A Theory for the Surface Atlantic Response to Thermohaline Variability

HELEN L. JOHNSON AND DAVID P. MARSHALL

Department of Meteorology, University of Reading, Reading, United Kingdom

(Manuscript received 2 January 2001, in final form 17 July 2001)

ABSTRACT

The response of the upper, warm limb of the thermohaline circulation in the North Atlantic to a rapid change in deep-water formation at high latitudes is investigated using a reduced-gravity ocean model. Changes in deep-water formation rate initiate Kelvin waves that propagate along the western boundary to the equator on a timescale of months. The response in the North Atlantic is therefore rapid. The Southern Hemisphere response is much slower, limited by a mechanism here termed the ''equatorial buffer.'' Since to leading order the flow is in geostrophic balance, the pressure anomaly decreases in magnitude as the Kelvin wave moves equatorward, where the Coriolis parameter is lower. Together with the lack of sustained pressure gradients along the eastern boundary, this limits the size of the pressure field response in the Southern Hemisphere. Interior adjustment is by the westward propagation of Rossby waves, but only a small fraction of the change in thermohaline circulation strength is communicated across the equator to the South Atlantic at any one time, introducing a much longer timescale into the system.

A new quantitative theory is developed to explain this long-timescale adjustment. The theory relates the westward propagation of thermocline depth anomalies to the net meridional transport and leads to a ''delay equation'' in a single parameter—the thermocline depth on the eastern boundary—from which the time-varying circulation in the entire basin can be calculated. The theory agrees favorably with the numerical results. Implications for predictability, abrupt climate change, and the monitoring of thermohaline variability are discussed.

1. Introduction

The thermohaline overturning circulation transports approximately 1 PW of heat northward in the Atlantic basin (Macdonald and Wunsch 1996). When released to the atmosphere in the region of the North Atlantic jet stream, this heat intensifies the storm track and is responsible for warming western Europe by as much as 5°–10°C (Rahmstorf and Ganopolski 1999). The thermohaline circulation is global in scale, transporting approximately 20 Sv (Sv $\equiv 10^6$ m^3 s^{-1}) of warm water northward across the equator. It may therefore affect tropical sea surface temperatures to which the entire global atmosphere is particularly sensitive. Changes in thermohaline circulation strength could therefore have a significant impact on climate (Vellinga and Wood 2002, hereafter VW).

In principle we should be able to observe the strength of the meridional overturning circulation at high latitudes. For example, Curry et al. (1998) monitor Labrador Sea Water thickness as a proxy for deep-water formation rate, and Marotzke et al. (1999) propose using the basinwide zonal pressure gradient. A key challenge, given these observations, is to determine the response of the ocean at other latitudes, as well as the response of the atmosphere.

Previous work has largely focused on the steady-state thermohaline problem, concerning either the structure of the mean circulation [beginning with Stommel and Arons (1960)] or its multiple steady states (Stommel 1961). Kawase (1987) and Cane (1989) looked at the spinup problem and showed that the abyssal ocean responds to changes in the deep-water formation rate via the propagation of Kelvin waves around the North Atlantic, on timescales of less than a year. More recent work by Yang (1999) demonstrates that the surface ocean responds in a similar way. Marotzke and Klinger (2000) suggest, however, that the ocean responds much more slowly, with a timescale set by advection in the deep western boundary current. General circulation model (GCM) studies (e.g., McDermott 1996; Suginohara and Fukasawa 1988) identify Kelvin waves during the spinup of the thermohaline circulation but find that adjustment takes place over decades to centuries. The resolution of this issue is key. If we want to understand the impact on climate of a change in thermohaline circulation strength, the first step is to understand how the surface ocean itself responds.

The thermohaline circulation can influence the atmosphere only by rearrangement of the sea surface tem-

Corresponding author address: Helen L. Johnson, Department of Meteorology, University of Reading, P.O. Box 243, Reading RG6 6BB, United Kingdom.
E-mail: h.l.johnson@reading.ac.uk

perature (SST) field (or the sea ice distribution). Yang (1999) shows a correlation between the deep-water formation rate at high latitudes and tropical patterns of SST. Because of the deep convective boundary layer in the Tropics, the atmosphere here responds strongly to even small SST anomalies, so the potential for feedback on the atmosphere is large. Yang (1999) demonstrates the relevance of oceanic teleconnections to climate. [The importance of *atmospheric* teleconnections is well established: see, e.g., the review by Trenberth et al. (1998)]. We follow Yang's lead, looking at the teleconnections associated with the thermohaline circulation using a dynamical approach rather than the more conventional statistical one.

During the course of preparing this paper we have become aware of similar work carried out by Huang et al. (2000). They study the global adjustment of the thermocline in response to a change in the deep-water formation rate. Their emphasis, like that of the thermohaline literature in general, is very much on the millenial timescale solution to the problem. While they comment on the shorter term, transient response, they present no quantitative analysis. It is this decadal and shorter timescale response, however, that is crucial for the rapid climate change problem and is the focus of this study.

In this paper our aims are to

- describe the dynamical response of the upper, warm limb of the thermohaline circulation, on decadal and shorter timescales, to a change in deep-water formation rate at high latitudes;
- develop and test a quantitative theory of this response; and
- establish the importance of the equator in restricting abrupt changes in thermohaline overturning to the North Atlantic, through a mechanism that we term the "equatorial buffer."

To achieve these aims we use a highly idealized reduced-gravity model, similar to that of Huang et al. (2000). It represents only the upper, warm branch of the thermohaline circulation, which is likely to be most important for the atmospheric response on timescales shorter than a few years. The model's simple geometry allows us to obtain a detailed mechanistic understanding, which we expect to carry over in essence (if not in detail) to a continuously stratified ocean. Note that we do not attempt to model the deep-water formation process itself. We simply prescribe this at the northernmost edge of the domain and aim to establish the response at lower latitudes.

The paper will be structured as follows. The key concepts of the equatorial buffer are laid out briefly in section 2. Section 3 illustrates the dynamical response using a shallow-water model. In section 4 the theory for the longer timescale Rossby wave adjustment is developed. Results and their implications are discussed in section 5.

2. The equatorial buffer

The equator plays two important roles in the adjustment of the surface ocean to a change in deep-water formation. First, as noted by Kawase (1987), the equator acts as a barrier to the propagation of Kelvin waves on the western boundary. Here, however, we wish to introduce a second important mechanism, the equatorial buffer, which limits the magnitude of the response in the South Atlantic on short timescales.

Suppose the rate of deep-water formation in the high-latitude North Atlantic is suddenly perturbed:

- Mass continuity leads to a meridional velocity in the surface layers of the ocean on the western boundary, in geostrophic balance with an anomaly in pressure, which then propagates southward along the western boundary as a Kelvin wave at a speed c_g of order 1 m s^{-1}. This response is essentially the surface equivalent of that demonstrated by Kawase (1987) for the deep ocean.
- To leading order the anomaly is in geostrophic balance:

$$\mathbf{u} = \frac{1}{\rho f}\mathbf{k} \times \boldsymbol{\nabla} p, \qquad (1)$$

  where $\mathbf{u}$ is the velocity, ρ is the density, $f = 2\Omega \sin\phi$ is the Coriolis parameter, p is the pressure, and $\mathbf{k}$ is a unit vertical vector. Thus a smaller pressure gradient is required to support a given transport anomaly as latitude decreases. The pressure anomaly on the western boundary therefore reduces in amplitude as the signal moves equatorward and is small at the equator. Equatorial Kelvin waves transmit the resulting small thermocline displacement across the equator, and it then propagates both northward and southward along the eastern boundary as a coastal Kelvin wave. The amplitude of this signal is tiny compared to that on the western boundary because it has passed through the equatorial region.
- Along the *eastern* boundary a meridional pressure gradient cannot be sustained, since to balance it a Coriolis force would require a velocity across the coastline, and all other terms in the momentum equation are small. Pressure along the eastern boundary is therefore uniform on timescales longer than the transit time of a Kelvin wave. On the *western* boundary, where time dependence, inertia, and friction all become significant, a finite pressure gradient can be maintained.
- The thermocline displacement is communicated into the interior by the radiation of Rossby waves from the eastern boundary. Propagation speed is a strong function of latitude, so the thermocline displacement is spread over a large area, keeping its amplitude small. It is therefore the radiation of Rossby waves that prevents the reamplification of the pressure signal as the east coast Kelvin waves move poleward. Meridional transport south of the equator responds only as a result of adjustment by these Rossby waves.

Because of the restriction on zonal pressure gradients imposed by the equatorial region, the forcing communicated to the ocean interior via Rossby waves is constantly changing: we might think of the Tropics as a *buffer* zone that limits the size of the pressure field response (and hence the response in meridional transport) in the South Atlantic. This is analogous to a chemical buffer that prevents large changes in the concentration of a species. Only a small portion of the change in thermohaline circulation strength is communicated to the South Atlantic at any one time. The long adjustment timescale in the system may therefore be considerably longer than the Rossby wave transit time. A similar situation arises in delay-oscillator models of the El Niño system, where the period of the oscillation is longer than the propagation time of the individual equatorial waves (e.g., Suarez and Schopf 1988).

The implication, then, is that the North Atlantic responds very rapidly to changes in deep-water formation rate, but that the rest of the global ocean adjusts on much longer timescales. If deep-water formation were to shut down completely, the northward heat transport associated with the meridional overturning circulation would cease in a few months in the North Atlantic but would continue for several years south of the equator. This implies a convergence of heat in the Tropics, where the atmosphere is particularly sensitive to sea surface temperature anomalies.

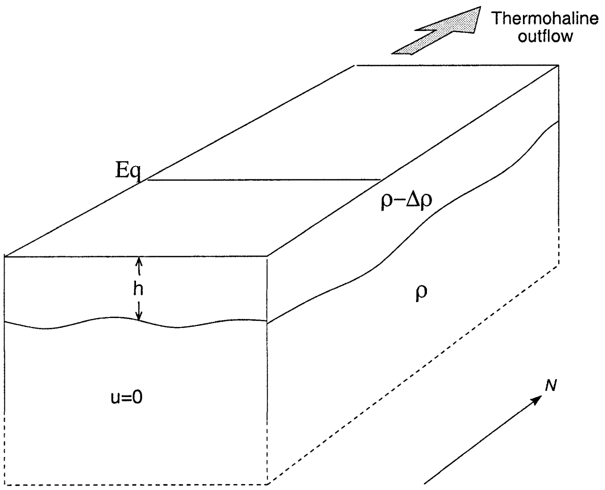

FIG. 1. Shallow-water model with moving surface layer and infinitely deep, motionless, lower layer. Thermohaline overturning is represented by a prescribed outflow from the surface layer on the northern boundary. Model domain is a sector ocean 50° wide, extending from 45°S to 65°N.

3. Shallow-water model

As a test bed for our ideas we use a simple reduced-gravity shallow-water model. Our aim is to elucidate the mechanisms of the ocean's response. The model is therefore chosen to be as simple as possible, while still representing at a reasonable resolution the dynamics we believe to be important.

a. Model formulation

The warm limb of the thermohaline circulation is represented by a shallow-water model with a moving surface layer (initial depth 500 m) and an infinitely deep, motionless lower layer (Fig. 1). Deep-water formation is represented by a prescribed outflow from the surface layer on the northern boundary (the sinking is assumed to take place outside the model domain). Experiments (not shown) reveal that our results are unaffected by the presence of mean wind-driven gyres, so in the spirit of keeping the model as simple as is reasonable we neglect wind-driven dynamics.

There has been much debate in the literature on the location of the upwelling required to balance the deep-water formation at high latitudes. Opinion is divided on whether it is distributed evenly throughout the global ocean [as assumed for convenience in the original Stommel and Arons (1960) model], is associated with enhanced mixing along ocean boundaries and rough topography (Polzin et al. 1997), or occurs mainly in the Southern Ocean (Döös and Webb 1994; Gnanadesikan 1999). Here we sidestep this issue and set the diapycnal upwelling to zero. We feel that this is reasonable, given that the Atlantic represents only a small fraction of the global ocean area, so different upwelling prescriptions will have a relatively small effect on the net meridional transport in the South Atlantic. In our model the upwelling required to balance the thermohaline outflow on the northern boundary is therefore represented using a "sponge" region to the south of the domain in which surface layer thickness is relaxed to a uniform value. The results are not sensitive to the details of this sponge, except in that the thickness value we relax to sets the surface layer thickness in the southwest corner of the domain.

Our restriction to a single basin is justified by the buffering effect of the Tropics. The response south of the equator is small, and we expect the response in other basins to be even smaller. The model domain is therefore an idealized sector ocean 50° wide and stretching from 45°S to 65°N. (Similar experiments have been conducted in an Atlantic domain with realistic coastlines and show essentially the same results. They will not be discussed here.)

The model equations are the standard nonlinear shallow-water equations:

$$\frac{\partial \mathbf{u}}{\partial t} + (f + \xi)\mathbf{k} \times \mathbf{u} + \boldsymbol{\nabla} B = A\nabla^2\mathbf{u} \qquad (2)$$

$$\frac{\partial h}{\partial t} + \boldsymbol{\nabla} \cdot (h\mathbf{u}) = 0, \qquad (3)$$

where $\xi = \partial v/\partial x - \partial u/\partial y$, $B = g'h + (u^2 + v^2)/2$, A is the lateral friction coefficient, $f = 2\Omega \sin\phi$, Ω is the rotation rate of the earth, ϕ is the latitude, $\mathbf{u}$ is the velocity, h is the surface layer thickness, ρ_0 is the average density, $\mathbf{x}$ is the horizontal position vector, $\mathbf{k}$ is a unit vertical vector, and t is time. The initial layer thickness h_0 is 500 m and the reduced gravity g' is 0.02 m s^{-2}. No normal flow and no slip boundary conditions are applied.

The equations are discretized on a C grid, with a resolution of 0.25°. On a C grid, the Kelvin wave speed is unaffected by the spatial resolution (Hsieh et al. 1983), although wave structure and damping are altered. Previous studies of the transient response to a change in thermohaline forcing have been run at far too coarse a resolution to resolve Kelvin waves. For example, Marotzke and Klinger (2000) conduct their experiments on a B grid at 3.75° by 4° resolution, and gravity wave speed is slowed considerably as a result. Here the midlatitude deformation radius $\sqrt{g'h}/f \sim 50$ km is resolved by approximately two grid points.

b. Response

Our interest lies in the rapid climate change problem: what would be the response on timescales of months to decades if the meridional overturning circulation were to suddenly weaken or shut down? In our model the dynamical response is well described by linear theory, so we consider here the cleaner problem of a sudden switch on of 10 Sv deep-water formation in an ocean initially at rest. The evolution of the system is followed after an instantaneous change from 0 to 10 Sv in the prescribed thermohaline outflow from the surface layer on the northern boundary.

Figure 2 shows a series of snapshots of the thickness field after a thermohaline outflow of 10 Sv is switched on at time $t = 0$ in the northwest corner of the domain. Similar to Kawase (1987), after the initial Kelvin wave signal on the western boundary reaches the Tropics ($\sim$1 month), further Kelvin waves can be clearly seen propagating along the equator and both northward and southward along the eastern boundary. The thickness signal is then carried westward into the basin interior by Rossby waves radiated from the eastern boundary. Equatorial waves are apparent after 3 and 4 months. Note that no signal continues southward into the Southern Hemisphere on the western boundary and that only a small fraction of the thickness signal at high latitudes reaches the eastern side of the basin (due to the buffering effect of the equator). Huang et al. (2000) and Yang (1999) find similar coastal and Rossby wave propagation timescales in their models. Marotzke and Klinger (2000) report a transient signal that follows the same pathways but takes much longer (45 years) to reach the equator.

More careful study of the model thickness field shows that, because information is transmitted so quickly along the equator and eastern boundary, surface layer thickness is approximately uniform along these Kelvin wave paths and shoals slowly as the interior adjusts (Fig. 3). This is in contrast to the western boundary where a strong meridional gradient persists throughout. Physically, the surface layer is shoaling because more water is being "sucked" into the western boundary current than is being returned from the interior, due to the delay involved in the propagation of information across the basin. Figure 4 shows a time series of the meridional transport as a function of latitude. Again the initial Kelvin wave can be seen and takes about one month to reach the equator. The equator's role as a buffer is then clear, with adjustment to equilibrium taking place over a period of about 50 years.

These results suggest a timescale separation approach, splitting the response into a very fast Kelvin wave response, which acts to make the pressure uniform along the equator and eastern boundary, and a much slower Rossby wave response, buffered at the equator, which propagates the eastern boundary pressure anomalies into the basin interior.

4. Theory for longer timescale adjustment

We now develop a theory for the longer timescale adjustment through the propagation of Rossby waves. As noted above, there is a large separation between the timescale of the initial Kelvin wave response (months) and the annual–decadal timescale of interest here. If we assume that the Kelvin waves are, in fact, *infinitely* fast so that the thickness of the surface layer along the equator and eastern boundary adjusts *instantly* to the signal arriving at the western side of the equator, then we can simplify the problem considerably.

a. Dynamical assumptions

We begin by listing the underlying assumptions. In the interior we assume geostrophic balance,

$$\mathbf{u} = \frac{g'}{f}\mathbf{k} \times \nabla h, \tag{4}$$

and a linearized continuity equation:

$$\frac{\partial h}{\partial t} + H\nabla \cdot \mathbf{u} = 0, \tag{5}$$

where H is the initial surface layer thickness. Other processes, such as nonlinearity and dissipation, are assumed to be important only in the western boundary current region where they are required in order to close the circulation. Dissipation is only of note to the theory in that it allows a meridional gradient in surface layer thickness to be maintained on the western boundary; as in Marotzke (1997), on the eastern boundary where h_e is uniform (see Fig. 3). The inclusion of wind forcing would lead to extra (Ekman) terms in these equations but would not alter the essential dynamics of the response.

The volume of the western boundary current region

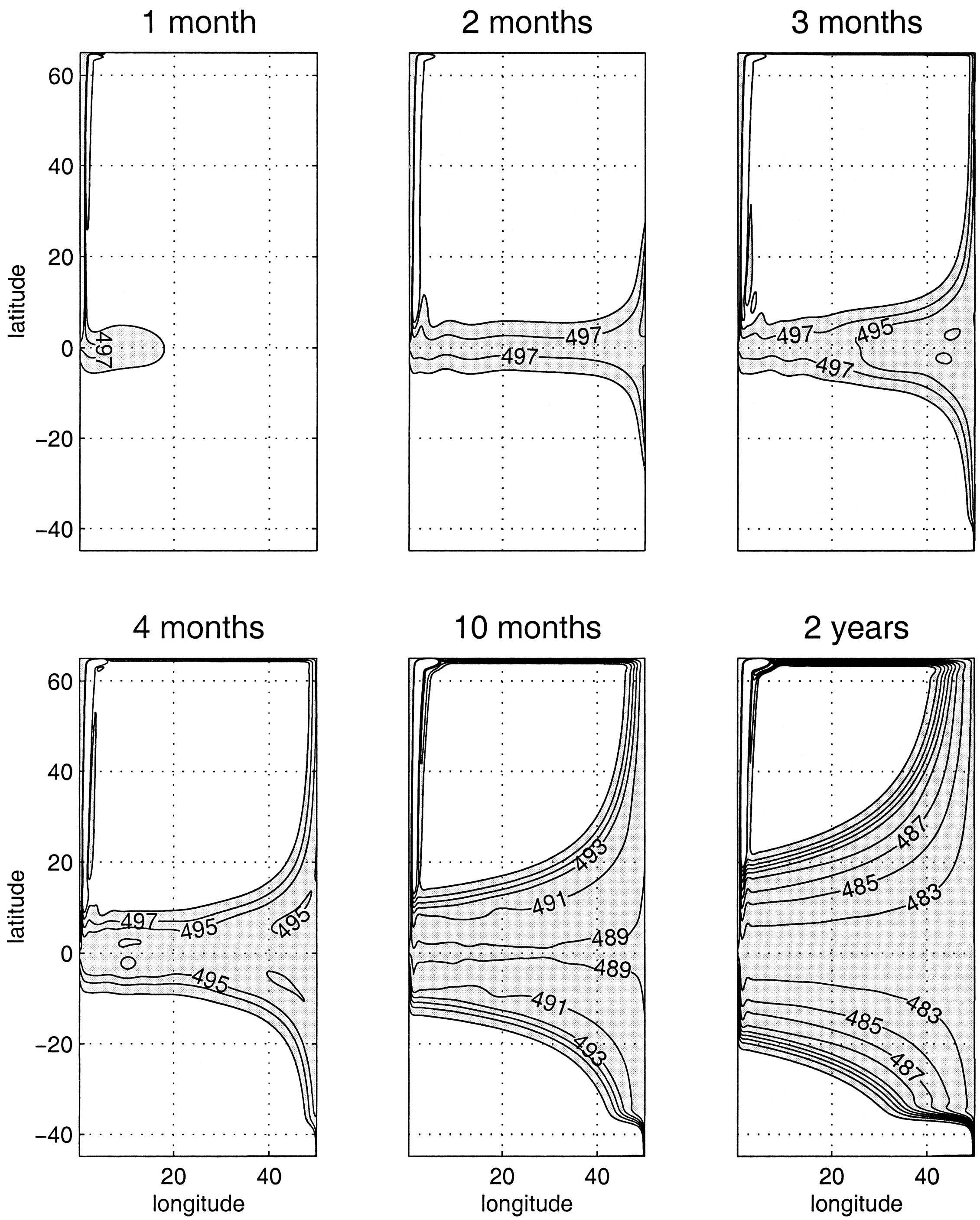

FIG. 2. Surface layer thickness after a thermohaline overturning of 10 Sv is switched on at time $t = 0$ in the northwest corner of an ocean initially at rest. There is no wind forcing, and the surface layer is initially 500 m deep. The contour interval is 2 m, and thicknesses less than 499 m are shaded. Note that the thickness anomaly on the western boundary is much greater than that in the interior (the layer thickness is approximately 350 m in the northwest corner of the domain; see Fig. 3 for details), but extra contours are not plotted here. The southernmost 10° of the domain comprises the sponge region, where we might expect the dynamics to be somewhat unrealistic.

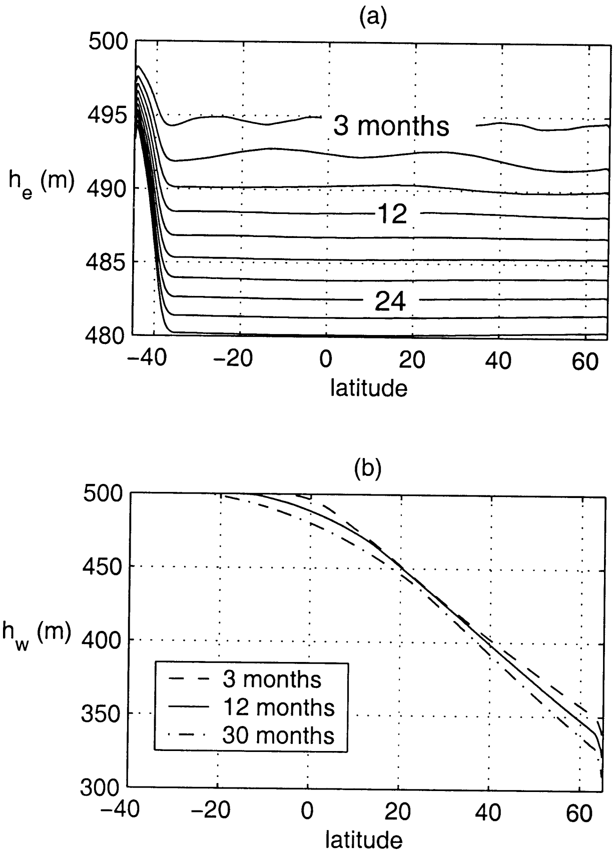

FIG. 3. Evolution of surface layer thickness on (a) eastern and (b) western boundaries. In (a) the lines show layer thickness at intervals of 3 months. Note that eastern boundary thickness is approximately uniform in y, and shallows with time, while there is a meridional gradient in western boundary layer thickness throughout. The southernmost 10° of the domain comprises the sponge region, where we might expect the dynamics to be somewhat unrealistic.

is considered to be negligible compared with the rest of the ocean, that is,

$$\int_w^b h\,dx \ll \int_w^e h\,dx, \tag{6}$$

where the integral limits w, b, and e represent the western boundary, the outside of the western boundary current, and the eastern boundary, respectively. For a boundary current of width 50–100 km in a basin 5000 km wide this introduces an error of less than 5%.

b. Meridional mass budget

The net northward volume transport T at any given latitude is defined as

$$T(y) = \int_w^e hv\,dx = \frac{g'}{2f}(h_e^2 - h_w(y)^2) \tag{7}$$

where h_e and h_w are the surface layer thickness on the eastern and western boundaries. We aim to predict this transport as a function of latitude and time.

Zonally integrating the continuity equation we find that

$$\frac{\partial}{\partial t}\int_w^e h\,dx = -\frac{\partial}{\partial y}T(y). \tag{8}$$

Physically, the divergence of transport depends upon the rate at which the zonally integrated layer thickness is shallowing in any given latitude band.

c. Rossby wave mass budget

Substituting (4) into (5) gives

$$\frac{\partial h}{\partial t} - c\frac{\partial h}{\partial x} = 0, \tag{9}$$

where c is the long baroclinic Rossby wave speed:

$$c = \frac{\beta g' H}{f^2}, \tag{10}$$

and $\beta = (2\Omega/R)\cos\phi$ is the gradient in the Coriolis parameter. The interior adjustment process is governed, then, by the propagation of long nondispersive baroclinic Rossby waves. Equation (9) only holds within the basin interior. In the western boundary current the importance of inertia means it is no longer valid. Zonally integrating (9) from just outside the western boundary current gives

$$\frac{\partial}{\partial t}\int_b^e h\,dx = c(h_e(t) - h_b(y, t)), \tag{11}$$

where h_b is the thickness of the surface layer immediately outside the western boundary current. The two terms on the right-hand side can be interpreted as volume fluxes associated with the westward propagating Rossby waves (the zonal component of the mass budget; see Fig. 5). The change in zonally integrated thickness at any given latitude is determined by the difference between the thickness anomaly communicated by Rossby waves into the western boundary current (h_b) and that propagated into the interior on the eastern side of the basin (h_e).

Because h_b is set by Rossby wave propagation from the eastern boundary, it is given by the history of h_e:

$$h_b = h_e\left(t - \frac{L}{c}\right), \tag{12}$$

where L is the width of the domain, and L/c is the time taken for a Rossby wave to cross the basin.

d. Equation for h_e

Since the volume of the boundary current is negligible, the right-hand sides of (8) and (11) can be equated. Substituting for h_b then gives

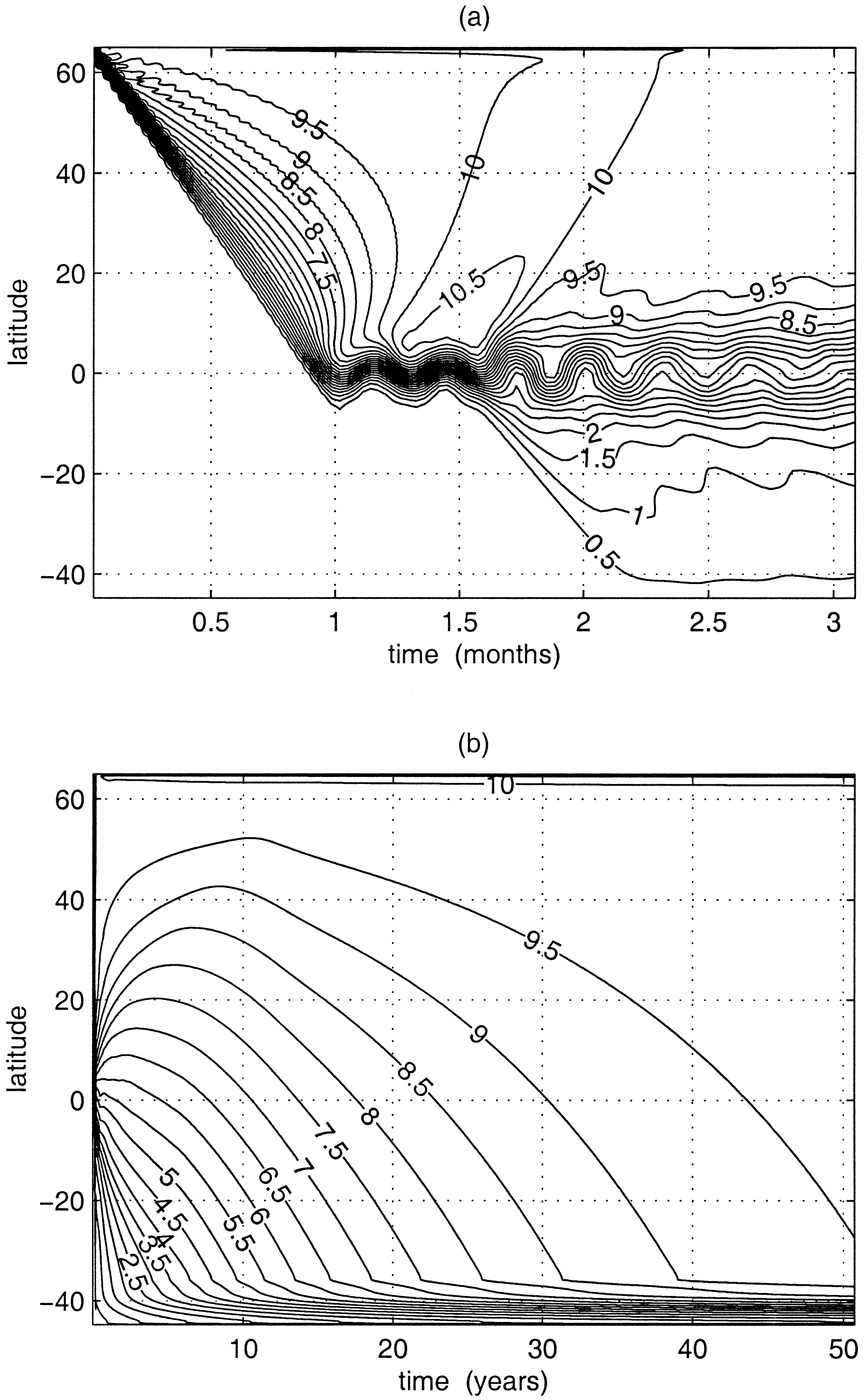

FIG. 4. Net northward volume transport (Sv) as a function of latitude and time after a thermohaline overturning of 10 Sv is switched on at time $t = 0$ in the northwest corner of an ocean initially at rest: (a) the initial Kelvin wave response and (b) the longer timescale adjustment. Note that the southernmost 10° of the domain comprises the sponge region, where we might expect the dynamics to be somewhat unrealistic.

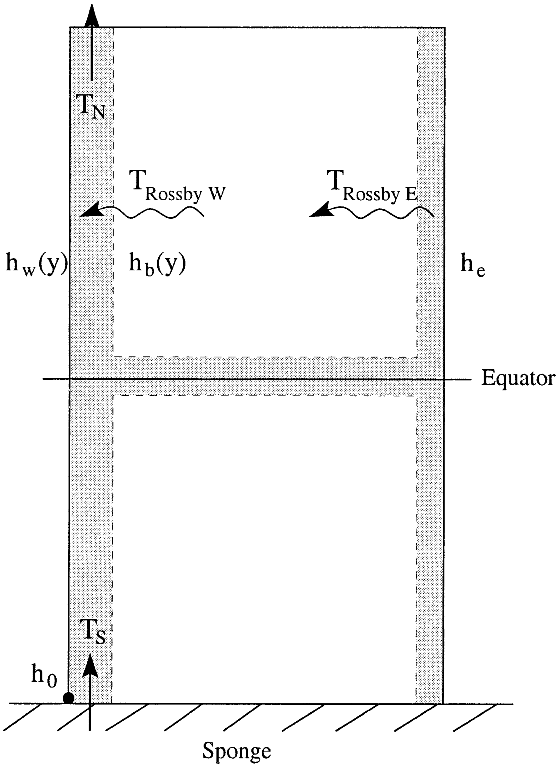

FIG. 5. Rossby wave mass budget. Westward propagating Rossby waves communicate a thickness anomaly of h_e into the interior from the eastern boundary and of h_b into the western boundary. The surface layer shoals with time as a result.

$$\frac{\partial}{\partial y}T(y, t) = c[h_e(t - L/c) - h_e(t)]. \tag{13}$$

The y dependence on the right-hand side enters through the lag since both the speed of long baroclinic Rossby waves and the width of the basin vary with latitude.

This gives us a delay equation from which, given the thickness of the layer on the eastern boundary, we can calculate the net transport across any latitude circle. All that remains is to establish how h_e varies as a function of time.

When (13) is integrated over the latitudinal extent of the domain we obtain

$$h_e(t) = \frac{1}{\int c\,dy}\left[\int h_e\left(t - \frac{L}{c}\right)c\,dy - T_N + T_S\right]. \tag{14}$$

Here T_S is given by the zonally integrated momentum balance at the southern boundary:

$$T_S = \frac{g'}{2f_S}(h_e^2 - h_0^2), \tag{15}$$

where h_0 is the prescribed layer thickness on the western boundary at the south of the domain (one might imagine that this is set by Kelvin waves propagating around Cape Horn from the Pacific); T_N is given by the imposed northern boundary condition. Hence only T_N, h_0, and the previous time history of h_e must be specified in order to obtain $h_e(t)$ and $T(y, t)$.

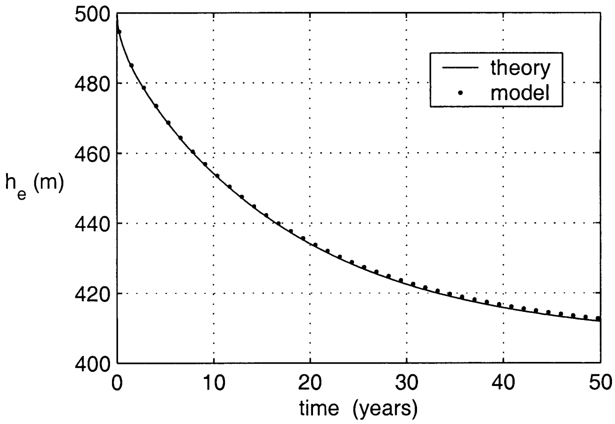

FIG. 6. Surface layer thickness on the eastern boundary. The solid line represents the theory, and the dotted line shows average eastern boundary layer thickness from the shallow-water model.

e. Comparison of model and theory

The delay equation is solved numerically to obtain a time series of h_e. To prevent the wave speed c from becoming infinite at the equator and to reflect the importance of equatorial dynamics in the real system, c is capped at the speed of the fastest equatorial Rossby waves. All westward propagating signals within a few degrees of the equator travel at one-third of the gravity wave speed, that is, at a speed $\sqrt{g'h}/3$. The eastern boundary thicknesses from the theoretical model and the full shallow-water model are shown in Fig. 6. The agreement between the two is remarkably good. Over the first 50 years the surface layer shoals by about 90 m.

If the southern boundary condition is linearized, then (14) has an exact analytical solution of the form

$$h_e(t) = \Delta h e^{-t/\tau} + h_{\mathrm{eqbm}}, \tag{16}$$

where Δh is the total change in eastern boundary thickness during the adjustment to equilibrium, and the adjustment timescale τ is given by

$$\tau = -\frac{f_S A}{g'H}, \tag{17}$$

where A is the area of the domain, f_S is the value of the Coriolis parameter at the southern boundary, and H is the average layer thickness. For the parameters used here we find $\tau = 16$ yr. Note that the adjustment time depends upon the latitude of the southern boundary of the domain. Shoaling of the surface layer continues until the zonal gradient in thickness at the southern boundary can support a meridional transport of 10 Sv. The farther this boundary is from the equator, the greater the change in h_e that is required to obtain a 10 Sv transport, in-

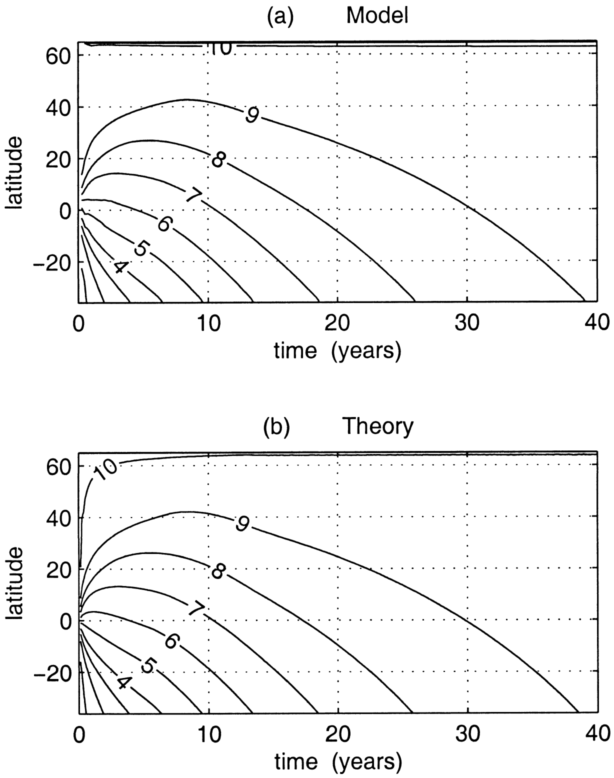

FIG. 7. Net northward volume transport (Sv) as a function of latitude and time from (a) the shallow-water model and (b) theory. The sponge region to the south of the shallow water domain is not shown.

creasing the adjustment timescale. However, if we set T_S equal to zero at time $t = 0$, then the rate at which the layer initially shallows is given by

$$\left.\frac{dh_e}{dt}\right|_{t=0} = -\frac{T_N}{A} \tag{18}$$

and is independent of the southern boundary condition. Details of the analytical solution are given in the appendix.

The structure of the circulation throughout the entire basin (both interior geostrophic circulation and boundary current strength) can be obtained relatively simply from the time series of h_e shown in Fig. 6. As an illustration, Fig. 7 shows the net meridional transport as a function of latitude and time. Figure 7a shows the results from the shallow-water model, and Fig. 7b shows the transport calculated using Eq. (13). The two are in broad agreement. This suggests that the dynamical adjustment of the surface ocean can be explained to a large degree by the simple theory developed in section 4.

5. Discussion

As in Kawase (1987), we have demonstrated that the ocean responds to a change in thermohaline forcing via the propagation of Kelvin waves along the western boundary to the equator. Upon reaching the equator the signal is deflected across the basin as an equatorial Kelvin wave and then travels in both directions along the eastern boundary, radiating Rossby waves that propagate westward into the interior. We have then continued where Kawase left off to demonstrate that, through the equatorial buffer mechanism, the equator acts to limit the size of the response in the South Atlantic, prevent high-frequency variability in thermohaline forcing from being transmitted to the Southern Hemisphere and introduce long adjustment timescales into the system. We have gone on to develop a new quantitative theory, based on a single variable—the thermocline thickness on the eastern boundary—to describe this surface ocean response.

Our simple, mechanistic, shallow water approach has many limitations. The thermocline is only very crudely represented by one active layer of uniform initial depth. The change in stratification with latitude in the real North Atlantic could substantially reduce the speed of coastal Kelvin waves at high latitudes (Greatbatch and Peterson 1996) and consequently lengthen the timescales involved. Our model also assumes vertical sidewalls and neglects all topographic influences on the surface dynamics. In reality, the signal is likely to be carried by some kind of hybrid coastally trapped wave propagating along the topographic shelf (Huthnance 1978) and interacting with overflows and other topographic features. We also neglect the effects of wind forcing and assume no background flow.

While the above limitations will undoubtedly modify the details of the response in a more realistic system, the theory relies on just a few fundamental ingredients that hold in general and not just in the simple model used here. As a consequence of geostrophy: (i) large pressure gradients are not possible in the Tropics, and (ii) pressure gradients cannot be maintained along eastern boundaries (where the existence of a Coriolis force to balance the pressure gradient force would require a flow through the boundary). These are the two key features of the equatorial buffer. The only other essential elements of the adjustment process are the existence of a rapid boundary wave that acts to communicate transport anomalies in the western boundary current from high to low latitudes, and a westward propagating Rossby wave mechanism in the interior. We therefore believe the essence of the ideas presented here to be robust.

The implications of our study are broad, but can be loosely grouped under three headings.

a. Heat transport

If the thermohaline circulation were to suddenly weaken significantly or shut down, then less heat would be transported northward by the surface ocean, and we might expect North Atlantic SSTs to cool. Meridional heat transport in the North Atlantic would cease very quickly, with heat transport south of the equator affected much more slowly. In our model, the ocean would adjust to this heat convergence by a deepening of the thermocline and an accumulation of heat in the ocean interior. Huang et al. (2000) suggest that this thermocline deepening may compensate for or even offset the negative SST anomalies associated with a reduced northward heat advection. For a 10 Sv change in thermohaline overturning, they find that the thermocline depth alters by between 50 and 100 m, with the largest changes occurring in the Tropics and along the eastern boundary of each basin. Surface layer thickness in our model shoals by 80–90 m during the 70 years over which adjustment takes place. To establish whether the dominant effect is likely to be a net warming of the upper ocean, or a large change in the amount of heat exchanged with the atmosphere, a detailed representation of the mixed layer and air–sea fluxes would be required. GCM experiments show a decrease in SST over most of the North Atlantic, but often show a much smaller warming over large parts of the Southern Hemisphere (e.g., VW), which could perhaps be explained by thermocline deepening.

b. *Predictability and monitoring*

For a known, single, instantaneous change in thermohaline forcing at high latitudes we have shown that, given a history of thermocline thickness on the eastern boundary, the surface ocean adjustment at every latitude can be predicted. If the thermohaline circulation were to suddenly shut down completely, the theory outlined in section 4 would therefore give a first-order estimate of the response over the next few decades. In reality, however, thermohaline forcing is unlikely to undergo a step change on less than decadal timescales. With a forcing that varies in time ($T_N \neq$ const), the rapid response in the North Atlantic means that the lead time for predictability is limited to the time it takes a Kelvin wave to propagate from the forcing region, along the western boundary, and then along the equator.

While this need to know the recent thermohaline forcing at high latitudes results in little real predictive power, our theory has much potential for thermohaline monitoring and nowcasting. Given a time series of surface layer thickness at any point on the eastern boundary, together with a measure of thermohaline circulation strength at high latitudes, the structure of the entire basinwide circulation can be recovered. From measurements made at just two points [e.g., dynamic height on eastern and western boundaries at a single latitude, as suggested by Marotzke et al. (1999)] monitoring of the overturning circulation strength throughout the entire basin could be achieved. Since we can compute the net volume flux across each latitude circle in the surface ocean (with an implicit return flow at depth), the theory suggests that measurements of hydrographic properties along the boundaries should provide strong constraints for inverse calculations.

c. *Modeling abrupt climate change*

The shallow-water model experiments conducted here are at a resolution of 0.25°, just enough to resolve Kelvin waves, which decay exponentially away from the boundary over an internal deformation radius. Like Hsieh et al. (1983) though, we find that the Kelvin wave speed is unaffected by changes in resolution on our C grid; for example, at as coarse a resolution as 4°, while the structure of the waves is somewhat unrealistic, the speed at which the pressure anomalies propagate is unaffected (not shown). This implies that we may not need to rely on very high resolution models for future studies of the abrupt climate change problem. The work of Hsieh et al. (1983) does suggest, however, that we should be careful in our interpretation of results from B-grid models, where resolution does become a factor in wave speed.

While the shallow-water model used here is fully nonlinear, its response to thermohaline variability is well described by our linear theory. This is a remarkable result and suggests that progress can be made on the rapid climate change issue using, for example, a planetary geostrophic model.

The recent work of Marotzke and Klinger (2000) on thermohaline adjustment has been something of a puzzle. In their idealized GCM experiments, adjustment takes place via a propagating signal that travels along the same Kelvin wave pathways but takes 45 years to reach the equator. The signal shows up clearly in the vertically integrated vertical velocity field. The equivalent diagnostic in our model is the Lagrangian rate of change of surface layer thickness (Dh/Dt), in which we see a rapid Kelvin wave signal reaching the equator after 1 month (not shown). The reason for the discrepancy between the two timescales is not clear. The Marotzke and Klinger (2000) model is run at coarse resolution on a B grid, with the Kelvin wave speed slowed considerably as a result, but not by enough to reconcile the differences. It is forced by a strong relaxation to a prescribed surface density field, unlike the ‘‘mass extraction’’ style of forcing used here, which could be a factor. Further work is needed to resolve this issue.

6. Concluding remarks

In order to take these ideas further, there is a need for simple process-based studies using GCMs to see if the theory carries over to a more complex stratification, geometry, and forcing structure. While several experiments have been conducted in which a salinity/freshwater anomaly has been imposed in the North Atlantic (e.g., VW), more analysis of the initial stages of these runs is needed in order to establish the importance of coastally trapped waves and the equatorial buffer. In terms of observations, the next logical step forward is to establish monitoring stations at the eastern and western boundaries of the Atlantic basin, as suggested by Marotzke et al. (1999).

This study is motivated in part by the potential for feedback on the atmosphere. Given some knowledge of changes in thermohaline overturning circulation strength, can we determine the response of the atmosphere? It would be interesting to pursue these ideas by coupling the kind of shallow-water model used here to a mixed layer model and a simple atmosphere.

Acknowledgments. We are grateful to Ted Johnson for suggesting the analytical solution outlined in the appendix, and to Joel Hirschi, Chris Hughes, Jochem Marotzke, Michael Vellinga, Carl Wunsch, and two anonymous reviewers for comments on a preliminary draft. This work was funded by the U.K. Natural Environment Research Council.

APPENDIX

Analytical Solution

If we assume an exponential solution of the form,

$$h_e(t) = \Delta h e^{-t/\tau} + h_{\mathrm{eqbm}}, \tag{A1}$$

and a linearized southern boundary condition,

$$T_S = \frac{g'H}{f_S}(h_e - h_0), \tag{A2}$$

where Δh is the total change in eastern boundary thickness during the adjustment to equilibrium, H is the average layer thickness, f_S is the value of the Coriolis parameter at the south of the domain, and h_0 is the initial layer thickness, then Eq. (14) becomes

$$A\left\{\int (1 - e^{L/c\tau})c\,dy - \frac{g'H}{f_S}\right\} = \left\{\frac{g'H}{f_S}(h_{\mathrm{eqbm}} - h_0) - T_N\right\}e^{t/\tau}, \tag{A3}$$

where the integral is calculated over the full latitudinal extent of the domain.

Equating terms that are exponential in time gives

$$h_{\mathrm{eqbm}} = \frac{T_N f_S}{g'H} + h_0 \tag{A4}$$

and sets the new equilibrium value of h_e.

Balancing steady terms gives

$$\int (1 - e^{L/c\tau})c\,dy = \frac{g'H}{f_S}. \tag{A5}$$

The exponent $L/c\tau$ is the ratio of the time taken for a Rossby wave to cross the basin (at a given latitude) to the adjustment timescale. Since propagation time ranges from months to years, while adjustment to equilibrium occurs over decades, it can be assumed to be small, and the exponential can be approximated by

$$e^{L/c\tau} \approx 1 + \frac{L}{c\tau}. \tag{A6}$$

This holds only marginally at the northernmost and southernmost extents of our domain, but works well in practice for the parameters used here. Substituting (A6) into (A5) leads to

$$\tau = -\frac{f_S A}{g'H}, \tag{A7}$$

where $A = \int L\,dy$ is simply the area of the domain. Note the sensitivity of the timescale to the southern boundary position. For the shallow-water model discussed in section 2 the parameters are $g' = 0.02$ m s^{-1}, $H = 500$ m, and $f_S = -10^{-4}$ s^{-1}, and the domain area is 5×10^{13} m^2, which gives an adjustment timescale of about 16 years.

If we set $T_S = 0$ initially, then the initial rate of adjustment is given by

$$\left.\frac{dh_e}{dt}\right|_{t=0} = -\frac{\Delta h}{\tau}. \tag{A8}$$

Since the initial value of h_e is h_0 and in the light of (A4),

$$\Delta h = -\frac{T_N f_S}{g'H}. \tag{A9}$$

So,

$$\left.\frac{dh_e}{dt}\right|_{t=0} = -\frac{T_N}{A}. \tag{A10}$$

For $T_N = 10$ Sv this is an initial shoaling of 6 m yr^{-1}.

It should be noted that this analytical solution is somewhat contrived in that it assumes an exponential decay in eastern boundary thickness prior to $t = 0$.

REFERENCES

Cane, M. A., 1989: A mathematical note on Kawase's study of the deep-ocean circulation. *J. Phys. Oceanogr.,* **19,** 548–550.

Curry, R. G., M. S. McCartney, and T. M. Joyce, 1998: Oceanic transport of subpolar climate signals to mid-depth subtropical waters. *Nature,* **391,** 575–577.

Döös, K., and D. J. Webb, 1994: The Deacon cell and other meridional cells of the southern ocean. *J. Phys. Oceanogr.,* **24,** 429–442.

Gnanadesikan, A., 1999: A simple predictive model for the structure of the oceanic pycnocline. *Science,* **283,** 2077–2079.

Greatbatch, R. J., and K. A. Peterson, 1996: Interdecadal variability and oceanic thermohaline adjustment. *J. Geophys. Res.,* **101** (C9), 20 467–20 482.

Hsieh, W. W., M. K. Davey, and R. C. Wajsowicz, 1983: The free Kelvin wave in finite-difference numerical models. *J. Phys. Oceanogr.,* **13,** 1383–1397.

Huang, R. X., M. A. Cane, N. Naik, and P. Goodman, 2000: Global adjustment of the thermocline in response to deepwater formation. *Geophys. Res. Lett.,* **27,** 759–762.

Huthnance, J. M., 1978: On coastal trapped waves: Analysis and numerical calculation by inverse iteration. *J. Phys. Oceanogr.,* **8,** 74–92.

Kawase, M., 1987: Establishment of deep ocean circulation driven by deep-water production. *J. Phys. Oceanogr.,* **17,** 2294–2317.

Macdonald, A. M., and C., Wunsch, 1996: An estimate of global ocean circulation and heat fluxes. *Nature*, **382,** 436–439.

Marotzke, J., 1997: Boundary mixing and the dynamics of three-dimensional thermohaline circulations. *J. Phys. Oceanogr.*, **27,** 1713–1728.

——, and B. A. Klinger, 2000: The dynamics of equatorially asymmetric thermohaline circulations. *J. Phys. Oceanogr.,* **30,** 955–970.

——, R. Giering, K. Q. Zhang, D. Stammer, C. Hill, and T. Lee, 1999: Construction of the MIT ocean general circulation model and application to Atlantic heat transport sensitivity. *J. Geophys. Res.,* **104** (C12), 29 529–29 547.

McDermott, D. A., 1996: The regulation of northern overturning by Southern Hemisphere winds. *J. Phys. Oceanogr.,* **26,** 1234–1255.

Polzin, K. L., J. M. Toole, G. R. Ledwell, and R. W. Schmitt, 1997: Spatial variability of turbulent mixing in the abyssal ocean. *Science,* **276,** 93–96.

Rahmstorf, S., and A. Ganopolski, 1999: Long-term global warming scenarios computed with an efficient coupled climate model. *Climatic Change,* **43,** 353–367.

Stommel, H. M., 1961: Thermohaline convection with two stable regimes of flow. *Tellus,* **13,** 224–230.

——, and A. B. Arons, 1960: On the abyssal circulation of the world ocean II. Idealized model of the circulation pattern and amplitude in oceanic basins. *Deep-Sea Res.,* **6,** 217–233.

Suarez, M. J., and P. S. Schopf, 1988: A delayed action oscillator for ENSO. *J. Atmos. Sci.,* **45,** 3283–3287.

Suginohara, N., and M. Fukasawa, 1988: Set-up of deep circulation in multi-level numerical models. *J. Oceanogr. Soc. Japan,* **44,** 315–336.

Trenberth, K. E., G. W. Branstator, D. Karoly, A. Kumar, N. C. Lau, and C. Ropelewski, 1998: Progress during TOGA in understanding and modeling global tele-connections associated with tropical sea surface temperatures. *J. Geophys. Res.,* **103** (C7), 14 291–14 324.

Vellinga, M., and R. A. Wood, 2002: Global climatic impacts of a collapse of the Atlantic thermohaline circulation. *Climatic Change,* in press.

Yang, J., 1999: A linkage between decadal climate variations in the Labrador Sea and the tropical Atlantic Ocean. *Geophys. Res. Lett.,* **26,** 1023–1026.