Multiple Equilibria in a Global Ocean General Circulation Model

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ABSTRACT

A global ocean general circulation model is forced using various upper boundary conditions (BCs) on temperature and salinity. Solutions are obtained under restoring and *mixed* BCs (i.e., a restoring condition on the upper-level temperature but using a fixed, specified surface salt flux).

Salt flux anomalies are temporarily applied under mixed BCs, and solutions are obtained in which overturning associated with deep-water formation is either present or absent in the North Atlantic and either vigorous or weak in the North Pacific. A comparison between these solutions helps to clarify the role North Atlantic deep-water formation plays in maintaining the current climate.

The surface heat fluxes differ substantially between the solutions, and their very existence is dependent upon these differences. As a result they are not multiple equilibria of the ocean model alone. Instead, they should be regarded as multiple equilibria of a very crude coupled atmosphere-ocean system. As the global ocean is substantially altered between the equilibria, it is unreasonable to expect that the parameters in the heat flux formulation will remain unchanged. Consequently, the calculated heat flux anomaly may be in error and there is, therefore, no guarantee that the additional equilibria will exist in more sophisticated models. If multiple equilibria do actually exist in such models, they could be quite different from those obtained under the restorative condition.

1. Introduction

Interpretation of the ocean modeling results of Stommel (1961), Rooth (1982), Welander (1986), Bryan (1986), Marotzke et al. (1988), Weaver and Sarachik (1991), Marotzke and Willebrand (1991), Stocker and Wright (1991), and Huang et al. (1992) has led to a growing consensus that the thermohaline circulation can exist in more than one stable state under *identical surface forcing*.

The studies of Bryan (1986), Marotzke et al. (1988), Weaver and Sarachik (1991), Marotzke and Willebrand (1991), and Stocker and Wright (1991) employed mixed boundary conditions (BCs) in which the upper-level temperature is restored to current climatological values of sea surface temperature (SST), while a fixed salt flux is applied. The fixed flux of salt is used to model the absence of immediate restorative feedbacks on sea surface salinity anomalies.

Under these conditions various states with different meridional heat transports were obtained in which North Atlantic Deep Water formation (NADWF) is either absent or very subdued. The existence of such states is supported by geological evidence (Broecker et al. 1985).

The models they employed ranged from an ocean general circulation model (OGCM) of a single, flat basin (Bryan 1986), through zonally averaged global models (Marotzke et al. 1988; Stocker and Wright 1991), to a global OGCM with no bathymetry and highly idealized continental geometry (Marotzke and Willebrand 1991). In this study a global OGCM with bathymetry and a more realistic continental outline is employed to obtain some of the states described by the earlier workers. Solutions obtained include states in which deep-water formation is either present or absent in the North Atlantic and either vigorous or weak in the North Pacific. Two of the states will be contrasted in order to highlight some of the salient features of the thermohaline circulation in the current climate.

The main objectives here, however, are to first clarify what is usually meant by the term *identical surface forcing*. As a result, it will become clear that all of the solutions obtained here and those found in earlier studies are not multiple equilibria of the ocean model alone. Instead we will show that they should be regarded as either different forced solutions or as multiple equilibria of very crude coupled models. The second main objective is to show that the use of a restorative condition on the upper-level temperature (as in all of the modeling results discussed so far) for the purpose of investigating climatic change on a global scale has a fundamental weakness.

The model used is described in the following section.

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The results are presented in section 3 and are discussed in section 4. Conclusions are drawn in the final section.

2. Description of the ocean model

The OGCM used here is the latest version of the Geophysical Fluid Dynamics Laboratory (Princeton) code (Pacanowski et al. 1991) based on the work of Bryan (1969) and Cox (1984) and given the acronym MOM (Modular Ocean Model).

The horizontal tracer grid was initially designed for coupling to an atmospheric model (Moore and Reason 1993). It is an R21 Gaussian grid, which has a longitudinal spacing of 5.625° and a latitudinal spacing of approximately 3.2° .

There are 12 vertical levels ranging in thickness from 25 m at the surface to 900 m in the deep ocean. The horizontal eddy viscosity is artificially large ($9 \times 10^5 \text{ m}^2 \text{ s}^{-1}$) to ensure that the western boundary currents have horizontal scales that are resolved by the coarse grid. The horizontal eddy diffusivity is $2.5 \times 10^3 \text{ m}^2 \text{ s}^{-1}$. The vertical eddy viscosity and diffusivity take the same numerical values of 20, 1.5, and $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in the first, second, and subsequent levels, respectively, in order to crudely simulate a surface mixed layer. Further details regarding the configuration of this model have been given by Moore and Reason (1993).

All of the results that will be presented here represent solutions that have been obtained using the acceleration techniques discussed by Bryan (1984), unless otherwise stated.

3. Results

a. Preliminary experiments

The preliminary experiments conducted are represented in Fig. 1 and will now be briefly described to

provide an overview. The same solutions will then be discussed in more detail.

First, the OGCM is integrated while restoring the upper-level temperature and salinity (with a time lag of 20 days) to annually averaged climatological values based on the data compiled by Levitus (1982). This leads to the solution *R*. The salt and heat fluxes, S_F and Q_F , are then diagnosed over a period of about 50 surface years, as did Weaver and Sarachik (1991) and Moore and Reason (1993). Averaging is necessary because the surface fluxes fluctuate, especially at high latitudes where deep convection occurs. The salt flux is used when mixed BCs are applied, whereas the heat flux field is used in a later experiment.

The immediate feedback on sea surface salinity anomalies can be avoided by imposing *mixed* BCs in which the salt flux is fixed to S_F but the temperature is restored. After further integration under these conditions, the second solution, M_1 (mixed BC solution 1), is obtained.

A salt flux anomaly, ΔS_{F1} , equal to the average salt flux over the Atlantic (from 35°S to 70°N), is then subtracted from the diagnosed values over the Atlantic and is added to the North Pacific (north of 30°N) after it is scaled to ensure that the anomaly has a zero mean. This leads to the next solution, *A* (anomaly solution).

The initial motivation for this experiment was to test the hypothesis that NADWF is maintained by the net loss of freshwater vapor from the Atlantic (Broecker 1991; Stocker and Wright 1991). At the same time it would reveal whether or not North Pacific DWF could occur if the low evaporation rates that prevail in the current climate were artificially enhanced.

The last solution represented in Fig. 1 is obtained by further integration after reverting back to the original mixed BCs (without the salt flux anomaly) to obtain M_2 (mixed BC solution 2). We will see that M_1 and

FIG. 1. Schematic representation of the preliminary experiments conducted. Solutions are obtained under restoring (*R*) or mixed (M_1 and M_2) boundary conditions (BCs). Solution *A* is obtained under mixed BCs but with an anomalous salt flux.

M_2 are fundamentally different. We now consider each of the solutions R , M_1 , A , and M_2 in more detail.

The zonally averaged mean meridional circulation for the (a) global, (b) Pacific, and (c) Atlantic oceans in R are presented in Figs. 2a–c, respectively. In this study we will be particularly concerned with the overturning in the northern oceans, especially that associated with NADWF. In Fig. 2c we see that under restoring BCs the overturning has a magnitude of about 12 Mt s^{-1} . The salt and heat fluxes (S_F and Q_F , respectively) diagnosed from R are presented in Fig. 3.

Upon a switch to mixed BCs Weaver et al. (1991) found that the transition featured variability on decadal time scales. Moore and Reason (1992), using a model very similar to that employed here, obtained variability on the same time scale during the transition, but only if the model had a constant depth—if bathymetry was included no such variability was evident. In our case the transition features small ($<0.05 \text{ Mt s}^{-1}$) fluctuations in the overturning streamfunction about an essentially smooth transition.

The Atlantic solution under mixed BCs, M_1 , is presented in Fig. 4. The temperature, salinity, passive tracer concentration, and the mean meridional circulation are presented in Figs. 4a–d, respectively. The passive tracer is continuously released over the western half of the North Atlantic's subpolar gyre and is restored to a value of 1000 in order to better understand the dynamical processes underlying the overturning.

The temperature, salinity, and passive tracer distributions all indicate a region of near-vertical homogeneity over the southern half of the subpolar seas, which is associated with penetrative convection. The passive tracer is advected and diffused south from this location at depths centered around 2500 m (Fig. 4d), highlighting the complexity of the processes involved in determining the overturning.

The impact of Antarctic Bottom Water is evident in Fig. 4d, as the negative overturning cell near the bottom, and in Fig. 4c, as a reduction in the concentration of the passive tracer. The overturning associated with NADWF is substantially enhanced, with a magnitude of about 22 Mt s^{-1} . This is similar to observational estimates, but this may be fortuitous since the value of the overturning varies from model to model and is dependent upon the details of the salt and heat flux forcing employed (Maeier-Reimer et al. 1991; Moore and Reason 1992), as well as vertical diffusion (Bryan 1986) and the way in which convection is handled (Marotzke 1991).

The salt flux anomaly described above, which freshens the North Atlantic but increases the salinity in the North Pacific, is then applied. After integration a solution is obtained in which overturning in the northern North Atlantic (Fig. 5a) has largely ceased (compare this figure with Fig. 4d) but is now occurring in the North Pacific with a magnitude of about 23 Mt s^{-1} (Fig. 5b).

a Meridional mass transport (MT/s): Global

CONTOUR FROM -55 TO 40 BY 5

b Meridional mass transport (MT/s): Pacific

CONTOUR FROM -55 TO 40 BY 5

c Meridional mass transport (MT/s): Atlantic

CONTOUR FROM -10 TO 12 BY 1

FIG. 2. The zonally averaged mean meridional circulation (or overturning) for the (a) global, (b) Pacific, and (c) Atlantic oceans under restoring BCs.

This is an interesting result: DWF can occur in either basin given a net surface divergence of fresh water, which would seem to indicate that the presence of DWF

FIG. 3. The surface salt and heat fluxes diagnosed over 50 surface years from experiment *R* (i.e., under restoring BCs).

is not critically dependent upon the special geometry or bathymetry of the North Atlantic. Apparently, the presence of semienclosed northern seas (like the Norwegian and Greenland seas) that are separated from the main basin by sills are not crucial in determining whether or not deep water is formed.

This is, however, a simplistic interpretation since the salt flux forcing applied is a function of both the oceanic and atmospheric climatology. The geometrical and bathymetric features of the ocean help to determine the atmospheric climatology via their impact upon SSTs, roughness lengths, and albedos, for example.

a Zonally averaged temperature

CONTOUR FROM -2 TO 30 BY 1

c Zonally averaged passive tracer

CONTOUR FROM 0 TO 1000 BY 25

b Zonally averaged salinity

CONTOUR FROM 27 TO 36 BY .2

d Meridional mass transport (MT/s): Atlantic

CONTOUR FROM -9 TO 22 BY 1

a Meridional mass transport (MT/s): Atlantic

CONTOUR FROM -13 TO 10 BY 1

b Meridional mass transport (MT/s): Pacific

CONTOUR FROM -55 TO 40 BY 5

FIG. 5. The overturning in the (a) Atlantic and (b) Pacific under mixed BCs with an anomalous salt flux added. This solution is referred to as A in the text.

Consequently the evaporation, precipitation, runoff, and therefore the salt flux itself, cannot be regarded as entirely independent of the geometrical and bathymetric differences between the North Atlantic and the North Pacific.

In other words, the oceanic differences do not prevent DWF from occurring in the North Pacific given the appropriate salt flux forcing. The same differences might, however, impact upon the atmosphere in such a way that the appropriate forcing cannot be established.

If the application of the anomalous salt flux is then halted and the original flux, S_F , is again applied under mixed BCs (M_2 , Fig. 6), the North Pacific overturning is substantially weakened but still evident, while very little ($<1 \text{ Mt s}^{-1}$) overturning resumes in the North

a Meridional mass transport (MT/s): Atlantic

CONTOUR FROM -11 TO 10 BY 1

b Meridional mass transport (MT/s): Pacific

CONTOUR FROM -60 TO 40 BY 5

FIG. 6. The overturning in the (a) Atlantic and (b) Pacific under mixed BCs. Solution A served as the initial conditions. The salt flux anomaly was removed during the integration. This solution is referred to as M_2 in the text.

Atlantic. At first this seems surprising—one might have expected that the solution would have reverted back to something like M_1 . This is not the case; a new and fundamentally different equilibrium solution has been obtained.

A comparison of the solutions helps to assess the impact of NADWF in the global thermohaline circulation and its importance in helping to maintain the current climate. Difference fields, $M_2 - M_1$, are presented in Fig. 7. The overturning maintains anomalously high (relative to the zonal average) sea surface temperatures over the northern North Atlantic and southeast of southern Africa, but anomalously low temperatures over the South Atlantic and Southern oceans. At depth the temperature is higher in the presence of overturning but lower over the Pacific.

FIG. 4. The (a) temperature, (b) salinity, (c) passive tracer concentration, and (d) mean meridional circulation in the Atlantic Ocean under mixed BCs. Solution R provided the initial conditions. This solution is referred to as solution M_1 in the text.

FIG. 7.

CONTOUR FROM -2.2 TO .4 BY .1

FIG. 7. The difference between the two solutions obtained under the same mixed BCs, that is, $M_2 - M_1$. Differences in temperature and salinity at the surface and at a depth of 1575 m, panels (a)–(d), and the horizontal velocities at depths of 370 m and 2350 m, panels (e) and (f), are presented.

The overturning in the North Atlantic is associated with raised salinities over the entire Atlantic basin, especially the surface waters, which occurs by freshening deeper waters elsewhere. The role of NADWF in the current climate, therefore, extends well beyond the North Atlantic as we might have expected on the basis

of the work done by Gordon (1986), and Semtner and Chervin (1988).

To help understand why this is so, the difference between the horizontal velocities are presented (Figs. 7e and 7f) at depths of 370 m and 2350 m. It is evident that the enhanced overturning is associated with an

enhanced western boundary current system—the Gulf Stream and its undercurrent. The same plots illustrate that these two currents are a part of a global circulation encompassing the Agulhas and Circumpolar currents, as well as flow through the Indonesian Passage. It is interesting to note that the deep flow into the Atlantic has two components: one moving eastward through Drake Passage, the other westward around the Cape of Good Hope.

b. Multiple equilibria or forced solutions?

We have just described two very different solutions (M_1 and M_2), both obtained under mixed boundary conditions—a fixed salt flux, and restoring to the same climatological values of SST—but with different initial conditions. Similar solutions have been obtained by others using simpler models and have been described as multiple equilibria (Bryan 1986; Marotzke and Willebrand 1991).

For these solutions to be multiple equilibria of the ocean model alone they must have identical boundary conditions; otherwise they represent solutions that differ because they are forced differently. The forcing is provided by the surface fluxes of momentum, salt, and heat. The first two are identical. The third flux is determined by the formula

$$Q_F = \alpha(T - T^*), \quad (3.1)$$

where α is a constant, T the upper-level temperature, and T^* is the climatological value of the SST. The difference between this field in experiments M_1 and M_2 is presented in Fig. 8. Substantial differences are evident

over the northern North Atlantic and Southern oceans. Hence, the solutions do not represent multiple equilibria of the OGCM alone. The *parameterization* of the surface heat flux is the same but the *actual fluxes* are different.

If heat is conserved, then in a statistically steady state (*i.e.*, a state in which the statistics of the flow are stationary) an accumulation of heat at any latitude must be balanced by a surface heat imbalance; otherwise, the temperature will rise. This conservation property can be expressed mathematically as

$$d[Q]/d\Phi = RL(\Phi)\langle [Q_F] \rangle, \quad (3.2)$$

where Q is the total meridional heat transport at a given latitude Φ , L is the basin width, and R the radius of the earth. The angle and square brackets denote zonal and temporal averaging, respectively.

Marotzke and Willebrand (1991) and Stocker and Wright (1991) presented meridional heat transports for each of their solutions in their flat-bottom ocean models and found significant differences between them. As their stable solutions are very nearly statistically steady states, we can infer that the zonally and temporally averaged surface heat fluxes must also differ, and so their solutions are not multiple equilibria of the ocean model alone either.

This point is worth emphasizing. If multiple equilibria exist for the OGCM alone, then surface fluxes of heat, salt, and momentum must have the same statistics (*i.e.*, mean, variance, *etc.*) at every latitude and in every basin. So the climatic significance of the multiple equilibria does not lie in their differing impact upon the atmosphere but rather on the possibility that the cou-

FIG. 8. The difference between the net surface heat flux, $M_2 - M_1$.

pled system will respond to a given perturbation in a different way.

The possibility that the solutions actually differ given the coincidence of the transports everywhere seems intuitively unlikely. Nevertheless, we test for the existence of multiple equilibria by integrating A using fixed flux conditions on both the heat and salt. The resulting solution for the mean meridional circulation in the North Atlantic is presented in Fig. 9. This should be compared with Fig. 2c, the equivalent plot under restoring BCs. The solutions are very similar, indicating that the solution that had no NADWF depended upon the sustained surface heat flux anomaly for its existence. When this anomaly was removed, NADWF returned and the ocean reverted back to a state very similar to its original state R .

As a further test the same experiment was repeated without using the acceleration technique of Bryan (1984). A one-day time step on the tracer equation and a 20-minute time step on the momentum and streamfunction equations were used at every vertical level (after Marotzke and Willebrand 1991) and again, A moved back close to R .

A number of experiments were conducted to see if A could be destabilized by adding a second salt flux anomaly field in the presence of the stabilizing anomalous heat flux field. This was done by trial and error. It was (eventually) found that if an anomaly equivalent to one-half the salt flux over the Atlantic Basin (from 35°S to 70°N) was added to the western half of the North Atlantic's subpolar gyre for a minimum of 450 years, A could be destabilized. The interesting point to note is that after this and under mixed BCs the OGCM moved into a state very similar to M_1 !

Since the existence of M_2 is dependent upon the anomalous heat fluxes, the major differences between M_1 and R might also be dependent on the difference

in the heat flux between them. To test this M_1 was integrated, applying both Q_F and S_F . It went into a state barely distinguishable from R . As a final test, this particular experiment was repeated without employing the acceleration technique. The same transition occurred. Thus, if similar boundary conditions are imposed, similar solutions are obtained.

4. Discussion

We have seen that the existence of the solution M_2 , in which no NADWF is forming, is dependent upon the anomalous surface flux of heat. These fluxes are determined using the simple restoring formula in Eq. (3.1). Haney (1971) showed that the zonally averaged value of the surface heat flux could be approximated in this way, but with T^* as an apparent air temperature.

Ocean modelers have traditionally used the SST as proxy for T^* , as the two are, in most regions, within a couple of degrees Celsius of each other. However, large differences do exist in some regions (Oberhuber 1988). It is also unrealistic to use a proportionality factor, α , which is either constant (Haney 1971) or varying in the meridional direction only (Han 1984; Oberhuber 1988). As a result, the use of the formula together with the substitution of SST for T^* limits the ability to model surface heat fluxes and (vertically averaged, meridional) heat transports.

While these inadequacies should be kept in mind, there is an even more fundamental problem with its use in studies of climatic change: T^* is presumed to be unaltered. This means that different solutions obtained using (3.1) represent climates in which the ocean is significantly different but the climate in general (for which T^* is but one indicator) is presumed to be unaltered—a contradiction.

Studies of climatic change that employ a restoring BC on temperature (e.g., mixed BCs) amount to using a very crude coupled model. This is because the additional solutions owe their existence to the difference in boundary conditions—only one solution was obtained when the OGCM is totally decoupled by imposing fixed surface fluxes of both salt and heat.

Given the crudeness of this coupled model, there is no guarantee that any new, modeled oceanic climates, obtained under mixed BCs, will have counterparts in reality. Both atmospheric and cryospheric responses to such large flux anomalies might substantially alter the evolution of the ocean climate.

On the other hand, Manabe and Stouffer (1988) obtained two stable states in a coupled climate model, which incorporated an atmospheric GCM and sea-ice model—one with and one without NADWF. It is important, however, to keep a few points in mind. First, the surface freshwater flux into the OGCM had to be corrected in order to obtain a reasonable simulation of the current climate, and the corrections have magnitudes that rival the magnitudes of the fluxes them-

FIG. 9. The mean meridional circulation in the North Atlantic after A was integrated using the surface flux fields diagnosed from R and presented in Fig. 3. This should be compared with Fig. 2c.

selves, especially over the northern North Atlantic. The heat fluxes are not presented, but there is no reason to expect that they will be any better. How then do we know that the atmospheric response (and therefore, the fluxes of heat and fresh water) to the SSTs exhibited by the OGCM when NADW is not forming will be appropriate?

Furthermore, we have seen that a salt flux anomaly over the western subpolar gyre can destabilize the state in which no NADW is forming, even under the stabilizing influence of the modeled heat flux. In the same state sea ice formation could be expected to be promoted, but the sea ice model incorporated by Manabe and Stouffer (1988) does not reject brine into the OGCM upon formation.¹ It will therefore be interesting to see if the same conclusions regarding the stability of this state are reached when studies are performed with coupled models having increased resolution and better physics.

Mixed BCs have also been used in studies of the stability of NADWF to high-latitude freshenings (Wright and Stocker 1991; Maier-Reimer and Mikolajewicz 1989). Under these conditions the current climate appeared to be very sensitive to such perturbations. However, the evolution of the climate state away from one in which NADW is forming is not reflected in the heat fluxes: the upper-level temperatures are restored to unaltered proxy values for the apparent air temperatures. Again, the ocean climate has changed substantially, but this change is not incorporated into the surface heat flux parameterization. As a result, the stability of NADWF to such perturbations remains an open question.

In M_2 we saw that if NADWF is shut off, then the heat fluxes estimated from the restoring condition on temperature promoted the stability of this state by warming the surface of the northern North Atlantic and thereby stabilized the water column. In reality there is no guarantee that the atmospheric response to such a shutoff will include a heat flux with this sign. Furthermore, the lower prevailing temperatures in this state will promote sea ice formation and thereby alter the salt flux, which we have seen can restart NADWF. Such a feedback is obviously precluded under conditions of a fixed salt flux. Other inadequacies of specifying a fixed salt flux have already been noted (Bryan 1986; Welander 1986).

Hence, the stability of both states—one in which NADW is forming and the other in which it is not—cannot be properly assessed under mixed boundary conditions.

5. Conclusions

A coarse-grid version of the GFDL global OGCM (Pacanowski et al. 1991) was integrated under various upper boundary conditions on temperature and salinity. Solutions were obtained under restoring and mixed BCs, that is, a restoring condition on the upper-level temperature but using a fixed, specified salt flux.

Implicit in the use of the conventional restorative condition on temperature to an apparent air temperature, T^* (or its proxy, SST), is the assumption that T^* remains fixed. This assumption is valid provided that the current climate is not dramatically altered. Yet the global ocean, the “flywheel” of the current climate, has been dramatically altered. It is, therefore, inconsistent to suppose that the parameter is unchanged.

Furthermore, since large heat flux anomalies form in the northern part of the North Atlantic it is unlikely that sea ice production, and therefore brine rejection, will not be affected. This limits the usefulness of a fixed salt flux condition in this context.

Salt flux anomalies were applied under mixed BCs and various solutions were obtained in which overturning associated with DWF was either weak or vigorous in the North Atlantic and North Pacific. A comparison between two of these solutions helped to clarify the role North Atlantic DWF plays in maintaining the current climate.

In each case the net surface heat fluxes differed substantially, and it was shown that the solutions presented here and in similar studies depend upon these differences for their existence. This indicates that none of the solutions presented are multiple equilibria of the OGCM alone but are different because the forcing is different.

This study highlights the fact that the solutions (i.e., M_1 and M_2) can be regarded as multiple equilibria but in a very crude coupled atmosphere–ocean model. In this model the climatic change evident in the ocean does not extend to the heat flux parameterization, which is based upon the observed relationship between SST and heat flux in the *current* climate, “ M_1 .” It might, therefore, be inappropriate in the new climate, M_2 , particularly in those regions where there are sustained and substantial surface heat flux anomalies. As a result the equilibria in this system need have no counterpart in more realistic coupled models. If, on the other hand, counterparts do actually exist, then they might be quite different to those modeled under mixed BCs.

Finally, we saw that DWF could be promoted in the North Pacific by imposing a freshwater flux divergence anomaly out of this basin and into the North Atlantic. This would seem to indicate that geometrical considerations are not crucial in determining whether or not deep water will be formed in a given basin. Geometrical

¹ Note added in proof: The model *does* in fact reject brine upon ice formation. The fact remains, however, that the surface freshwater flux is corrected and so the statement at the end of the paragraph remains valid.

features of the ocean basin might, however, impact upon the atmospheric circulation in such a way that this sort of anomalous freshwater flux can never be established. Thus, the reason why no deep water is formed in the North Pacific is probably not completely independent of the ocean geometry or bathymetry.

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