Rio de la Plata water.

mation of the differential equations that govern the behavior of the natural system into computer-solvable algebraic equations. The truncation of the spectrum of scales at a chosen mesh size immediately divides processes into spatially resolved and unresolved ones, the latter requiring a physical closure assumption. A good example of a closure scheme for processes taking place on spatial scales too small to be resolved by a climate model is the wind-induced turbulent mixing below the sea surface. Since this turbulence stirs up water from depths of tens or even hundreds of meters, it strongly affects sea surface temperature. Disregarding or poorly parameterizing it, therefore, has dire consequences on the representation of air-sea exchange processes in our models.

Errors associated with the inevitably imperfect physical closure of unresolved processes are compounded by errors introduced by solving algebraic instead of differential equations; these so-called truncation or discretization errors mainly affect the resolved scales. Hence, climate forecasts are fraught with a mixture of physical closure errors and numerical truncation errors.

Notwithstanding efforts by groups such as the Program for Climate Model Diagnosis and Intercomparison (PCMDI) at Lawrence Livermore National Laboratory (http://www.pcmdi.llnl.gov), the climate community is still largely unable to separate the effects of physical and numerical errors on a climate forecast. One of the few tools at our disposal, as already mentioned, is developing multiple climate models that employ identical physical-closure schemes but are based on different numerics. This approach leads to the need for what was earlier referred to as genetic diversity in climate models. The differences between level and layer models arguably provide such diversity and hence open the door to experimentation aimed at separating physical from numerical model errors. A few examples of such experimentation are given below.

El Niño-like Variability in Climate Models. Much of the discussion about global warming focuses on the question of whether the currently observed global temperature rise can be attributed to the inherent natural variability of the ocean-atmosphere system or whether it is a consequence of increased greenhouse gas concentrations. In order to clarify this question through numerical simulation, one obviously needs a climate model with a proven ability to simulate the multitude of ocean-atmosphere feedback mechanisms giving rise to natural variability.

The biggest observed climate variability on interannual time scales is associated with the so-called El Niño–Southern Oscillation (ENSO) phenomenon. The ocean-atmosphere system in the tropics is known to switch back and forth between two states, one of which (La Niña) is characterized by strong trade winds and strong upwelling of cold subsurface water in the equatorial eastern Pacific, whereas the other (El Niño) is characterized by weak trade winds and weak upwelling. Both states appear to be self-sustaining in the sense that strong/weak upwelling caused by strong/weak trade winds tends to support the underlying wind anomaly. A particular signal telling the coupled system to initiate the switch from one state to the other has not yet been identified. Efforts to predict that switch, therefore, have not advanced beyond the stage of what may euphemistically be described as early detection.

The ENSO coupled mode is often used as a yardstick for how well a climate model handles internal variability. Most coupled models are actually capable of producing an ENSO-like variability mode (AchutaRao and Sperber 2002), but a fair amount of parameter tuning is usually required before those models come close to simulating the observed amplitude, frequency, and spatial anomaly pattern of the genuine ENSO. Tuning attempts usually focus on the turbulence closure scheme for the oceanic and atmospheric boundary layers, but changing the scale selectivity of the model by modifying the computational mesh can also have a surprisingly strong effect.

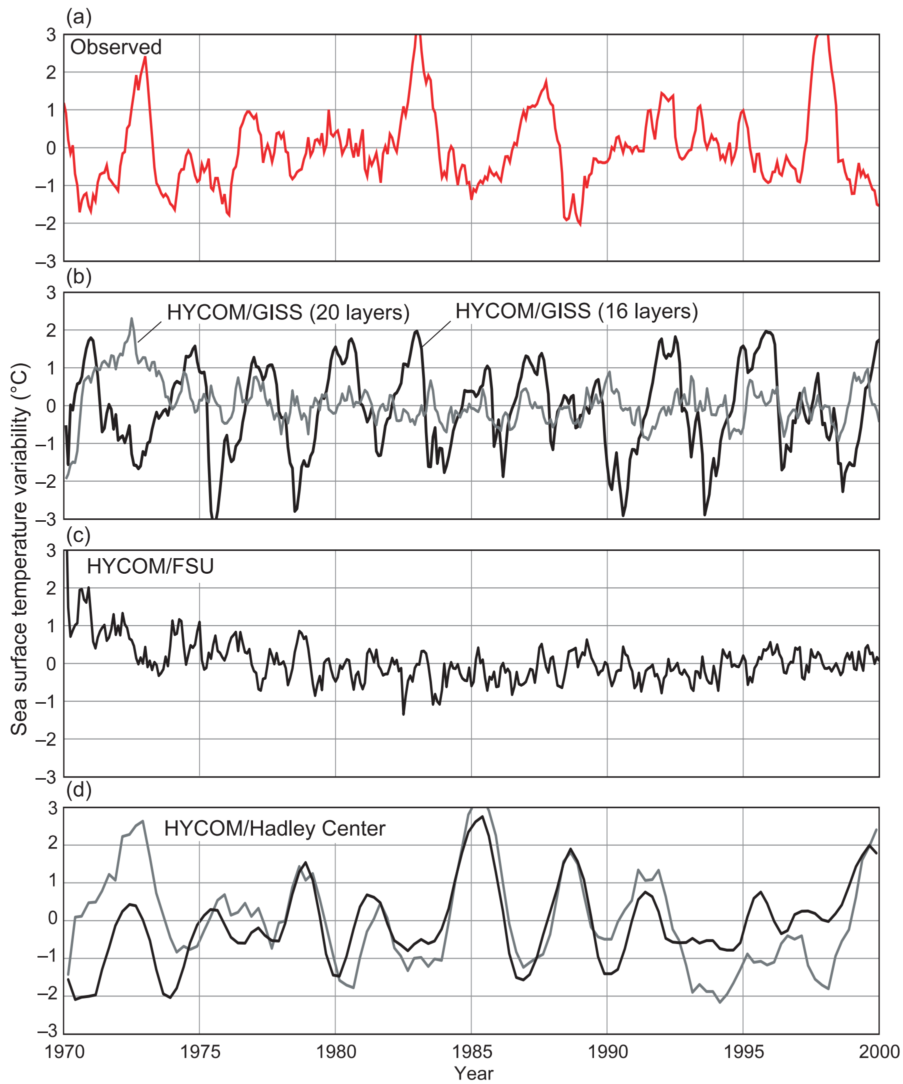


Figure 6. El Niño Variability in Climate Models

An observed 30-year time series of El Niño–related sea-surface temperature variability shown in (a) (Niño3 index, °C) is compared with corresponding time series obtained from three atmospheric circulation models, all of which have the oceanic component HYCOM in common: (b) model from the Goddard Institute for Space Studies (GISS) at NASA; (c) model from Florida State University (FSU); and (d) model from the Hadley Centre in the United Kingdom. Two curves within a panel indicate two runs based on different parameter choices: number of layers in (b) and different turbulence surface mixing in (d). The large model-to-model variation in Niño3 amplitude is largely unexplained and the subject of intense research.

(Graphs (b) and (d) are courtesy of Shan Sun from NASA/GISS and Alex Megann from the Southampton Oceanography Centre.)

The point just made is illustrated in Figure 6, in which an observed temperature time series from the equatorial Pacific highlighting ENSO variability is compared with corresponding time series obtained from three climate models that have the oceanic component HYCOM in com-

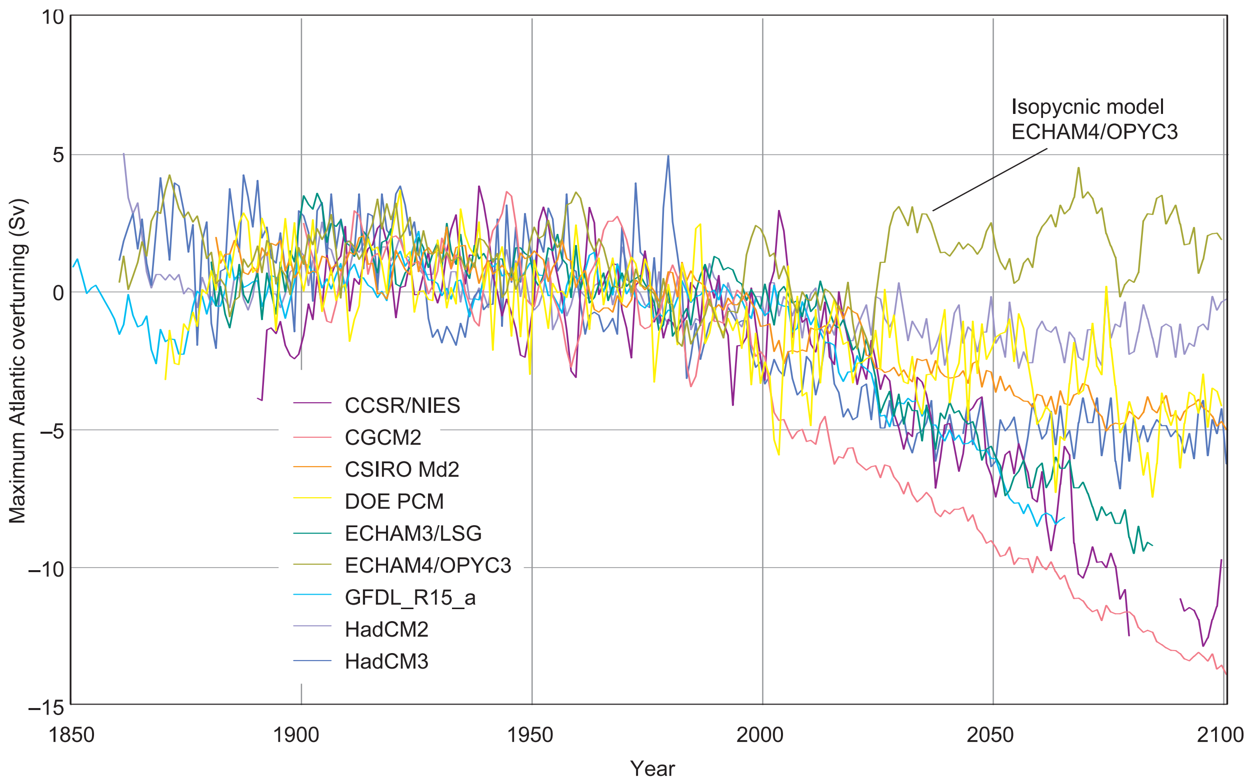


Figure 7. Effects of Global Warming in the Atlantic Overturning Rate

The curves in this plot represent changes in the Atlantic overturning rate (1 Sv = 106 m^3 s^{-1}) from the gradual doubling of atmospheric CO_2 in nine coupled climate models. Overturning rates are plotted relative to each model's average over the period from 1960 to 1990. A reduction by 15 to 20 Sv amounts to a total shutdown of the overturning. Whereas the eight level models show a decreasing overturning rate, the isopycnic, or layer, model ECHAM4/OPYC3 does not indicate a slowdown of that rate under the conditions described above. **(Reproduced courtesy of IPCC 2001.)**

mon. As expected, the amplitude of the ENSO mode depends on which atmospheric-model component the ocean is coupled to. But the large difference in the two GISS/HYCOM results (b) is mainly caused by changing the target densities and vertical mesh spacing in HYCOM. A grid configuration that minimizes the vertical extent of the depth-coordinate subdomain in the eastern Pacific, thereby allowing the isopycnic subdomain to rise close to the surface, seems to favor large-amplitude El Niño variability in the model. It is tempting to attribute this phenomenon once again to the superior thermal insulation properties of the isopycnic vertical coordinate.

As stressed in the introduction, a model may well be able to satisfactorily predict long-term global change caused by extraneous factors such as increased greenhouse gas concentrations even if it does a less-than-perfect job in simulating ENSO.

Atlantic Overturning during Global Warming. Changes in ocean circulation, particularly in the strength of the meridional overturning circulation (MOC) in individual basins, are considered plausible triggers of rapid climate change (Broecker 2003). What began as a highly technical discussion of this issue has recently seeped into more popular publications (*Fortune*, February 26, 2004; *The Observer*, February 22, 2004). Given the pivotal role played by the Atlantic in moving heat to high northern latitudes (as highlighted in Figure 2), climate researchers are keenly interested in processes that have led to a periodic weakening or outright shutdown of the Atlantic MOC since the last ice age. Foremost among the processes that can trigger such effects is the buildup of a freshwater cap in the subpolar Atlantic by melting land and sea ice. Since seawater density at near-freezing temperatures depends almost entirely on salinity, accelerated ice melt during global warming could conceivably create a strong enough vertical density contrast in the subpolar Atlantic to inhibit the sinking of surface water to the bottom, thereby suppressing the MOC.

Such a shutdown can easily be simulated in an ocean model by imposing an appropriate high-latitude freshwater source. The question is, "how robust a feature is the Atlantic MOC in a climate model?" In other words, is the threshold for an MOC shutdown by ice melt in the model the same as the threshold in the real ocean? Figure 7, taken from the 2001 climate assessment report by the Intergovernmental Panel on Climate Change, indicates that there are vast differences among models in predicting the rate at which the Atlantic MOC will slow down during global warming. Interestingly, from among nine climate models, only an isopycnic coordinate, or layer, model does not indicate a slowdown of the MOC during gradual doubling of atmospheric carbon dioxide (CO_2). This observation suggests that the type of vertical coordinate in an ocean model can greatly influence the outcome of a climate forecast—for reasons touched upon earlier in this article. Further support for the still tentative notion that layer models predict a more stable behavior of the Atlantic MOC during global warming than the eight level models shown in Figure 7 can be found in Sun and Bleck (2001b). Note, however, that the jury is still out on whether "more stable" is synonymous with "more realistic."

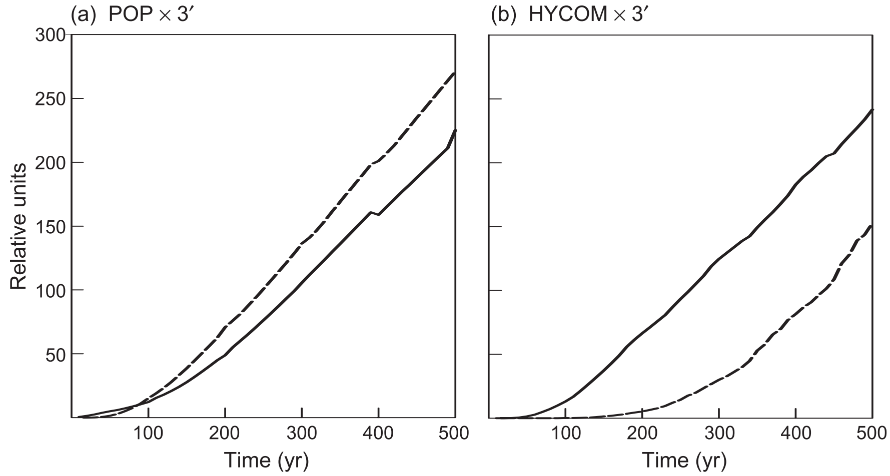


Figure 8. Comparing Carbon Sequestration Results
The two plots show the gradual accumulation of tracer material representing CO_2 (arbitrary mass units) in the top 10 m of the world ocean in (a) POP and (b) HYCOM. The conditions were continuous tracer release at two near-bottom points next to the North American continental shelf off Delaware and California. The tracer injected off Delaware is represented by the solid line; the one injected off California by the dashed line. The source strength at both sites is 1 mass unit per day (36,000 units per century). The discrepancies between the POP and HYCOM results are a manifestation of the uncertainty attributable to numerical approximations in ocean circulation models.

Transport of Sequestered CO_2 in the Ocean. A standard question asked of a climate prediction model is whether its "equilibrium" climate, obtained by running the model for a long time (several centuries) with a time-invariant mixture of atmospheric greenhouse gases and constant solar-energy output (that is, with fixed boundary conditions), resembles the observed climate. Given a century or two, an energy imbalance of a few watts per meter squared, less than 1 percent of the standard solar energy input, will gradually melt the polar ice caps or bring on an ice age in the model. Since the heat capacity of the atmosphere is negligible compared with that of the ocean, radiative imbalances are primarily accumulated in the ocean (including its frozen component). There they create long-term trends in the thermal structure, which sooner or later will disrupt the overturning circulation and the associated poleward heat transport. Interestingly, the drift in global surface temperature accompanying these changes may be as small as a fraction of a degree. (That is why sea surface temperature maps, often presented as an indicator of the performance of an ocean model, are of limited usefulness.)

Much time is being invested at the Laboratory and elsewhere into studying the sensitivity of the modeled MOC to changes in the boundary conditions ("forcing") at the sea surface. In many of these studies, for the sake of computational economy and to avoid contaminating the ocean simulation with atmospheric model errors, the ocean is driven by observed values of temperature, precipitation, wind, or other factors rather than by an atmospheric model that properly reacts to the evolving surface conditions in the ocean model. However, replacing an interacting atmospheric model with prescribed surface fields elicits unforeseen responses in the ocean model. Efforts at Los Alamos to compare the performance of layer and level models in ocean-only experiments have been frustrated by the realization that ocean models show different degrees of tolerance to physically imperfect surface forcing.

Nevertheless, enough progress has been made over the years in formulat-

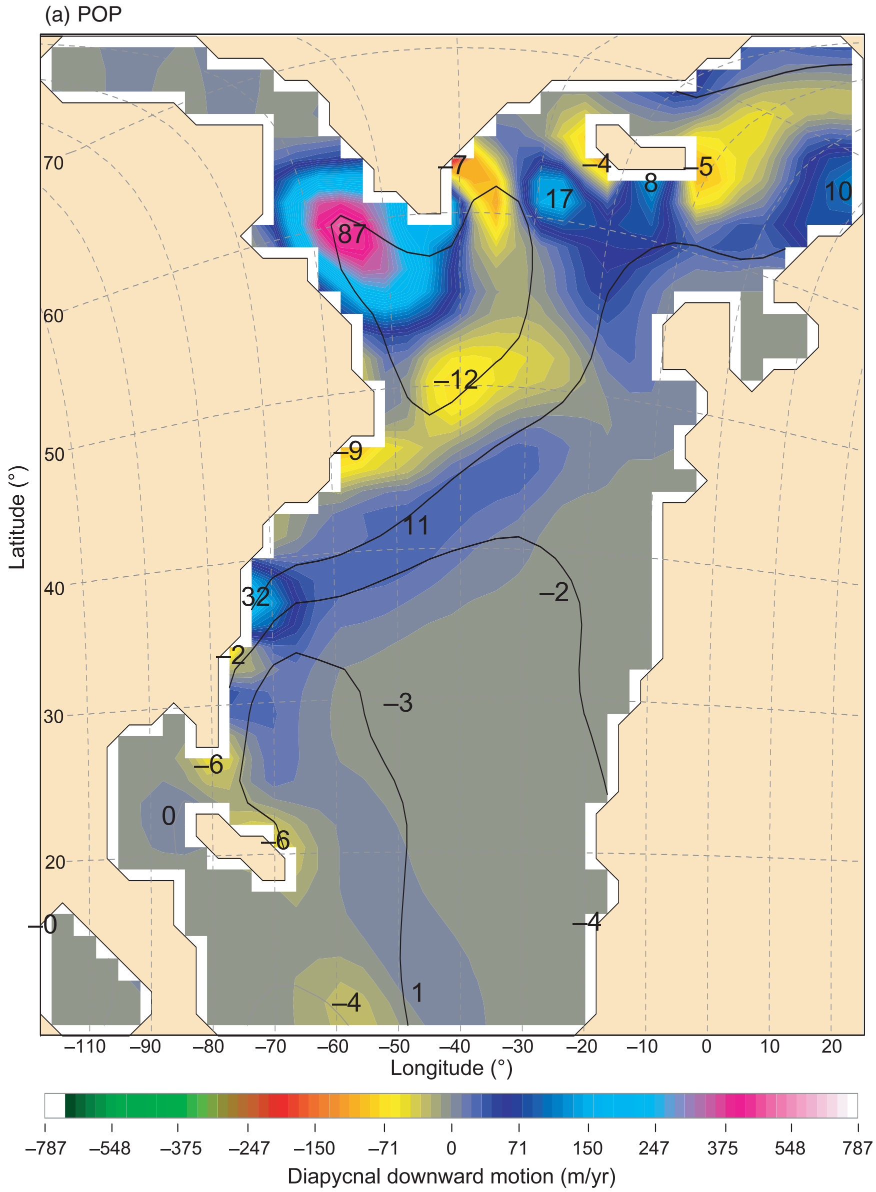


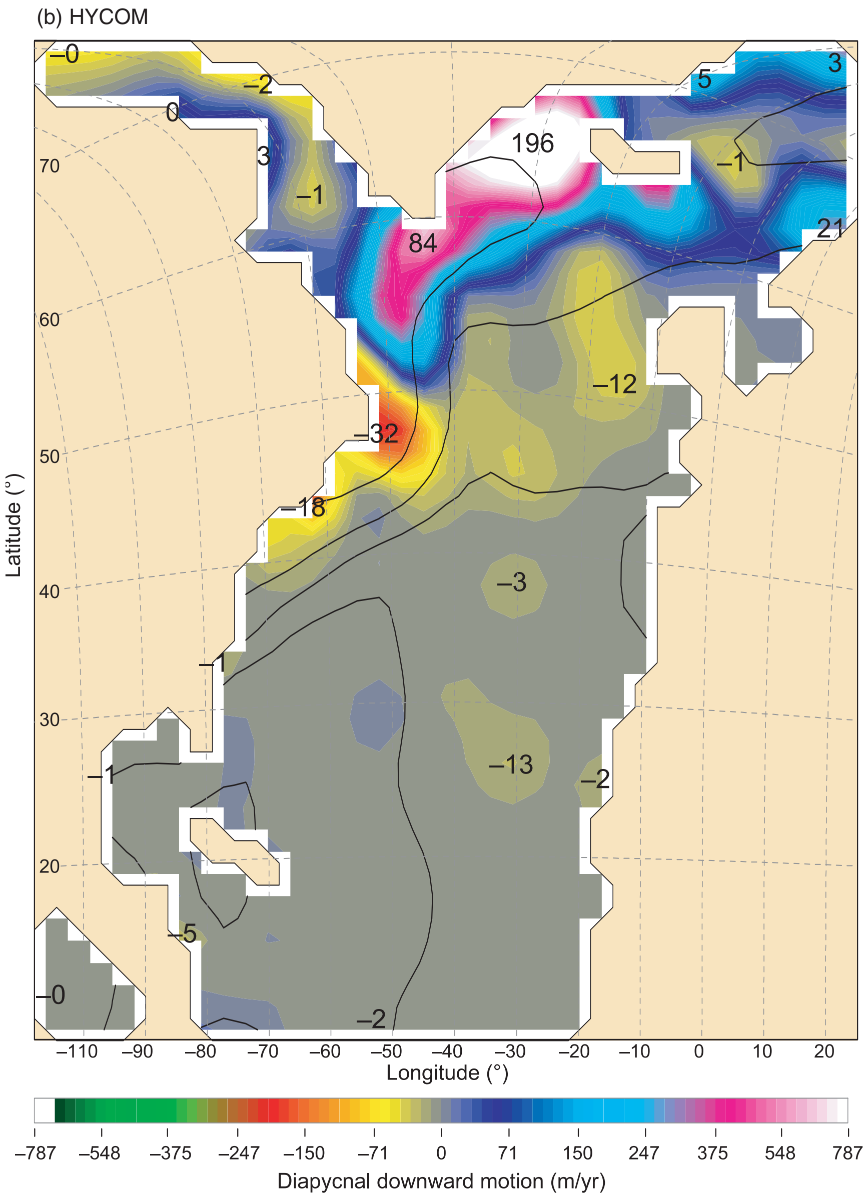
Figure 9. The Downwelling Limb of the Atlantic Overturning Circulation in POP and HYCOM
These isopycnic-coordinate views of the thermally forced Atlantic circulation were obtained with two coarse-mesh models: (a) POP and (b) HYCOM. Each view shows North America at left and Europe and Africa at right. Greenland (grossly deformed by the map projection) is seen at the top. Color contours represent the time-averaged rate (meters per year, positive downward) at which water crosses an isopycnic surface near the interface between the warm and cold limbs of the Atlantic overturning circulation. Numbers overlaying the patches of upwelling and downwelling indicate the total diapycnal mass flux (in units of 0.1 Sv, positive down) associated with each patch. Also shown are sea-surface height contours (at 20-cm intervals), a proxy for streamlines of surface currents. The figure illustrates that numerically dissimilar ocean models will disagree on the strength and geographic distribution of the downwelling limb of the overturning circulation even when subjected to identical surface boundary conditions.

ing internally consistent surface boundary conditions to produce reasonably steady and realistic equilibrium circulation states in ocean-only experiments. These circulation states can be used for a variety of practical applications, among them studies of the efficacy of CO_2 sequestration in the world ocean.

Among options currently under discussion for slowing down greenhouse gas-induced global warming is pumping liquefied CO_2 into the abyssal ocean. Regardless of the potential ecologic side effects or the economic feasibility of this approach—not to be discussed in this article—oceanic carbon sequestration presents an interesting test case for studies aimed at comparing ocean models.

One question that can be addressed through numerical simulation is how much time it would take for CO_2 injected into the deep ocean to come back to the surface. Figure 8 shows the results of such a simulation in which an inert tracer representing CO_2 is continually being released close to the sea floor at two points located at 37°N near the continental shelf off the American East and West Coasts. The curves show the globally averaged near-surface buildup of that tracer as it gradually works its way through the global ocean. This buildup provides a semiquantitative measure of how soon the sequestered CO_2 is likely to re-enter the atmosphere through transport and diffusion alone.

The experimental details can only be sketched here. The simulation is performed with both POP and HYCOM configured on the same



noneddy-resolving horizontal grid and subjected to identical seasonally varying atmospheric forcing. POP uses 25 levels in the vertical direction, whereas HYCOM uses 16 layers. The tracer is transported "offline" using the two models' seasonally varying circulation states averaged over consecutive 3-month intervals. The offline approach is chosen for computational economy. The time step in most fluid models is set by the time it takes for the fastest signal supported by the model equations to propagate from one grid point to the next. In the ocean model, the fastest signals (gravity waves) travel in excess of 200 meters per second (ms^{-1}), but advection by currents is at least 100 times slower. Hence, offline tracer advection, in which gravity waves are not an issue, can be done with a 100 times longer time step, and hence 100 times faster than in the full ocean model itself.[2]

In preparation for tracer transport, horizontal mass fluxes from both models are transformed into isopycnic fluxes (fluxes along isopycnic surfaces) from which the missing diapycnal component is deduced by mass continuity. This transformation is performed to ensure that global ocean-ventilation processes, whose action is modeled most coherently in isopycnic coordinates, act similarly in both models. Plots of diapycnal mass flux fields (Figure 9) indeed indicate that both models maintain an Atlantic overturning circulation that is in fair agreement with the available observational evidence (refer to Figure 2).

The vertical flux fields in Figure 9 are a study in model-to-model variability in their own right. Both models clearly depict the ocean basins surrounding southern Greenland as the region anchoring the downwelling limb of the Atlantic overturning circulation, but differences in local detail are obvious. Note that vertical motion is analyzed here in potential-density space; hence, it depicts areas where individual seawater parcels get either lighter or denser with time. Consequently, upwelling and downwelling patches in Figure 9 coincide with regions where the ocean exchanges heat with the atmosphere. Given that atmospheric cyclones thrive on surface heating (the notorious Cape Hatteras storms are a good example), the different MOC downwelling patterns indicated in Figure 9 are likely to result in large differences in regional weather. Storminess in the Irminger Sea, east of Greenland, for example, would be affected by the surface heat-flux differences indicated in Figure 9(a).

[2] Ongoing efforts at Los Alamos and elsewhere try to lengthen the time step in ocean models by filtering out gravity waves, but the ensuing mathematical complexities are daunting. Gravity waves do serve a purpose, both in reality and in the model: They repair deviations from "geostrophic" equilibrium, a particular balance between velocity and pressure field, on which fluids on a rotating planet rely to counteract the deflecting effect of the Coriolis force.

Overall, the Atlantic MOC appears to be stronger in HYCOM than in POP, consistent with the earlier discussion about differences in vertical diffusion control in level and layer models. Reduced momentum mixing in the vertical direction, that is, lower drag on the wind-driven surface flow, may also be the cause for the somewhat stronger surface circulation in HYCOM. This difference is indicated in Figure 9 by the tighter spacing of sea-surface height contours in the right panel compared with those in the left panel. In geostrophically balanced flow, sea-surface height contours are a proxy for streamlines, like isobars on a weather map.

The salient result from Figure 8 is that, after 500 years, POP has brought 1.5 times more material sequestered off California back to the surface than HYCOM. Model-to-model differences are much smaller for the material sequestered off Delaware. Since the circulation off the U.S. East Coast is dominated by strong opposing boundary currents representing the cold and warm limbs of the Atlantic MOC, a feature not found off the West Coast, large differences in dispersion from the Pacific and Atlantic release sites are to be expected. HYCOM accentuates those differences more than POP.

These results, which represent ongoing work and remain to be confirmed by additional experiments, are tendered here as a first attempt at quantifying the circulation-related uncertainties in simulating the feasibility of abyssal sequestration of CO_2. These uncertainties are compounded, of course, by uncertainties about the chemical behavior of CO_2 at great depths.

Concluding Remarks

This article has presented some of the tools used by the research community to assess the uncertainty in decadal to century-scale climate prediction. For the discussion, climate prediction has been cast as a boundary-value problem in which the boundary values of interest (forcings) are assumed to be known. In other words, forcing uncertainties, which are a major point of debate in their own right, have not been considered. Instead, the focus in this article is on error sources within climate models. Limiting the number of climate-relevant natural processes, as well as parameterizing processes that are deemed important but take place on scales too small to be resolved by the model's space-time mesh, creates one type of errors: type 1, or physical-closure, errors. The conversion of the underlying differential equations into computer-solvable algebraic equations, which mainly affect processes the model is designed to resolve explicitly, results in another type of errors: type 2, or numerical, errors.

To guide future model development, the effects of these two error types on the performance of a climate model need to be separated. At least in principle, one can separate those effects either by manipulating type-1 errors (by, for example, adding/subtracting earth system processes or refining certain physical closure schemes) or by quantifying type-2 errors through solving the same physical problem with numerically dissimilar models. Unfortunately, experimenting with different mesh sizes in a climate model—the approach usually taken to establish the proximity of a numerical to a "true" solution—typically does little to disentangle the two error types because physical closure assumptions often are tailored to a particular mesh size and are not expected a priori to work well if the resolution is changed.

Los Alamos is making important contributions in this area by supporting the development and use of multiple ocean models in climate simulation. The numerical diversity in the Laboratory's model ensemble is achieved by support of both level and layer ocean models. The former discretize the underlying differential equations on a Cartesian grid whereas the latter use a material, or Lagrangian, vertical coordinate tied to the oceanic potential-density field, a proxy for entropy. Vertical dispersion of physical properties is handled very differently in these two types of models. Because subsurface oceanic processes are adiabatic (except for mixing) and hence are governed by the entropy conservation law on centennial time scales and beyond, numerically different approaches to satisfying the second law of thermodynamics can lead to profoundly different equilibrium circulation states in long-term ocean simulations. In fact, the sensitivity of the model solution to discretization (type-2) errors in the thermodynamic and dynamic equations often overshadows the sensitivity to the physical-closure assumptions (type-1 errors).

Shortcomings of layer models having to do with the difficulty of defining constant-density surfaces in unstratified regions (regions in which water density does not vary with depth) have led to the development of so-called hybrid-coordinate models, which also are included in the ocean model mix at Los Alamos. ■

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Further Reading

AchutaRao, K., and K. R. Sperber. 2002. Simulation of the Elf Niño Southern Oscillation: Results from the Coupled Model Intercomparison Project. *Clim. Dyn*. **19** (3–4): 191.

Bleck, R., C. Rooth, D. Hu, and L. T. Smith. 1992. Salinity-Driven Thermocline Transients in a Wind- and Thermohaline-Forced Isopycnic Coordinate Model of the North Atlantic. *J. Phys. Oceanogr*. **22**: 1486.

Bleck, R. 2002. An Oceanic General Circulation Model Framed in Hybrid Isopycnic-Cartesian Coordinates. *Ocean Model.* **4** (1): 55.

Broecker, W. S. 1991. The Great Ocean Conveyor. *Oceanogr*. **4**: 79.

Hirt, C. W., A. A. Amsden, and J. L. Cook. 1974. An Arbitrary Langrangian-Eulerian Computing Method for All Flow Speeds. *J. Comput. Phys.* **14** (3): 227.

IPCC. 2001. Climate Change 2001: The Scientific Basis—Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Edited by J. T. Houghton, Y. Ding, D. J. Griggs, M. Noguer, P. J. van der Linden, X. Dai, et al. Cambridge: Cambridge University Press.

Malone, R. C., R. D. Smith, M. E Maltrud, and M. W. Hecht. 2003. Eddy-Resolving Ocean Modeling. *Los Alamos Science* **28**: 223.

Schmitz Jr., W. J. 1996. "On the World Ocean Circulation. Vol. 1, Some Global Features/North Atlantic Circulation." Woods Hole Oceanogr. Inst. Tech. Rep. WHOI-96-03.

Sun, S., and R. Bleck. 2001a. Thermohaline Circulation Studies with an Isopycnic Coordinate Ocean Model. *J. Phys. Oceanogr.* **31** (9): 2761.

———. 2001b. Atlantic Thermohaline Circulation and Its Response to Increasing $C0_2$ in a Coupled Atmospheric—Ocean Model. *Geophys. Res. Lett.* **28** (22): 4223.

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