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Climate Prediction: *The Limits of Ocean Models*

Peter H. Stone

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Climate Prediction: *The Limits of Ocean Models*

Peter H. Stone[*]

Abstract

We identify three major areas of ignorance which limit predictability in current ocean GCMs. One is the very crude representation of subgrid-scale mixing processes. These processes are parameterized with coefficients whose values and variations in space and time are poorly known. A second problem derives from the fact that ocean models generally contain multiple equilibria and bifurcations, but there is no agreement as to where the current ocean sits with respect to the bifurcations. A third problem arises from the fact that ocean circulations are highly nonlinear, but only weakly dissipative, and therefore are potentially chaotic. The few studies that have looked at this kind of behavior have not answered fundamental questions, such as what are the major sources of error growth in model projections, and how large is the chaotic behavior relative to realistic changes in climate forcings. Advances in computers will help alleviate some of these problems, for example by making it more practical to explore to what extent the evolution of the oceans is chaotic. However models will have to rely on parameterizations of key small-scale processes such as diapycnal mixing for a long time. To make more immediate progress here requires the development of physically based prognostic parameterizations and coupling the mixing to its energy sources. Another possibly fruitful area of investigation is the use of paleoclimate data on changes in the ocean circulation to constrain more tightly the stability characteristics of the ocean circulation.

Contents

1. INTRODUCTION

The oceans are a player of fundamental importance in the climate system. One important role is the transport of heat by oceanic circulations. These circulations carry about two petawatts of heat poleward in both hemispheres (Ganachaud and Wunsch, 2003). This may be compared to the total poleward heat transport in the whole climate system, about 5.5 petawatts (Trenberth and Caron, 2001). The ocean transport profoundly influences latitudinal variations in climate (Seager *et al.*, 2002). It also affects the global mean climate by affecting the amount of sea ice in high latitudes. Because of its high reflectivity, sea ice has a substantial effect on the amount of solar energy absorbed by the climate system, and thus changes in the amount of sea ice can cause global warming or cooling. Another important role of the oceans is the mixing of heat into the

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deep oceans. This mixing determines how rapidly surface temperatures change (Hansen *et al.*, 1985). In a global warming scenario, if the mixing is strong the surface warming will be retarded. Thus any attempt to model or predict climate change requires a good understanding of how the oceans operate.

That our understanding of the climate system as a whole has not yet reached the level where reliable projections can be made is obvious from the lack of robustness of climate change projections made with different state-of-the-art climate models. For example, Cubasch and Meehl (2001) compared projections of changes in the meridional overturning circulation in the North Atlantic from 10 different coupled atmosphere-ocean general circulation models (GCMs) for the same global warming scenario. This circulation is illustrated in **Figure 1**. The poleward flow near the surface is primarily associated with the Gulf Stream. This circulation is particularly important for climate, because it transports more heat than the circulations in any other ocean basin, and has a substantial warming effect on mid and high latitudes of the Northern Hemisphere (Seager *et al.*, 2002). Estimates of the strength of the overturning circulation range from 16 to 25 Sv (Macdonald and Wunsch, 1996; Ganachaud, 2003; Sv = one Sverdrup = 10^6 m^3/s). However the simulated changes in this circulation by 2100 varied from no change to a decrease of 14 Sv. Since this result comes from coupled models, it is not possible to identify any single component of the climate system, such as the oceans, as being the source of the differences, without further analysis.

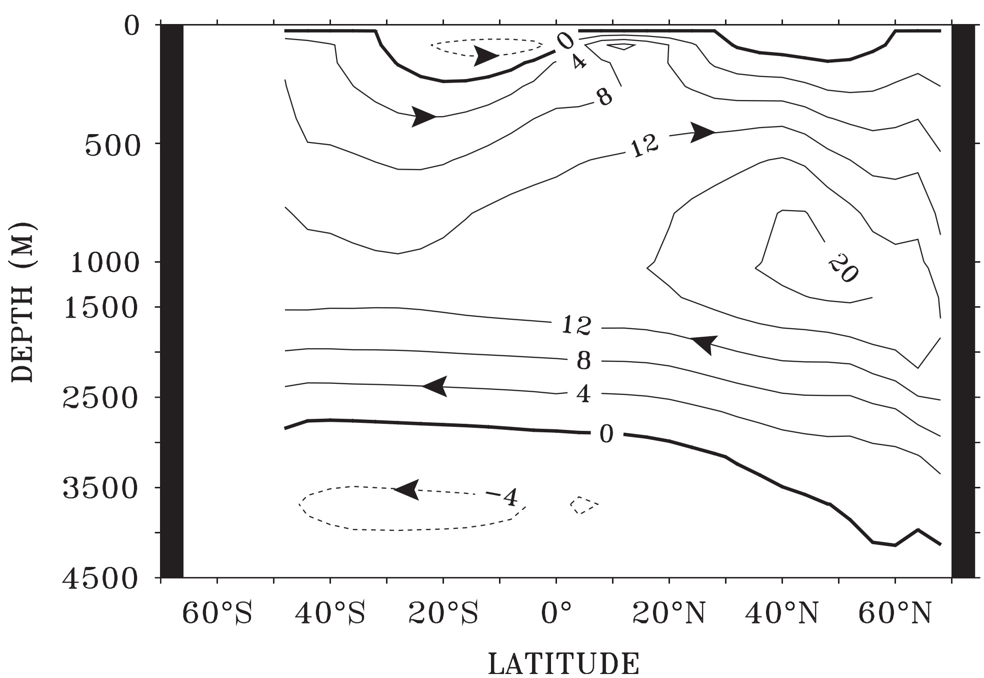

Figure 1. Typical model simulation of the stream function of the zonal mean overturning circulation in the North Atlantic. Depth is given on the vertical axis and latitude on the horizontal axis. Adapted from Huang *et al.* (2003).

An analysis which does implicate the ocean component of the climate models has been carried out by Sokolov *et al.* (2003). They found that model differences in projections of changes in global mean surface temperature could be attributed to differences in two model characteristics. One is the model's climate sensitivity, defined as how much the global mean surface temperature would increase if the concentration of CO_2 in the atmosphere were doubled and the climate system were allowed to equilibrate. This sensitivity depends primarily on atmospheric processes such as how clouds change when climate changes. These processes are not well understood and are represented in different ways in different models. The second model characteristic is the rate at which perturbations in the heat flux between the atmosphere and ocean are mixed into the deep oceans.

Figure 2 shows how 11 different coupled atmosphere-ocean GCMs differ with respect to these two characteristics. In the figure, the rate of heat uptake by the deep oceans is measured by the global mean value of a coefficient which describes the effective rate at which heat anomalies are mixed into the deep ocean. In the figure the square root of this coefficient is plotted, since the depth to which heat penetrates at a given time is proportional to the square root of the coefficient. As shown in Figure 2, this depth varies between models by a factor of two and one half. The rate of heat uptake is not well constrained by the available observations (Forest *et al.*, 2002), so none of these models can be ruled out by comparing them with the observations. Similarly we cannot be sure that any of them are right.

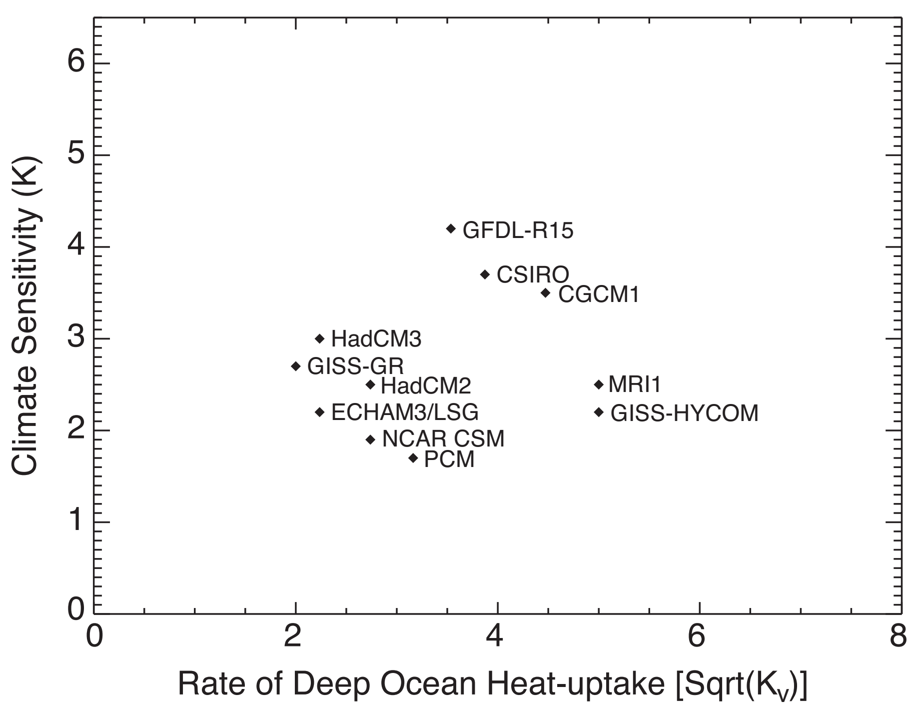

Figure 2. Properties of 11 different coupled GCMs. Vertical axis: climate sensitivity. Horizontal axis: a parameter measuring the depth to which heat has penetrated in the deep ocean (see text). Adapted from Sokolov *et al.* (2003).

One likely source of the ocean model differences in the rate of heat uptake is the different representations of small-scale oceanic processes used in different models. The differences reflect our ignorance of these processes, and this is one potential obstacle to our current ability to predict climate change. This problem will be discussed in Section 2.

Another potential obstacle is the possibility that the circulation in the North Atlantic and its heat transport may be very sensitive to small changes in climate. Since this circulation is coupled to that of the rest of the oceans by the "conveyor belt" circulation, such changes would have global consequences. Uncoupled ocean models that show this possibility include simple box models (Stommel, 1961; Rooth, 1982; Welander, 1986), two-dimensional meridional plane models (Marotzke *et al.*, 1988), and three-dimensional numerical models (Bryan, 1986; Marotzke and Willebrand, 1991). They all show that the circulation is very sensitive to salinity perturbations, particularly at high latitudes, and that the circulations can have at least two states. One is like that currently existing in the North Atlantic Ocean, with a relatively strong poleward heat transport. The other has a much weaker circulation with very little poleward heat transport.

Paleoclimatic evidence also indicates that two states like these with very different climates can exist (Broecker *et al.*, 1985; Boyle and Keigwin, 1987; Broecker, 2003). Indeed, Broecker *et al.* (1985) suggest that sudden shifts in climate, such as that associated with the Younger Dryas event some 10,000 years ago, may have been caused by a sudden collapse in the circulation of the North Atlantic. How this phenomenon may limit predictions of climate change will be discussed in Section 3.

The limits on prediction described above could in principle be overcome if we could acquire data that is sufficiently extensive and accurate, and if our computers were sufficiently fast. However there may be a more fundamental limitation to our ability to predict changes in the oceans. The oceans' circulations are highly nonlinear, but only weakly dissipative. Such systems are potentially chaotic, *i.e.*, unpredictable past a certain time limit. This possibility will be discussed in Section 4. Finally, in Section 5, we will summarize our results and discuss possible paths for improving the predictions of ocean models and determining the limits of their predictability.

2. SMALL-SCALE OCEANIC PROCESSES

The ocean GCMs used in current climate models have coarse resolution; typical horizontal resolutions are in the range 1° to 3°. Thus there are many subgrid-scale processes that need to be parameterized in these models. In current practice these processes are generally decomposed into four components which are parameterized separately: diapycnal diffusion, isopycnal diffusion, mesoscale eddies, and convection. Diapycnal diffusion refers to diffusion perpendicular to constant density surfaces, while isopycnal diffusion refers to diffusion along constant density surfaces. Mesoscale eddies are eddies with typical spatial scales of about 100 km and typical periods of about 100 days. Energy spectra of the oceans show a peak at the frequency of the mesoscale eddies (*Wunsch,* 1981). The other parameterized processes occur at smaller spatial scales. There are major uncertainties and problems in current parameterizations of all these processes.

Diapycnal diffusion plays a particularly important role in determining the ocean's circulation, since it is the diapycnal mixing of heat and salinity from the ocean's surface into its depths that gives rise to the density gradients that drive the large-scale ocean circulation and its horizontal heat transports (Munk and Wunsch, 1998). In fact, scaling analyses and ocean GCM calculations show that the strength of the ocean circulations and heat transports are sensitive to the value of the diapycnal diffusion coefficient (Bryan, 1987; Marotzke, 1997). In a basin like the North Atlantic, the strength of the meridional overturning is approximately proportional to the 2/3 power of the coefficient and the poleward heat transport to the 1/2 power (Marotzke, 1997). The strength and heat transport are determined primarily by the values of the diapycnal diffusion at depths of 200 to 500 m in the tropics and subtropics (Scott and Martozke, 2002; Bugnion and Hill, 2003).

However OGCMs generally treat the diapycnal diffusion coefficients for heat, salinity, and momentum as constants, or as specified functions of depth. These representations are unlikely to be realistic. For example, one would expect the coefficients in general to depend on the shear and/or the stratification. Furthermore the values of the coefficients in the current climate are quite uncertain, with different measurements and estimates giving a range of 10^{-4} to 10^{-5} m^2/s (Munk and Wunsch, 1998). This is at least in part because they have strong spatial variations (*e.g.*, Polzin *et al*., 1997).

OGCM calculations show that vertical mixing by the other three subgrid-scale processes is strongest in high latitudes (Huang *et al.*, 2003a and 2003b). This is because the strong cooling of surface waters in high latitudes favors static instability and a vertical orientation of isopycnals. The former leads to convection; the latter leads both to isopycnal diffusion being predominantly vertical and to large amounts of potential energy being available for mesoscale eddies. The efficiency of all these processes is usually parameterized by specifying a constant diffusion coefficient.

The values of these coefficients are again poorly known. Estimates of the isopycnal diffusivity range from 500 to 2000 m^2/s (Hirst and Cai, 1994; Jenkins, 1991). The most popular parameterization of mesoscale eddies is the Gent-McWilliams parameterization, which requires the specification of both an isopycnal diffusion coefficient and a diffusion coefficient parameterizing the effect of the mesoscale eddies on the density field (Gent and McWilliams, 1990). The two diffusivities are commonly (but arbitrarily) taken to be the same. Eddy-resolving simulations show that in fact the mesoscale eddy diffusivity varies over a range of 10 to 10^7 m^2/s (Nakamura and Chao, 2000).

There are also theoretical reasons for questioning the adequacy of the parameterizations of high-latitude mixing. A fundamental limitation of the Gent-McWilliams parameterization is its assumption that mesoscale eddies' energy source is potential energy, whereas eddy-resolving simulations show that the kinetic energy of the mean flow is also an important source of eddy energy (Solovev *et al.*, 2002). In the case of parameterizations of convection, current schemes neglect the inhibiting effect of rotation on vertical motions (Marshall and Schott, 1999).

Finally we note that the calculation of the large-scale circulations in ocean GCMs is dependent on numerical schemes that are not perfect. Because of their inaccuracies there may be a significant amount of numerical diffusion, *i.e.*, artificial mixing, in a model. Indeed it has been suggested that the unusually rapid mixing of heat into the deep ocean found in a global warming scenario with the GISS-HYCOM model (Sun and Bleck, 2001; Sokolov *et al.*, 2003; see Figure 2) may be an artifact due to numerical diffusion in the HYCOM model (R. Bleck, personal communication).

3. STABILITY OF THE GLOBAL OCEAN CIRCULATION

As noted in the introduction, all ocean models show the possibility that the ocean circulation can be very sensitive to salinity perturbations and therefore to changes in surface freshwater fluxes. This sensitivity is closely associated with the fact that ocean models show the existence of more than one equilibrium state under some circumstances. These multiple equilibria arise because of a positive feedback associated with the advection of salinity in a circulation like that illustrated in Figure 1.

In this circulation the sinking is located in high latitudes, because that is where the surface waters are most dense. The density is a maximum there because the surface waters are coldest there. However the waters in high latitudes are relatively fresh compared to the subtropics because in high latitudes precipitation exceeds evaporation, while in the subtropics evaporation exceeds precipitation. Thus the poleward flow near the surface in a circulation like that shown in Figure 1 (basically the Gulf Stream) brings saltier water into high latitudes, and this tends to raise the density of the high latitude surface waters. Thus, this advection supplies a positive feedback to perturbations in the strength of the circulation. For example, if the circulation is weakened, the salinity advection weakens, the density of high latitude surface waters is decreased, and this weakens the circulation even more. Given a sufficiently strong initial decrease in the circulation, it will collapse. As noted earlier, paleoclimate evidence does indicate that similar state changes have occurred in the past.

This behavior can be illustrated in a model by tracing out a hysteresis loop (Stocker and Wright, 1991; Rahmstorf, 1995a). Two such hysteresis loops, calculated with the Rooth (1982) box model, are shown in **Figure 3**. The equilibrium strength of the meridional overturning circulation in the Atlantic Ocean is plotted vs. the moisture flux into high latitudes of the North Atlantic, F_1. A positive circulation means that there is a strong poleward heat flux into high latitudes of the North Atlantic, and, in this model, a weak poleward heat flux into high latitudes of the South Atlantic. A negative circulation implies the opposite. The former state is the one analogous to that of the Atlantic in the current climate.

As the figure shows, there is a range of values of the moisture flux where two equilibria exist. For smaller values of the moisture flux only the state with strong poleward heat flux in the North Atlantic can exist; for larger values of the moisture flux only the state with strong heat flux in the South Atlantic can exist. If the system is in the former state, a sufficiently large positive perturbation added to the moisture flux will cause this state to collapse to the other equilibrium

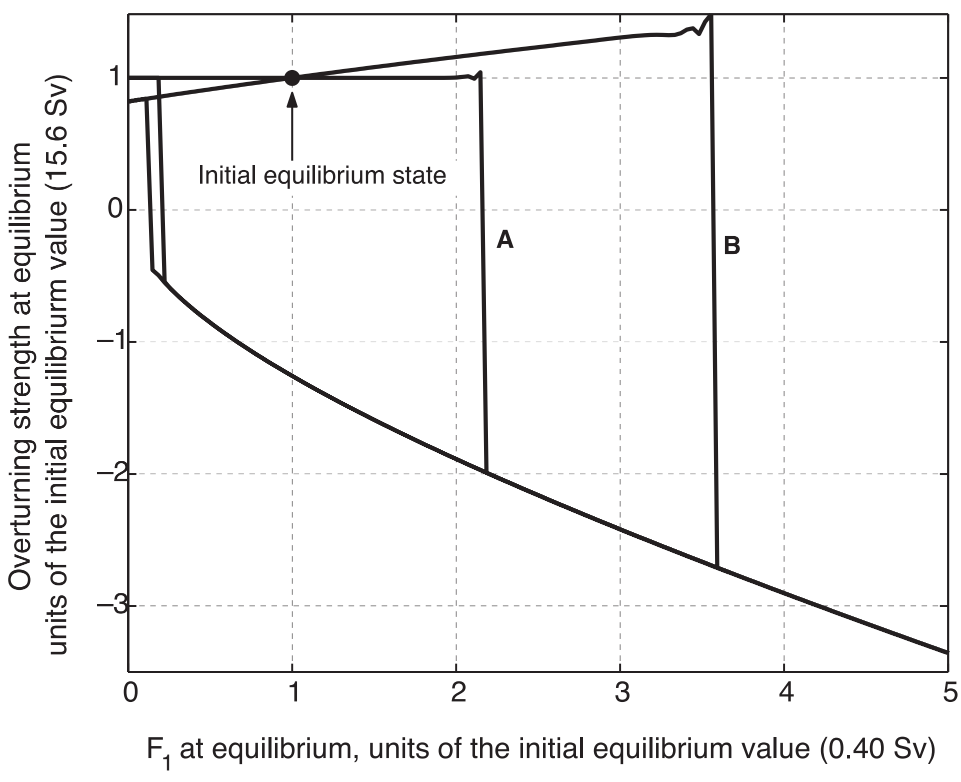

Figure 3. Hysteresis loops calculated from the Rooth (1982) box model with mixed boundary conditions. Vertical axis: strength of the meridional overturning circulation in the Atlantic normalized by its value in the current climate. Horizontal axis: atmospheric moisture flux from low to high latitudes in the Northern Hemisphere, normalized by its value in the current climate. Curve A assumes that the atmospheric moisture flux in the Southern Hemisphere is kept fixed at its value in the current climate. Curve B assumes that Southern Hemisphere flux is increased from its current climate value by 20% of the increase in the Northern Hemisphere.

state, with a consequent large change in the oceanic heat transport and climate. How big a perturbation is required to accomplish this depends on many things. One factor is illustrated by the difference of the two hysteresis loops shown in Figure 2. Curve A is plotted under the assumption that the moisture flux into high latitudes of the South Atlantic does not change when F_1 changes. Curve B shows how the equilibrium state depends on F_1 when there is a simultaneous perturbation of the moisture flux into the high latitudes of the South Atlantic equal to 20% of F_1. As the figure shows, increased moisture flux into southern high latitudes is a stabilizing influence, *i.e.*, it takes larger perturbations in F_1 to shift the system from one equilibrium state to the other.

A question of major importance to our understanding of the sensitivity of climate and its predictability is the question of where on the upper branch of the hysteresis loop the current climate is located. Ideally this question should be addressed with the most sophisticated state-of-the-art coupled GCMs. However to trace out such a curve with one of these models is not computationally feasible. To do so requires either very many integrations with different values of F_1, or a single integration in which F_1 changes very slowly so that the model will evolve through the whole series of possible quasi-equilibrium states. This would require 10,000 or more years of integration, and no coupled GCM has yet been used to calculate such a hysteresis loop.

Recently however hysteresis loops for 11 different models of intermediate complexity have been calculated as part of an intercomparison project for earth models of intermediate complexity (EMICs). EMICs are models which have less detail than state-of-the-art coupled GCMs, but do contain representations of all of the physical processes present in coupled GCMs, (Claussen *et al.*, 2002). The results were reported at a workshop at the annual meeting of the European Geophysical Society in April, 2003. There was no agreement among the models as to the position of the current climate. All the models did have the position being on the upper branch of the hysteresis loop, as it has to be in order to be consistent with the modern climate, but the locations varied from being far to the left of the hysteresis loop, in the monostable regime, corresponding to a very stable climate, to the position being in the bistable region near the bifurcation at the right side of the loop, corresponding to a state with very weak stability.

Actually the situation appears to be even more complicated than is indicated by the simple hysteresis loops illustrated in Figure 3. EMICs with an ocean GCM and realistic ocean bathymetry indicate the possibility of more than two equilibrium states, with the upper branch of the loop having a more complicated structure than that illustrated. In particular different states with somewhat different strengths for the overturning circulation are possible, depending on the sites of high latitude convection in the North Atlantic (Rahmstorf, 1995b).

The diversity of the model results for the state of the ocean circulation ultimately arises from the uncertainties in the input parameters for the climate models. One example is obvious from

Figure 3, *i.e.*, one needs to know accurately the values of the freshwater flux into the high latitudes of the Atlantic Ocean. Since these fluxes depend on precipitation and evaporation over the oceans, where measurements are sparse, the errors are large, of order ±30% (Schmitt *et al.*, 1989). In addition we note that the equilibrium states are not steady states, but rather contain fluctuations, presumably about a fixed climate state (see Section 4 and Figure 5 below). Also, if the climate forcing is not steady, as for example when greenhouse gases increase, the equilibrium states and the hysteresis loops will change.

Another major source of uncertainty involves again the uncertainty in small-scale oceanic mixing processes. **Figure 4** illustrates two hysteresis loops calculated from an EMIC which includes an ocean GCM (Kamenkovich *et al.*, 2002). In order to complete the calculations in a reasonable amount of time, the moisture flux into the North Atlantic was taken to evolve somewhat more rapidly than required for the plotted states to be precise equilibrium solutions,

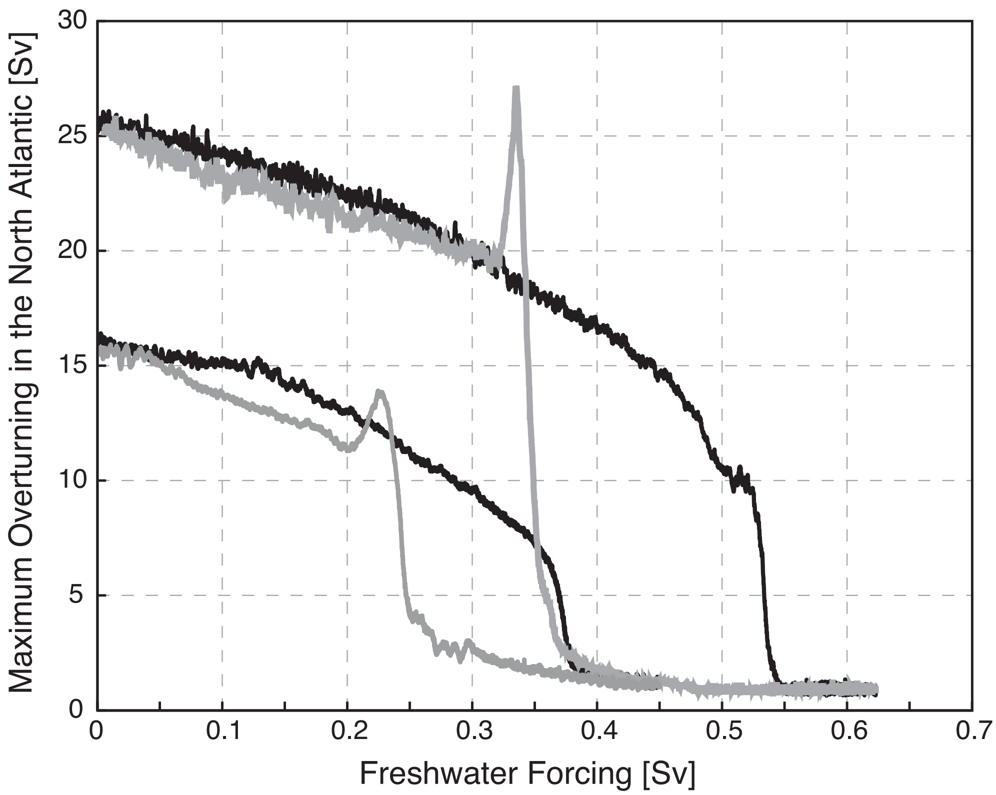

Figure 4. Hysteresis loops calculated with the MIT model of intermediate complexity (Kamenkovitch *et al.*, 2002). Vertical axis: strength of the meridional overturning circulation in the North Atlantic. Horizontal axis: moisture flux into the North Atlantic minus its value in the current climate. The states were traced out by starting with the current climate, then increasing the freshwater flux into the North Atlantic by 0.1 Sv/1000 years, and then after the circulation collapses, reversing the trend and returning to the current climate. The upper curve was calculated with a diapycnal diffusivity of 0.5 cm^2/s, the lower one with 0.2 cm^2/s. Adapted from Dalan (2003)

and thus the forward and return branches of the hysteresis loops do not coincide precisely. Note that in these calculations there was no change in the moisture flux into the South Atlantic, and that in Figure 4 on the horizontal axis is plotted the change in the moisture flux into the North Atlantic from that in the current climate, rather than the actual flux. The two hystersis loops were calculated for different values of the ocean model's diapycnal diffusion coefficient, the upper one being for 0.5 cm^2/s, and the lower one for 0.2 cm^2/s.

As shown in the figure the hysteresis loops are displaced considerably from each other, and correspondingly the stability properties of the system are quite different, with the system being much less stable with the smaller value of the diffusivity. The intersection of the hysteresis curves with the vertical axis gives the strength of the overturning circulation in the North Atlantic in the current climate for the two values of the diapycnal diffusivity. Unfortunately, as we noted earlier, the strength is uncertain.

4. CHAOTIC BEHAVIOR

As noted in the introduction, oceanic circulations are likely to be chaotic, *i.e.*, their evolution is likely to be very sensitive to the initial conditions. This behavior is well known in the atmosphere, and has been studied extensively with atmospheric GCMs. The results show that weather cannot in principle be predicted more than about two weeks in advance because small errors in the initial conditions grow so rapidly. The dynamical time scales in the oceans are much longer than in the atmosphere, of order decades and centuries rather than days, and this makes it much more difficult computationally to assess how chaotic behavior may limit the predictability of ocean circulations. There have only been two studies using ocean GCMs which have attempted to determine if such limits do exist. One by Griffies and Bryan (1997) (hereafter referred to as GB) looked at the predictability of fluctuations in the North Atlantic circulation; the other by Wang *et al.* (1999) (hereafter referred to as WSM) looked at the predictability of regime changes, *i.e.*, of changes between different branches of the hysteresis loops discussed in the previous section.

GB used a coupled atmosphere-ocean GCM in their study. They carried out a thousand-year integration with fixed forcing corresponding to the current climate. In this integration there were fluctuations in the strength of the meridional overturning circulation of the North Atlantic, as illustrated in the top of **Figure 5**. They then carried out an ensemble of 12 integrations in which the initial state of the oceans was taken from year 130 of the control run, but the initial state of the atmosphere varied, being picked from 12 different years in the control runs (but all from the same calendar date). Thus only the weather in the initial atmospheric state differed in the 12 runs. The results for the evolution of the strength of the meridional overturning circulation in the

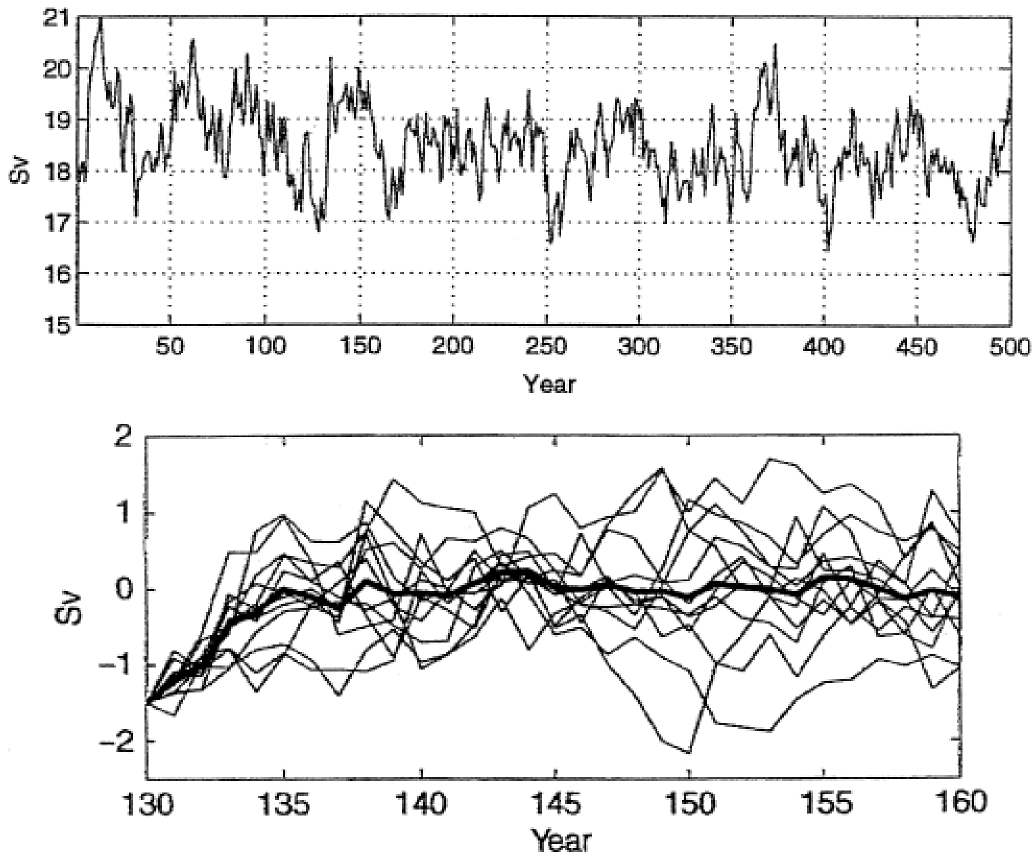

Figure 5. Top: strength of the meridional overturning circulation in the North Atlantic vs. time from a 500-year segment of a control run with the GFDL coupled GCM. Bottom: same as the top figure, except the difference in the strength of the circulation from the mean of the control run is plotted on the vertical axis, and the results are taken from 12 different experiments, all starting from the oceanic state at year 130 in the control run, but with different initial conditions in the atmosphere. The thick line indicates the mean of the 12 experiments. Adapted from Griffies and Bryan (1997).

North Atlantic are shown in the bottom of Figure 5. We see that the ensemble members diverge, and GB found using a statistical test that there is some reasonable predictability of the circulation strength only for the first 3 years. This result is the oceanic analog (for this model) of the prediction limit for atmospheric weather.

However from the point of view of climate, the GB result is not so relevant. The fluctuations in the circulation strength shown in Figure 5 are analogous to fluctuations in weather, and they all occur within the same climate regime. From the point of view of climate, a more interesting question is, what happens if the forcing changes? Is there a limit on our ability to predict regime changes? WSM examined this question using an ocean GCM with idealized global geometry. The ocean was forced by specified moisture fluxes and wind stresses, and the heat flux was calculated from a relaxation condition for the sea surface temperature. In the control run all these boundary conditions were based on the current climate. In addition a stochastic forcing was added to the wind stress boundary condition in order to mimic atmospheric weather fluctuations.

WSM then carried out an ensemble of runs in which the strength of the hydrological cycle in the Northern Hemisphere increased linearly, at a rate equal to 0.1% of the strength in the control run, per year. Thus the net precipitation in high latitudes of the Northern Hemisphere slowly increases and there is an equivalent increase in the net evaporation in low latitudes of the Northern Hemisphere. Three runs were carried out with three different choices for the initial value of the stochastic component of the wind stress. The results for the evolution of the strength of the meridional overturning circulation in the North Atlantic are shown in **Figure 6**.

Because of the very slow acceleration of the Northern Hemisphere hydrological cycle, the circulation evolves through a series of quasi-equilibrium states. In these equilibrium states the strength of the circulation does not change because the changes in precipitation and evaporation in the Northern Hemisphere in effect compensate each other. The increased precipitation in high latitudes reduces the density of the surface water there, but the increased evaporation in the

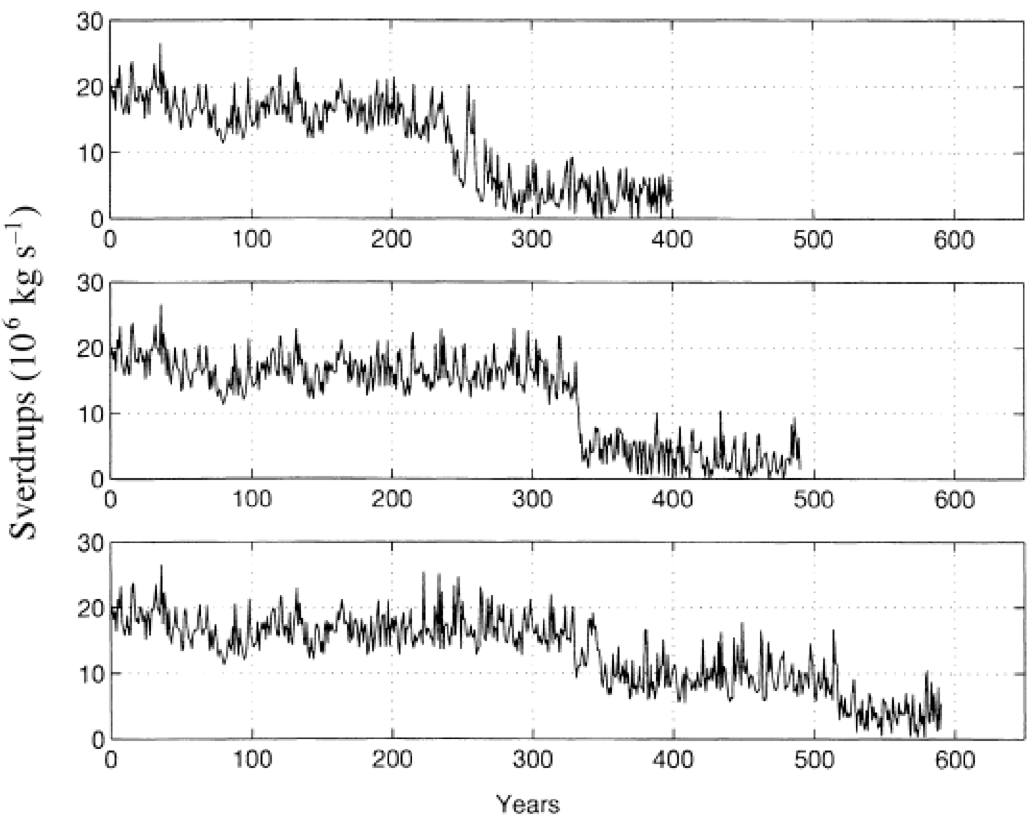

Figure 6. Strength of the meridional overturning circulation in the North Atlantic vs. time from 3 experiments with the WSM model in which the moisture flux into high latitudes of the North Atlantic slowly increased. The only difference between the experiments was the initial value of the atmospheric wind stress. Adapted from Wang *et al.* (1999).

subtropics increases the salinity of the subtropical surface waters, and this increases the advection of salinity into high latitudes. The effect of the latter on the density of the high-latitude surface waters just balances the effect of the former, because there is no net exchange of moisture between the atmosphere and ocean in the Northern Hemisphere as a whole. Thus the system evolves along a hysteresis loop like that shown by curve A in Figure 3.

During the initial phase of the experiments there are interannual fluctuations in the strength of the circulation which are comparable to those in the GB experiments (*cf.* Figures 5 and 6). However there is a striking difference in the nature of these fluctuations. In the WSM experiments, the fluctuations in all three experiments are identical for about 200 years, *i.e.*, the predictability time is much longer than in the GB experiments. One plausible reason for the difference is that the surface heat flux variations in the GB model were much larger and more realistic. Although GB found that the interannual variations in the ocean circulation were largely controlled by the internal ocean dynamics, surface heat fluxes did play a role, and their variations due to weather could have caused the loss of predictability compared to the WSM experiments. We note however that even the more realistic GB model has significant limitations. For example it has coarse horizontal resolution (~5°), which limits the ability to simulate realistic weather fluctuations, and the model can only reproduce the current climate by introducing large unphysical adjustments to the surface heat fluxes.

The more interesting aspect of the WSM experiments is what happened on the longer time scales. As discussed in the previous section the acceleration of the Northern Hemisphere hydrological cycle must eventually lead to a collapse of the strong North Atlantic circulation (as indicated by curve A in Figure 3). It does in all three experiments but, as shown in Figure 6, the timing of the collapse, and the nature of the transition between the two circulation regimes differ considerably. Evidently the differences in the initial condition do not matter until the system approaches a bifurcation, and then there is a complete loss of predictability.

The two studies just described clearly only touch the surface of the problem of how prediction of changes in the ocean’s circulation may be limited by chaotic behavior. For example, it is not clear from these experiments whether fluctuations in the surface heat flux or wind stress are more important in limiting predictability in the ocean circulation on long time scales. In addition neither study looked at how the predictability is affected by perturbations in the initial state of the oceans.

5. POSSIBLE PATHS FORWARD

Forecasts of global warming during the 21st century indicate that the earth is likely to reach global temperatures higher than any it has experienced for at least 100,000 years (IPCC, 2001). This would take the earth to a situation outside the previous experience of our own species as well as that of many others. Thus one of the most formidable scientific challenges facing society is the need to develop a better understanding of how the climate system operates and to predict, to the extent possible, the changes in climate and the environment that society must cope with in the future. Because of the great complexity of the climate system and the many different disciplines that are required to deal with it, this is arguably the most difficult scientific task that has been undertaken. Because the natural response times of the ocean lie in the range of decades to centuries, understanding and predicting its behavior is essential for planning for the next few centuries.

In our discussion of the oceans we have focused on three problems which limit our ability to predict the ocean's behavior. They are 1) our poor understanding of small-scale mixing processes, 2) our inability to characterize the stability characteristics of the ocean circulation, and 3) the presence of chaotic elements in the ocean's behavior. These problems are not independent. For example, the strength and behavior of the mixing properties affect the stability properties, and the stability properties influence the degree of chaotic behavior. In our discussion of the oceans, we focused on the North Atlantic because that is where ocean heat transports are strongest. However the circulations in the North Atlantic are not closed, but rather extend throughout the global oceans, as the "conveyor belt" circulation. Thus these problems are obstacles to understanding and modeling the whole global ocean. Because of these problems simulations of climate change with current state-of-the-art models are problematic.

With regard to the small-scale mixing processes, advances in computer speeds will considerably alleviate at least the problems associated with parameterizations of mesoscale eddies. Since typical scales of these eddies are of order 100 km, models with horizontal resolutions of order 1/10 degree will have much less need to parameterize their effects. Such resolutions should be achievable for global climate models in the near future. Because oceanic energy spectra peak at the frequency of mesoscale eddies, this should mark a major advance in our models' capabilities.

Unfortunately the other mixing processes occur on a much finer scale and thus ocean models will have to rely on subgrid-scale parameterizations for them for a long time to come. More observational estimates of vertical fluxes of heat and tracers, particularly in high latitudes, would be useful, but obtaining them is difficult and expensive. In this situation theoretical approaches may be the most fruitful. In particular one needs prognostic parameterizations rather than the empirical schemes based solely on the current climate that are commonly used in current ocean GCMs. One promising approach for improving current parameterizations is to use modern turbulence closure models to derive prognostic parameterizations (Canuto *et al.*, 2001 and 2002).

However even these parameterizations still require the specification of the flux of energy into the oceans that drives the mixing. The major sources of this energy are believed to be surface winds and tidal mixing (Munk and Wunsch, 1998). Thus climate changes which lead to changes in the surface winds might change the ocean mixing. Such an interaction has never been included in a climate model. Another potentially valuable step forward would be to couple these processes.

Because the stability characteristics of the ocean circulation depend on the small-scale mixing processes and surface flux climatologies (*cf*. Figures 3 and 4), improvements in our knowledge of both of these factors would help to determine the stability properties of the current climate. Paleoclimate data could also prove quite useful. There is considerable evidence indicating changes in the ocean's circulation regime in the past (*e.g.*, Broecker, 2003, and references therein) and these data could help constrain a fully coupled climate model to have the right stability properties.

The fundamental nature of the ocean's circulations, *i.e.*, their nonlinearity and weak dissipation, make it inevitable that their behavior will contain some chaotic elements. Computers have played a prominent role in advancing our knowledge of chaotic behavior in other systems, and in principle they could also do so for the oceans. The primary obstacle so far has been the inherently long time scales associated with the oceans. However increases in computer speeds are now reaching the point where one can envisage carrying out ensembles of runs over long time scales with EMICS whose ocean component is an ocean GCM. Similar studies using coupled atmosphere-ocean GCMs are likely to be feasible within a decade or so. One key question that needs to be addressed is the question of whether the major sources of error growth are fluctuations in the ocean or in the atmosphere, and if the latter, which surface flux fluctuations lead to the most rapid error growth. From the point of view of climate the key question that needs to be addressed is, to what extent does this error growth dominate over changes in forcing in controlling climate change?

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