Dynamics of the Gulf Stream/Deep Western Boundary Current Crossover. Part II: Low-Frequency Internal Oscillations*

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(Manuscript received 3 March 1995, in final form 3 February 1996)

ABSTRACT

A low-frequency oscillation in the Gulf Stream/deep western boundary current (DWBC) system is identified and its influences on several important aspects of the basin-scale circulation are investigated. An eddy-resolving regional primitive equation model is used to demonstrate that feedbacks between the Gulf Stream, with its associated northern and southern recirculation gyres, and the upper core of the DWBC can lead to self-sustaining large amplitude internal oscillations of roughly decadal frequency. The oscillator cycle is described as follows: The upper core of the DWBC is entrained under the Gulf Stream through interaction with the eddy-driven northern and southern recirculation gyres, as described in Part I of this study. Once entrained, the low potential vorticity DWBC water stabilizes the Gulf Stream and suppresses the eddy fluxes that maintained the interior recirculation gyres. This causes the upper DWBC to switch to a southward path along the western boundary, thus removing the source of the stabilizing low potential vorticity water to the Gulf Stream. The Gulf Stream quickly returns to its unstable state and the resulting eddy fluxes spin up the northern and southern recirculation gyres. At this point, the upper DWBC is reentrained and the cycle begins again. The frequency and amplitude of the oscillations are controlled by the efficiency of the entrainment mechanism, as demonstrated by its sensitivity to variations in the model forcing parameters. The oscillation strongly influences the penetration scale of the Gulf Stream and distribution of eddy variability, the separation latitude of the Gulf Stream, the effective age of the DWBC south of the crossover, and the pathways of the upper DWBC. The implications of such an oscillation on observing and modeling the thermohaline circulation are discussed.

1. Introduction

The world's oceans play an important role in the global climate system by transporting heat toward high latitudes, where it is exchanged with the atmosphere and strongly influences world weather patterns. Much of this net poleward heat transport is carried in the thermohaline mode of the ocean, that part of the ocean circulation that flows poleward in the upper ocean and is compensated for by an equatorward flow in the deep ocean. The high latitudes of the North Atlantic are a major source region for the formation of intermediate and deep waters (Pickart 1992; Schmitz and McCartney 1993). While individual components of this overturning circulation, such as the Gulf Stream and the deep western boundary current (DWBC), are readily observed, there is still much uncertainty regarding the strength and time dependence of the overturning cell, the pathways that the waters follow (particularily the deep return flow), and the interaction between the poleward and equatorward flowing components. In Part I of this study (Spall 1996, hereafter referred to as Part I), it was shown that the eddy-driven interior recirculation gyres north and south of the Gulf Stream could entrain part of the upper core of the DWBC under the Gulf Stream, consistent with the observations of Pickart and Smethie (1993). This splitting of the DWBC has large consequences on the mean properties of both the upper DWBC and the basin-scale circulation in the interior. It will be shown here that this entrainment can lead to very interesting and important time-dependent behavior as well.

It is generally believed that the upper limb waters are converted to deep and intermediate waters in relatively small high-latitude regions through buoyancy exchange with the atmosphere. It is expected that variations in the strength of this interaction with the atmosphere (which may result from changes in the atmospheric weather patterns or in the ocean circulation) will then influence the rate of water-mass conversion and the strength of the overturning cell. Numerical models suggest that sufficient freshening of the surface waters at high latitudes can lead to a shutdown in the formation of deep waters, fundamental changes in the ocean circulation, and greatly diminished meridional heat transports (Bryan 1986). Low-frequency oscilla-

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tions in the state of the ocean, and its meridional heat transport and heat content, can result from such interruptions of the thermohaline overturning mode (e.g., see Weaver and Sarachik 1991 and the references therein).

Understandably, there is much interest in estimating and modeling the strength and time dependence of the thermohaline circulation. The DWBC provides a relatively confined current from which to observe the lower limb; however, the interpretation of such observations are complicated by the existence of interior recirculation gyres and time dependence. Transient tracers, such as chlorofluorocarbons (CFCs), may be used to obtain an estimate of the time since water parcels were last in contact with the atmosphere, and thus indirectly an estimate of the strength and time dependence of the thermohaline circulation (Fine 1995). However, as discussed by Pickart et al. (1989), Smethie (1993), and Fine (1995), the indirect estimate provided by CFCs is influenced by nonlinear mixing, which will generally result in estimated ages that are older than in the real ocean. The spreading of CFCs may also be influenced by exchanges between the boundary currents and interior recirculation gyres; thus CFCs are best viewed as providing an upper bound to the effective age of water parcels since their formation. The existence of recirculation gyres is evident in many deep ocean basins (Schmitz and McCartney 1993; Spall 1994); however, their exchange with the boundary currents and their influence on the properties of the water transported in the DWBC are not well understood.

Variability in the thermohaline circulation, and thus in the apparent age of the DWBC, can result from both internal (e.g., time-dependent exchange with interior recirculation gyres) and external processes. Therefore, if we are to correctly interpret any observed variability, and represent and predict such variability in models of the ocean and coupled ocean–atmosphere system, a clear understanding of the controlling physics is necessary.

The present study is an extension of Part I and investigates the influences of the entrained DWBC water on the stability properties of the Gulf Stream and its feedback on the entrainment of the DWBC water. In particular, a new mechanism for low-frequency oscillations in the thermohaline circulation is presented. The model assumption is that the strength of the overturning cell and atmospheric forcing are constant in time; thus all of the variability arises from the internal dynamics of the system. Rather than being characterized by changes in, or from, the strength of the overturning cell, the present oscillation is indicated by alternate pathways for the flow of the intermediate waters. It will be shown that this has very large consequences for the characteristics of both the upper and deep ocean circulations. The present study also addresses several issues that have received little attention to date, including the interaction between the DWBC and interior recirculation gyres (see also Spall 1994), the role of mesoscale variability in low-frequency (decadal) variability, and the midlatitude feedback between the upper and lower limbs of the thermohaline circulation.

The paper is outlined as follows: The model equations, configuration, and forcing are summarized in section 2. The basic oscillating mechanism is presented, and its dynamics described, in section 3. The sensitivity of the oscillation to variations in the forcing parameters is discussed in section 4, and discussion and final conclusions are presented in section 5.

2. Model configuration and forcing

The model used in this study is based on the regional Gulf Stream model described by Thompson and Schmitz (1989) and applied to a series of idealized, three-layer experiments in Part I of this study. The model solves the primitive equations of motion using an isopycnal vertical coordinate. Primitive equations are appropriate for the present study because both the variations in bottom topography near the DWBC and the isopycnal displacements near the Gulf Stream are large compared to the layer thicknesses. The isopycnal formulation also offers a conceptual advantage for process-oriented analysis in that the diapycnal mixing can be set to zero. This allows for near conservation of potential vorticity following fluid parcels and lends itself well to the study of flow pathways within distinct water mass layers using potential vorticity budgets. The equations and boundary conditions solved by the model are given in Part I.

The model domain is a rectangle 2500 km in latitudinal extent and 3500 km in longitudinal extent with a uniformly sloping coastline along the northwestern portion of the domain, representing the continent of the United States, as shown in Fig. 1. The horizontal grid resolution is 25 km in both the zonal and meridional directions. The Coriolis parameter varies linearly with latitude as $f = f_0 + \beta y$ with $\beta = 2 \times 10^{-11}\ \mathrm{m}^{-1}\ \mathrm{s}^{-1}$. The Coriolis parameter at the southern extent of the model domain is $f_0 = 0.616 \times 10^{-4}\ \mathrm{s}^{-1}$, consistent with

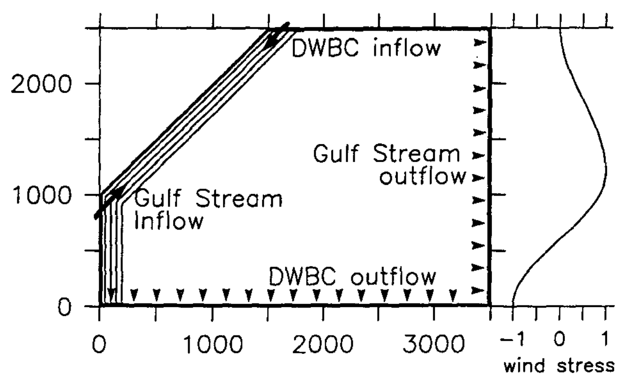

FIG. 1. Model domain and forcing configuration.

a latitude of 25°N. The maximum bottom depth is 5000 m and flat over the entire interior except for a 250-km-wide continental slope along the western boundary between 2500-m and 5000-m depth (uniform slope of 0.01 m/m). The Laplacian viscosity is 175 $m^2 s^{-1}$ and the bottom drag coefficient is 0.005 for all experiments.

The surface wind stress is applied in the zonal direction only and given by the analytic function

$$\tau_x = -\tau_m \cos(y\pi/L), \quad y < y_m \tag{1}$$

$$\tau_x = \tau_m(1 + \cos[(y - y_m)\pi/(L - y_m)])/2, \quad y > y_m. \tag{2}$$

The zero wind stress curl occurs at a latitude $y_m = 1200$ km and τ_m is the maximum value of the zonal wind stress. (The existence of the oscillation does not seem to be sensitive to the latitude of the zero wind stress curl.) This wind stress approximately represents the zonal average of the zonal component of the annual mean wind stress over the North Atlantic; the meridional distribution is indicated in Fig. 1 for $\tau_m = 1.0$ dyn cm^{-2}.

As described in Part I, the model is also forced through inflow/outflow boundary conditions at the edges of the model domain. An inflow in the upper layer along the western boundary centered at 925 km latitude of width 125 km represents that component of the Gulf Stream transport that flows off the shelf at Cape Hatteras. In the real ocean, this transport off the shelf is comprised of a wind-driven Sverdrup transport driven in the eastern North Atlantic (not resolved by the model domain) and Southern Hemisphere water that makes up part of the thermohaline circulation (Schmitz and McCartney 1993). No attempt is made to distinguish between these separate sources in the present study. The outflow boundary conditions in layer 1 require that the same amount of transport that enters through the western boundary must leave the domain somewhere through the eastern boundary (the entire eastern boundary is open). There is also an inflow 250 km wide specified in layers 2 and 3 along the northern boundary centered at 1725-km longitude over the sloping bottom. These inflows represent the upper and lower components of the DWBC that are formed at high latitudes. The DWBC waters are required to exit the model domain through the southern boundary such that total mass within each layer is conserved. For brevity, the value of the inflow boundary conditions will be specified as $\Gamma = (\Gamma_1, \Gamma_2, \Gamma_3)$, where Γ_k is the transport specified in layer k. In order to isolate the internal modes of variability from external processes, it is assumed here that the inflow conditions and surface forcing are held steady. This restriction explicitly excludes the possibility of feedbacks between the state of the ocean at midlatitudes and the water mass conversions at high latitudes.

Lack of explicit thermodynamics requires that the stratification be specified rather than be determined by the model physics and atmospheric forcing. This simplified adiabatic approach has several advantages from the process-oriented modeling point of view. The spin-up time is greatly reduced, thus allowing for a larger exploration of parameter space for a given computational expense. More importantly, the strength of the thermohaline circulation, as specified by the individual components of the DWBC, is under the explicit control of the user rather than being dependent on complex and poorly understood processes such as air–sea exchange, vertical convection, lateral spreading, and flow over complex topography. It is reemphasized that the present focus is on midlatitude processes; no attempt is made to model the entire cycle of the thermohaline circulation. The potential density of each of the layers is 26.25, 27.75, and 27.88, with respective resting layer thicknesses of 850 m, 900 m, and 3250 m. This stratification has been chosen to be representative of the observed stratification in the North Atlantic, within the constraints of a three-layer model and the requirement that the layer thicknesses not vanish during the course of integration. The first internal deformation radius at the central latitude of the model domain is 41 km.

3. Model results

The model is initialized at rest and integrated for a period of 100 years for each of the calculations in this paper. Because the model is adiabatic, the spinup time is relatively short, on the order of 10 years. The mean circulations resulting from the 100-year integrations are similar to those described in Part I, including a splitting of the upper DWBC into two paths at the crossover. The present focus is on the model variability, for a discussion of the mean circulation patterns the reader is referred to Part I.

a. An internal oscillator

The interaction between the upper DWBC and the Gulf Stream can lead to large amplitude, low-frequency oscillations in the basin-scale current system. One simple measure of the frequency and amplitude of such oscillations is indicated by time series of the basin integrated kinetic energy (total, not eddy energy) in each layer. The time series shown in Fig. 2a is for a calculation with transports $\Gamma = (50, 10, 10)$ Sv (Sv $\equiv 10^6$ $m^3 s^{-1}$) and a maximum surface wind stress $\tau_m = 0.5$ dyn cm^{-2}. These transport values are reasonably close to the observational estimates of Schmitz and McCartney (1993), and Pickart (1992). The kinetic energy in the layers corresponding to the Gulf Stream and the upper DWBC fluctuates by a factor of approximately 2 over the 100 years of integration. The events occur at nearly the same time in both layers, with a much smaller, but still noticeable, signal in the deepest layer. The frequency of the events varies, but is on average approximately once every 10 years (nine events in 100

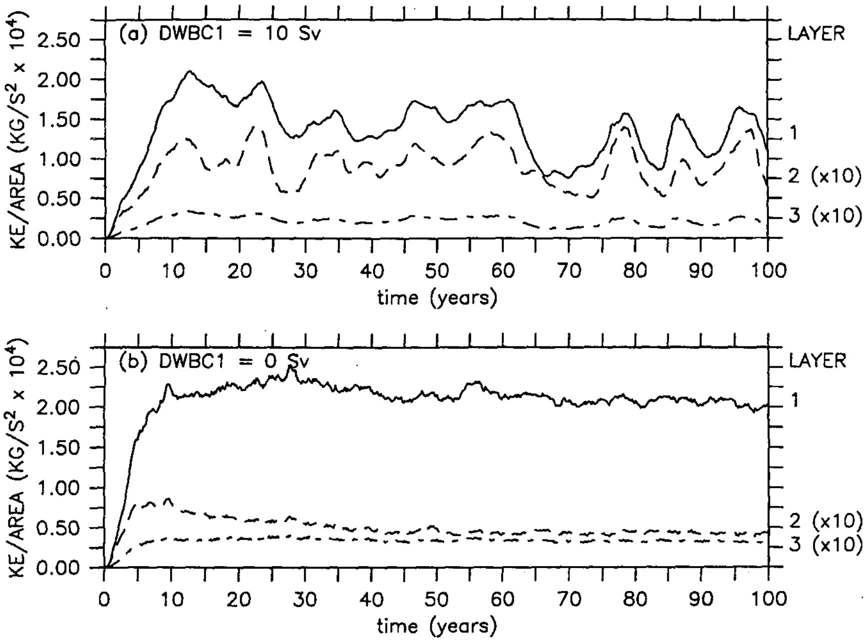

FIG. 2. Time series of basin integrated kinetic energy in all three model layers (layers 2 and 3 have been multiplied by a factor of 10 for clarity). (a) Central model run with transports Γ = (50, 10, 10) and (b) model run with no upper DWBC, transports Γ = (50, 0, 10).

years). The model fields are oscillating between high and low energy states. This oscillation is due to the presence of the upper DWBC, as is demonstrated by the time series shown in Fig. 2b for a calculation with the same wind stress and no transport in the upper DWBC, Γ = (50, 0, 10) Sv. [To be completely consistent with the reduction in layer 2 transport, one might also choose to either 1) reduce the layer 1 transport by 10 Sv or 2) increase the layer 3 transport by 10 Sv. The present forcing was chosen to isolate the upper DWBC as the key factor in the oscillation.] There is no evidence of the low-frequency oscillation in the absence of the upper DWBC, the energy spins up over the first 10 years of integration and remains nearly constant for the remaining 90 years.

The nature of the high and low energy states is revealed by the average of the transport streamfunction taken for the high energy events and the low energy events separately. The average transport streamfunction for the high energy events (years 11, 17, 23, 32, 47, 57, 78, 89, and 96) is shown in Fig. 3a for layer 1. The Gulf Stream is evident penetrating approximately 2000 km into the basin, flanked to the north and south by eddy-driven recirculation gyres. The mean of the low energy states (years 13, 20, 26, 37, 52, 70, 83, 92, and 100) exhibits a much shorter penetration of the Gulf Stream into the interior and much weaker recirculation gyres (Fig. 3b). The Gulf Stream only penetrates approximately 1000 km into the basin, half of what was found for the high energy years. The Gulf Stream is also found slightly farther to the south in the low energy states (more on this in section 3c). The oscillation in total kinetic energy results from the model Gulf Stream switching back and forth between a low energy state with weak recirculation gyres and short penetration scale and a high energy state with stronger recirculation gyres and a long penetration into the basin. This transition between states takes place in only 2–5 years. Similar low energy and high energy limit states, and patterns of variability, have been found in a single-layer quasigeostrophic wind-driven model by McCalpin and Haidvogel (1995). Although their extreme states are qualitatively similar to those found here, their model behavior appears to be more chaotic and sensitive to model forcing. It is interesting that a model with very different physics and forcing demonstrates a tendency to visit similar extreme states.

The intermediate depth energy and circulation also show large fluctuations between the high and low energy states. The high energy states are characterized by strong, zonally elongated recirculation gyres to the north and south of the upper layer Gulf Stream and a complete entrainment of the upper DWBC into the recirculation gyres, as shown in Fig. 4a. The entrained DWBC water flows under the Gulf Stream and into the southern recirculation gyre before returning to the western boundary south of the crossover. There is no Gulf Stream transport leaving the shelf in this density range. The entrainment of the upper DWBC below the Gulf Stream was shown in Part I to be driven by the

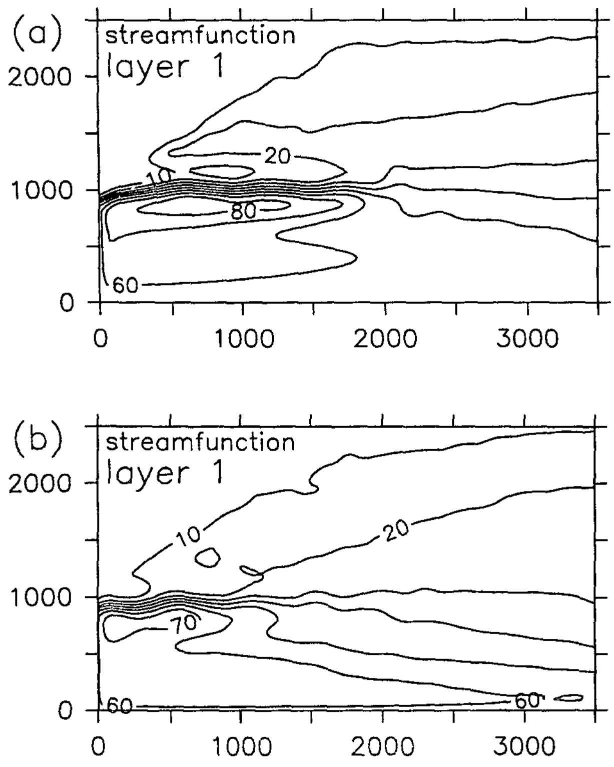

FIG. 3. Average layer 1 streamfunction (Sv ≡ 10^6 m^3 s^{-1}) for (a) high energy states and (b) low energy states.

eddy flux of potential vorticity at the crossover. The low energy states (Fig. 4b) have much weaker recirculation gyres and only partial entrainment of the upper DWBC. This splitting of the upper DWBC into an entrained branch flowing under the Gulf Stream and another branch flowing directly to the south was found in Part I and has also been inferred from observations by Pickart and Smethie (1993).

The relationship between the amount of entrained upper DWBC water and the energy level of the Gulf Stream is apparent by considering the time series of the amount of upper DWBC entrained at the crossover (Fig. 5). The entrainment was calculated once for each year from the annual average transport streamfunction as $\Gamma_2 - \Gamma_2'$, where Γ_2' is the net southward transport between the center of the southern recirculation gyre and the western boundary. There is a large entrainment of upper DWBC water under the Gulf Stream leading each of the high energy states seen in Fig. 2a, and a low level of entrainment (more of the upper DWBC flowing directly to the south at the crossover) preceding each of the low energy states. When the entrainment exceeds 10 Sv (the upper DWBC transport), it indicates that the flow along the western boundary is to the north. The entrainment/detrainment of upper DWBC leads each of the low-frequency fluctuations in basin-integrated energy by several years. This phase relationship suggests that the switching of the upper DWBC between a path that turns to the east (flowing under the Gulf Stream) and one that flows directly to the south (crossing under the Gulf Stream) triggers the switching of the Gulf Stream between the high energy and low energy states. It will be shown in the following section how the entrainment/detrainment of the upper DWBC results in fundamental changes in the state of the Gulf Stream and leads to a self-sustaining low-frequency oscillation.

b. The oscillator mechanism

The life cycle of the oscillator is most easily understood if one first considers the low energy state, with most of the upper DWBC flowing to the south crossing under the Gulf Stream. As discussed in Part I, the baroclinically unstable Gulf Stream spins up eddy-driven recirculation gyres to the north and south of the stream that extend westward toward the DWBC. Through the combined influence of these recirculation gyres and the eastward shifting of the upper DWBC at the crossover, the DWBC sheds low potential vorticity eddies into the interior, which are then carried to the east by the intermediate depth Gulf Stream. The divergence of the eddy

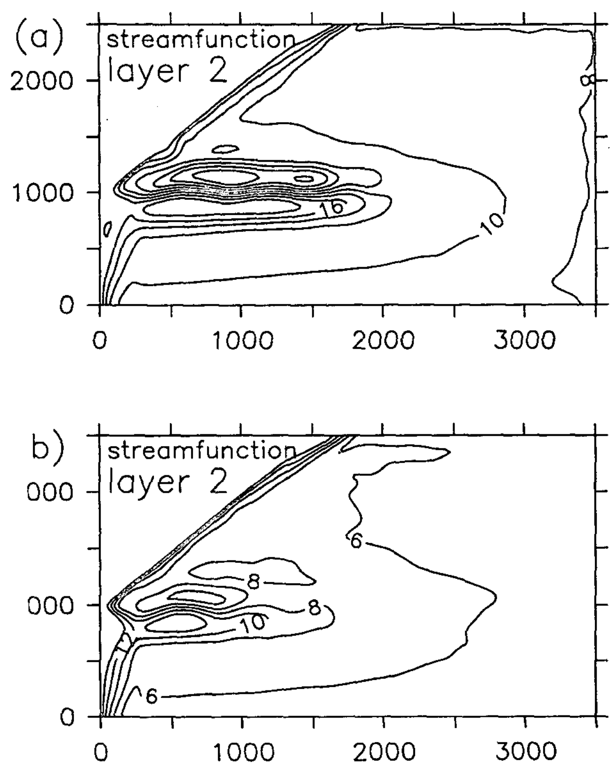

FIG. 4. Average layer 2 streamfunction (Sv) for (a) high energy states and (b) low energy states.

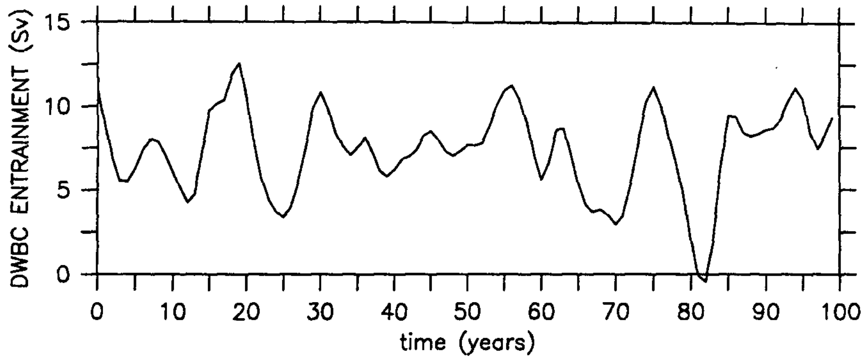

FIG. 5. Time-series of upper DWBC water entrained under the Gulf Stream (Sv).

potential vorticity flux resulting from these events drives a mean flow from the western boundary current at the crossover under the Gulf Stream toward the east. This mechanism splits the mean path of the upper DWBC into two branches, one flowing to the south and one flowing to the east. If the entrainment is strong enough, all of the upper DWBC water may be diverted to the east under the Gulf Stream, as is the case preceding each of the major oscillation events as indicated in Fig. 5.

The entrainment of the upper DWBC water leads to an accumulation of low potential vorticity in the southern gyre. One such example of this state is shown in Fig. 6 for fields in the vicinity of the crossover, averaged over model year 78. The streamfunction clearly shows the existence of the two recirculation gyres and the turning of the upper DWBC at the crossover under the Gulf Stream. The potential vorticity shows a tongue of low potential vorticity extending off the boundary at the crossover (1000-km latitude), which then splits into two branches, one extending east and one extending south. The eastward branch flows along the southern edge of the Gulf Stream as it recirculates in the southern gyre. This accumulation of low potential vorticity in the southern gyre results in a positive meridional gradient of the potential vorticity under the core of the Gulf Stream. In the low energy state (not shown) the potential vorticity under the Gulf Stream is nearly homogenized but has a slightly negative meridional gradient (higher potential vorticity to the south of the stream, low potential vorticity to the north). This state is consistent with the near homogenization of potential vorticity in recirculation gyres in the absence of any DWBC influence predicted by the theory of Rhines and Young (1982) and found in the two-layer steady quasigeostrophic model of Cessi (1988) and the three-layer eddy-resolving quasigeostrophic models of Holland et al. (1984).

A necessary condition for baroclinic instability in a layered model is that the meridional gradient of potential vorticity be of different sign in two of the model layers. This is the discrete representation of the familiar condition for continuously stratified models that the meridional gradient of the potential vorticity be zero somewhere in the basin (Pedlosky 1979). The meridional gradient in layer 1 across the Gulf Stream is always positive (low q, large layer thickness to the south). In the absence of DWBC influences, the meridional gradient in layer 2 is negative (high q, thinner layer thickness to the south), thus satisfying the necessary conditions for instability. The influence of mixing due to mesoscale eddies tends to homogenize the q in layer 2, but the exactly homogeneous limit is never quite achieved. The low potential vorticity diverted from the DWBC can lead to a local stabilization of the Gulf Stream in the western 2000 km of the basin by carrying low q water into the southern recirculation

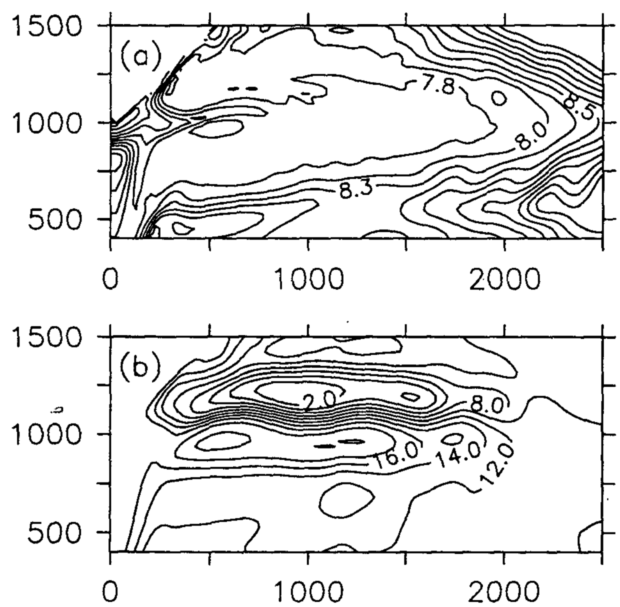

FIG. 6. Example of average layer 2 (a) potential vorticity and (b) streamfunction during the high energy states (shown here for model year 78). Note complete entrainment of upper DWBC and accumulation of low potential vorticity south of the Gulf Stream.

gyre and changing the sign of the meridional gradient in layer 2. A time series of the meridional gradient in layer 2 under the core of the upper-layer Gulf Stream, averaged between 250- and 1250-km longitude, is shown in Fig. 7 (solid line). There are distinct positive peaks in the meridional gradient corresponding to the high energy states in Fig. 2a. This demonstrates that the high energy states are also relatively stable states as a result of the entrainment of upper DWBC water. Each of the low energy states has a reduced meridional gradient in the potential vorticity, suggesting a decrease in stability. A local maximum in the meridional gradient of potential vorticity (slightly positive) is found under the Gulf Stream in the density range of the upper DWBC by Johns (1988), suggesting such a similar influence may be present in the real ocean.

The consequence of the change in meridional gradient of potential vorticity in layer 2 on the strength of the eddy fluxes is evident in the time series of the zonally averaged annual meridional eddy flux of potential vorticity ($\overline{v'q'}$) in Fig. 7 (short-dashed line). The eddy flux is nearly 180° out of phase with the meridional gradient of q. Positive eddy fluxes indicate that the upper ocean is unstable and that the eddies are working to pump zonal momentum downward from the upper layer to the intermediate layer and driving the recirculation gyres in layer 2 (Rhines and Holland 1979). The time periods of weak meridional gradients in q correspond to relatively unstable states with positive eddy fluxes. The increase in the meridional gradient of q stabilizes the upper Gulf Stream and the eddy fluxes quickly become small or negative (negative fluxes correspond to a deceleration of the layer 2 zonal flow). Since this is a zonally averaged measure of the meridional gradient, even in times when the average is positive, there may be regions where it is negative and the necessary conditions for instability are still met. The high basin-integrated energy results directly from the increased stability of the stream because it is able to penetrate far into the basin before breaking up due to instabilities. The correspondence between the meridional gradient of q and the strength of the deep Gulf Stream (and recirculation gyres) is indicated in Fig. 7 by the zonally averaged strength of the maximum zonal flow (long-dashed line).

The life cycle of the oscillator may be summarized as follows: When the upper DWBC is flowing to the south, the Gulf Stream is in a relatively unstable state because the meridional gradient in potential vorticity in layer 2 is negative under the Gulf Stream. The unstable stream pumps momentum downward from the upper layer into the intermediate and deep layers and spins up the northern and southern recirculation gyres. The recirculation gyres drive the entrainment of upper DWBC water at the crossover through the eddy flux mechanism discussed in Part I. A sufficiently strong diversion of the upper DWBC advects a plume of low potential vorticity water along the southern edge of the Gulf Stream and into the southern recirculation gyre. The Gulf Stream is stabilized by the presence of the low q water and the eddy fluxes diminish. In the absence of a continuous momentum supply by the eddy fluxes, the deep Gulf Stream and recirculation gyres spin down through diffusion and negative eddy fluxes on a timescale of several years. Once the recirculation gyres are gone, the entrainment mechanism that forced the diversion of the upper DWBC is no longer effective and the DWBC switches back to its southward flowing path. With the supply of DWBC water turned off, the stabilizing plume of low q water is rapidly mixed away and the Gulf Stream returns to its unstable state. This cycle then starts again.

It is stressed here that this is not a random or chaotic oscillator, but rather it is controlled by a specific, repeatable cycle that has associated with it a characteristic period determined by the efficiency of the DWBC entrainment and subsequent Gulf Stream stabilization. This will be discussed further in section 4.

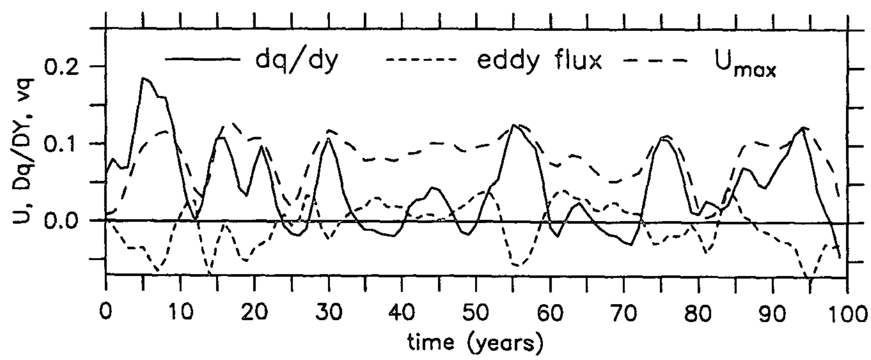

FIG. 7. Various terms under the core of the Gulf Stream, zonally averaged between longitude 250 km and 1250 km. Solid line: meridional gradient in potential vorticity (0.25×10^{-4} m^{-1} s^{-1}). Short dashed line: meridional eddy flux of potential vorticity $\overline{v'q'}$ (10^{-9} s^{-2}). Long dashed line: zonal velocity (m s^{-1}).

c. Consequences of the oscillation

The oscillation impacts many aspects of the circulation in addition to the basin-integrated energy. The consequences of the oscillation for three important properties of the large-scale circulation are presented here. In brief, variability in the entrainment of upper DWBC water causes oscillations in 1) the region of maximum eddy kinetic energy in the Gulf Stream, 2) the separation latitude of the Gulf Stream, and 3) the age of the upper DWBC water downstream (south) of the crossover.

1) EDDY KINETIC ENERGY

The eddy kinetic energy in layer 1 is shown in Figs. 8a and 8b for the high energy and low energy years, respectively. The eddy kinetic energy was calculated for each model year based on deviations from the mean flow for that model year. Figure 8 represents the averages of each of the high and low energy states over the 100-year integration (9 individual years for each state). The high energy state has very low variability in the western 1000 km of the basin, with the maximum at 1500 km longitude and penetration out to 2200 km longitude. The low energy state has a maximum value at approximately 700-km longitude and does not penetrate past 1500-km longitude. These patterns are consistent with the previous analysis indicating that the low energy states are quite unstable, and the high energy states are stabilized by the plume of DWBC in the west. The maximum value of the eddy kinetic energy is nearly the same in both states, approximately 250 $cm^2 s^{-2}$. These values are low relative to typical surface observations (Richardson 1983), but it must be remembered that they are averages over the whole main thermocline, not just the surface value, and also they were calculated from individual time series of 1 year, thus filtering out lower-frequency variability. The most important aspect of this result is the change in the pattern of variability, not the relative amplitude of the eddy energy.

A similar shift in the distribution of eddy kinetic energy is found for the intermediate and deep layers. This is similar to the shift in variability found in the two-layer model of Thompson and Schmitz (1989) when they added an increasingly strong DWBC in the deepest layer. This suggests that a diversion of the deep DWBC at the crossover through the present mechanism stabilized the western portion of their domain and resulted in the observed shift in eddy energy to the east. Their model was forced with more realistic winds and was configured in a model domain with the realistic bottom topography and coastlines.

The pattern of the upper ocean eddy kinetic energy commonly observed in realistic wind and buoyancy forced North Atlantic basin models, such as the Community Modelling Effort discussed by Stammer and Böning (1992), has a maximum very near to the coast. This is in contrast to the observed surface variability as inferred from satellite altimeter data, which shows a maximum approximately 1000 km offshore of the separation point (Stammer and Böning 1992). The present results suggest that an inadequate representation of the upper DWBC, or insufficient entrainment of this low potential vorticity water at the crossover point, could be responsible for the lack of Gulf Stream penetration in the model. This result would also be consistent with the separation point in the CME model, which is too far to the north, because the upper DWBC shifts the separation point to the south, as discussed further below (see also Thompson and Schmitz 1989 and Part I).

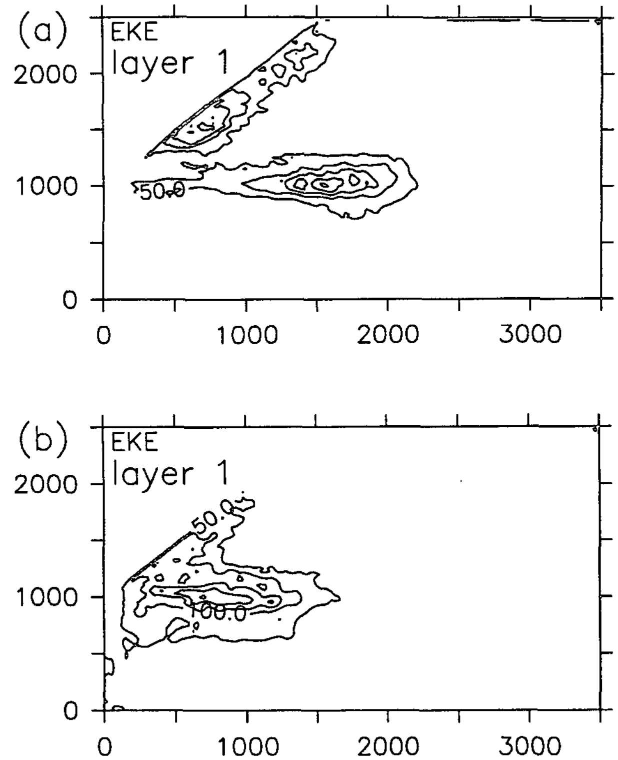

FIG. 8. Average eddy kinetic energy ($cm^2 s^{-2}$) in layer 1 for (a) high energy states and (b) low energy states (contour interval = 50 $cm^2 s^{-2}$).

2) GULF STREAM SEPARATION

The separation latitude of the Gulf Stream is strongly influenced by the entrainment and detrainment of the upper DWBC. The layer 1 zonal velocity representative of the annual average between longitudes 250 km and 1250 km is shown in Fig. 9 as a function of time and latitude. The positive core with maximum velocities of approximately 80 cm s^{-1} marks the position of the Gulf Stream, which flows nearly zonally at these longitudes. The recirculation gyres are indicated by westward

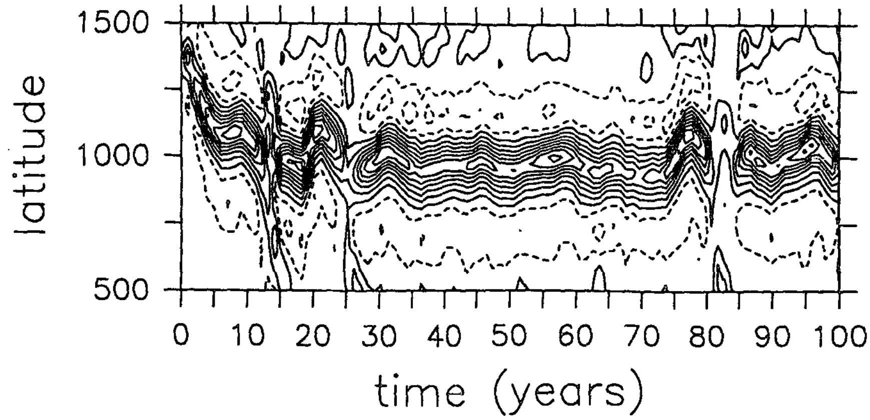

FIG. 9. Average layer 1 zonal velocity between longitude 500 and 1500 km as a function of latitude and time. High energy states are indicated by an increase in the maximum zonal velocity and northward shift in the core of the Gulf Stream; contour interval = 10 cm s^{-1} and first contours are at ±5 cm s^{-1}.

flows of approximately 10–15 cm s^{-1}. Each of the major oscillation events is seen in this figure. The transition to the high energy states is indicated by an increase in the amplitude of the maximum zonal flow and, to a lesser extent, in the strength of the recirculation gyres; this is the spinup driven by the eddy fluxes. Also remarkably coincident with each of these events is a northward shift of the core of the Gulf Stream and recirculation gyres. This shift can be as large as 200 km, but is more often on the order of 100 km. Calculations with weaker wind forcing result in larger amplitude shifts to the north that average 250 km. This northward shift in the core of the stream is caused by a northward shift in the separation point, not by the stream turning to the north after it has separated from the coast. The drop in penetration scale of the stream upon transition to the low energy states is also indicated by the decrease in the maximum zonal velocity after the peak events. Some of these events are so drastic that the Gulf Stream essentially disappears between 250 km and 1250 km, for example, years 27 and 83. This transition to the low energy state is also accompanied by a southward shift in the stream position.

The reason the oscillation influences the separation point so strongly is clear if we consider the mechanism of interaction between the upper DWBC and the Gulf Stream, as discussed by TS89 and Part I. The southward flowing DWBC causes the Gulf Stream to separate from the coast farther to the south than it otherwise would in the absence of the DWBC by advecting the sloping thermocline to the south. This mechanism is effective as long as the DWBC is flowing nearly perpendicular to the upper-layer stream, as is the case when there is little entrainment under the Gulf Stream. However, as the DWBC is diverted offshore, its influence on the Gulf Stream is decreased because it flows more nearly parallel to the stream at the crossover. When all of the upper DWBC is being diverted to the east, the component that is advecting the thermocline southward ($\mathbf{v}_2 \cdot \nabla h_1$) is greatly reduced. The Gulf Stream thus flows farther to the north before separation. Once the DWBC switches back to its southward path the mechanism is restored and the Gulf Stream separation shifts back to the south.

This sensitivity of the Gulf Stream separation point to the existence of the upper core of the DWBC is also evident in the frequency distribution of the variance in Gulf Stream position in the interior. Figure 10 shows the variance of the Gulf Stream position (defined as the latitude of maximum zonal velocity) for the case with no upper DWBC and the central case with 10 Sv transport in the upper DWBC. It is clear that the presence

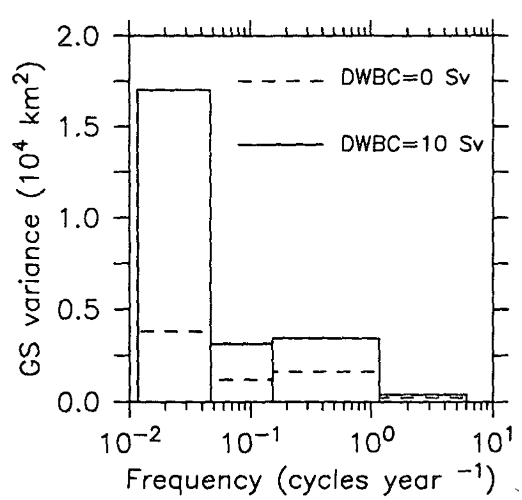

FIG. 10. Spectral distribution of the variance in Gulf Stream position for a case with no upper DWBC (dashed line) and a case with 10 Sv upper DWBC (solid line).

of the upper DWBC has increased the variability of the Gulf Stream path at all frequencies. The increase is by more than a factor of 2 for all periods less than 21 years and by more than a factor of 4 at the longest period (21–82 years). This variability results directly from the oscillating entrainment and detrainment of the upper DWBC and the shift in the separation point of the Gulf Stream. Clearly, the processes at the crossover have large consequences on the circulation far into the interior.

3) AGE OF THE DWBC

The properties of the upper DWBC are also strongly influenced as a result of the oscillation. The model has been integrated with an ideal age tracer that represents the time since the upper DWBC parcels were introduced to the basin (e.g., see Thiele and Sarmiento 1990 and Part I). The age in the core of the DWBC 500 km south of the crossover for the central calculation is shown in Fig. 11a by the solid line. Even though the upper DWBC inflow is steady, the age south of the crossover varies quite strongly with peaks of greater than 15 years. Each of these peaks in age is closely correlated with the entrainment of upper DWBC shown in Fig. 5. The transition between a ''young'' DWBC and an ''old'' DWBC can take place very quickly, on the order of a few years, as the Gulf Stream entrains and detrains the DWBC. As described in Part I, in the absence of the oscillation mechanism the quasi-steady entrainment of the upper DWBC at the crossover was found to increase the effective age of the upper DWBC south of the crossover significantly. However, the age at the southern boundary of a model calculation in which there is no oscillation remains at a nearly steady value of 5 years, as indicated in Fig. 11a for the dashed line. (This model calculation was forced with a reduced Gulf Stream transport of 25 Sv. The dependence of the oscillator on the model forcing parameters is discussed in the next section.) A similar indication of the increased low-frequency variability in the presence of the upper DWBC is indicated by the spectral distribution of age variance shown in Fig. 11b.

This increase in age downstream of the crossover during the high energy events is partially a result of the longer path that the DWBC water flows along in the entrained state. However, the same type of oscillation is seen in a passive tracer that does not increase its value with time, indicating that some of this increase in age is a result of mixing with older unventilated waters in the interior. Even in this steadily forced model, the internal dynamics of the crossover point can result in strongly time-dependent behavior downstream of the crossover. The existence of internal recirculation gyres, and the possibility of time-dependent exchange between the DWBC and the interior, can make correct interpretations of observed variability in the DWBC very difficult.

4. Sensitivity to forcing

The sensitivity of the oscillating mechanism to variations in the Gulf Stream transport, amplitude of the surface wind stress, and strength of the lower DWBC are now explored. The variance of the age in the DWBC south of the crossover is used as an indicator of the strength and frequency of the oscillation. The range of parameter values tested here is within realistic bounds for the North Atlantic; however, it is partially constrained by the model limitation that none of the layer thicknesses vanish and the desire to keep the model configuration (stratification, number of layer, etc.) the same for all experiments. The range of forcing parameters explored is sufficient to demonstrate the response of the oscillation amplitude and frequency to the model forcing, and to interpret those changes in terms of the processes at the Gulf Stream/DWBC crossover. This model is most appropriately viewed as a means to explore the basic dynamics of the system; hence, the quantitative values of the forcing parameters are not as important as the qualitative response to changes in those parameters.

a. Gulf Stream inflow transport

The existence of the oscillation is quite sensitive to the strength of the specified transport of the Gulf Stream leaving the shelf. For a Gulf Stream transport of 25 Sv there is very little evidence of variability in the DWBC age, as indicated in Table 1 for run 2 and by the dashed line in Fig. 11. Increasing the Gulf Stream transport to 37.5 Sv results in significant variability in the lowest-frequency band (Table 1, run 3), while a further increase to 50 Sv results in a very large peak in the decadal band (Table 1, run 1, and Fig. 11). The reason for this sensitivity is that the strength of the Gulf Stream transport strongly influences the efficiency of the entrainment mechanism, as discussed in Part I. Briefly stated, the stronger the Gulf Stream transport, the larger the change in the thickness of layer 1 where the Gulf Stream leaves the shelf and the further offshore the upper DWBC must flow in an attempt to conserve its potential vorticity. As shown in Part I, if the upper DWBC is deflected sufficiently offshore, it interacts with the interior recirculation gyres and sheds eddies of low potential vorticity water into the interior. This entrainment can lead to the oscillating mechanism discussed here. If the amount of DWBC entrained is insufficient to change the stability characteristics of the Gulf Stream (as with 25 Sv of Gulf Stream transport), the oscillation does not arise. For sufficient entrainment, the period of the oscillation is determined by the time it takes to stabilize the stream and shut down the entrainment mechanism. For 37.5 Sv the mechanism is relatively inefficient, the amplitude of the oscillation is small, and the period is long. For 50 Sv inflow the entrainment of upper DWBC is strong, the resulting

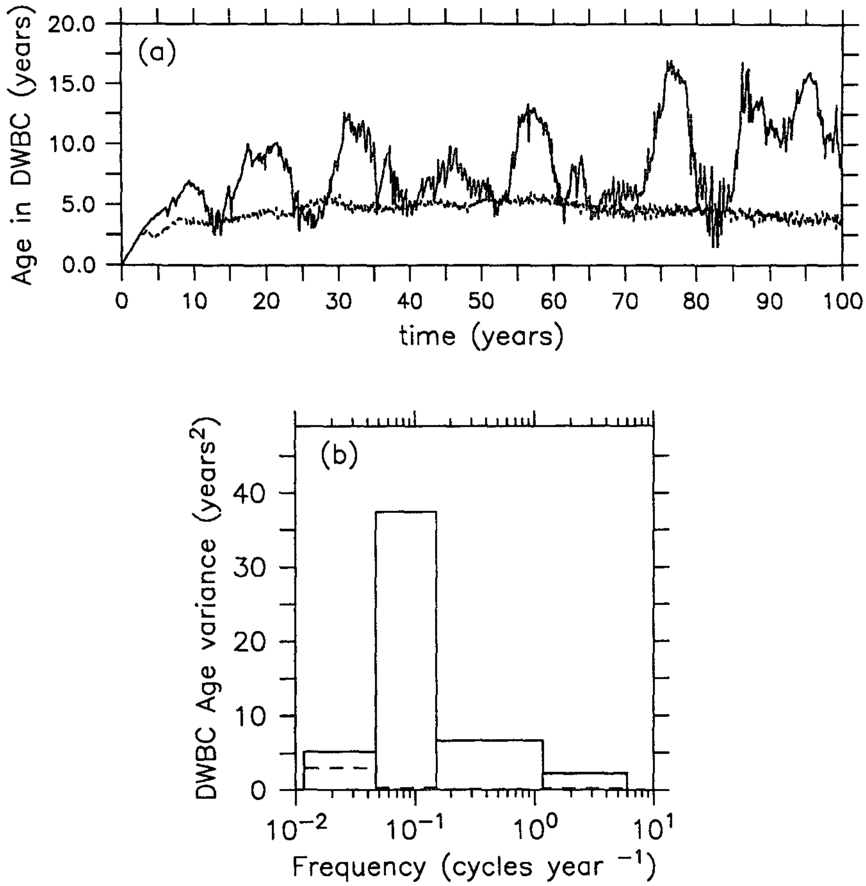

FIG. 11. (a) Age within the core of the upper DWBC at the southern model boundary for transports Γ = (50, 10, 10) (solid line) and Γ = (25, 10, 10) (dashed line). The solid line is an example of variability in the DWBC age induced by internal oscillations at the crossover; the dashed line is an example of the DWBC age in the absence of such low-frequency oscillations. (b) Spectral distribution of age variance for the same two calculations.

amplitude of the oscillation is larger, and the period of the oscillation is shorter.

b. Wind stress amplitude

The strength and frequency of the oscillation is also dependent on the amplitude of the wind stress. The pattern has been assumed to be the same as in the central case [given by Eqs. (1) and (2)] and constant in time. The amplitude of the oscillation is weakened for a maximum surface wind stress of τ_m = 0.25 dyn cm^{-2} compared to the central case of τ_m = 0.5 dyn cm^{-2}, as indicated in Table 1 for run 4. The spectral peak is decadal in both cases, although the low-frequency energy is larger for the weaker wind forcing. This decrease in amplitude and frequency of the oscillation with decreasing forcing is as expected because the lower wind results in a weaker Gulf Stream, weaker wind and eddy-driven recirculation gyres, and a less efficient entrainment of upper DWBC water at the crossover, as discussed in Part I. An increase in the strength of the wind stress to τ_m = 0.75 dyn cm^{-2} (run 5) results in an increase in the strength of the oscillation at both high (1–8-year periods) and low (21–85-year periods) frequencies. The increase in amplitude is expected as the stronger Gulf Stream results in stronger recirculation

TABLE 1. Model run conditions.

Run	Wind (dyn cm^{-2})	Γ_1 (Sv)	Γ_2 (Sv)	Γ_3 (Sv)	Age variance spectral period (years) 21–85	7–21	1–7	0.16–1
1	0.50	50	10	10	5.2	37.4	6.6	2.2
2	0.50	25	10	10	3.0	0.3	0.1	0.2
3	0.50	37.5	10	10	17.6	3.2	1.4	1.4
4	0.25	50	10	10	12.0	20.0	3.7	1.4
5	0.75	50	10	10	57.6	20.9	13.2	4.4
6	0.50	50	10	20	15.2	21.9	2.2	1.3
7	0.50	50	10	30	0.5	1.0	0.5	1.4

gyres and large entrainment of the upper DWBC. A typical amplitude of the variability in the age of the DWBC water is 7 years, greater than the mean time it takes to transit the model domain from the northern boundary. The maximum amplitude also shifts to lower frequencies with the increase in wind stress. This is because the entrainment of the upper DWBC water is no longer purely controlled by the eddy-driven recirculation gyres, but is now also influenced by the wind-driven recirculation gyres. The detrainment of the DWBC water continues on a longer timescale after the Gulf Stream is stabilized because the wind-driven circulation continues to entrain water and is not influenced by the stability of the stream. The increase in higher-frequency variability in the 1 to 8 year period is also found in the time series of the eddy fluxes, indicating that the increased efficiency of the eddy-driven entrainment mechanism resulting from the more unstable Gulf Stream is responsible for this increase.

c. *Lower DWBC transport*

As discussed in Part I, the efficiency of the upper DWBC entrainment also depends on the strength of the lower DWBC, or more precisely it depends on the interface slope between the upper and lower cores of the DWBC. The upper DWBC shifts offshore at the crossover in response to the deepening of the main thermocline below the Gulf Stream. The upper core is able to continue under the Gulf Stream while preserving its layer thickness (or potential vorticity) by sliding down the interface between the upper and lower cores of the DWBC. In the absence of interior recirculation gyres, the upper DWBC would preserve its potential vorticity and then continue to the south along the western boundary. However, for a sufficient offshore shift, the interior recirculation gyres are able to entrain some, or all, of the upper DWBC under the Gulf Stream. Thus, the efficiency of the entrainment depends on the distance the upper DWBC is shifted offshore at the crossover. It was shown in Part I that for a stronger lower DWBC there was less entrainment of the upper DWBC.

The strength and frequency of the oscillator are also influenced by the strength of the lower DWBC. As expected, as the lower DWBC transport is increased to 20 Sv (run 6), the entrainment mechanism is weakened and the strength of the oscillation decreases. There is also a shift toward lower frequencies because it takes longer for the less efficient entrainment mechanism to stabilize the Gulf Stream. For a sufficiently large lower DWBC transport of 30 Sv (run 7) the oscillation is essentially eliminated because the upper DWBC does not flow sufficiently offshore to be entrained by the interior recirculation gyres.

5. Discussion and conclusions

A new low-frequency oscillation in the Gulf Stream/deep western boundary current system is proposed. Eddy fluxes arising from an unstable Gulf Stream drive recirculation gyres to the north and south of the Gulf Stream in the basin interior. These recirculation gyres can entrain the upper core of the DWBC so that it flows under the Gulf Stream. This entrainment stabilizes the basic-state profile of the Gulf Stream by changing the meridional gradient of potential vorticity below the main thermocline. The decrease in eddy fluxes results in a decay of the northern and southern recirculation gyres and the return of the upper DWBC to a southward flowing path along the western boundary. With the loss of entrained DWBC water, the meridional gradient in potential vorticity below the thermocline becomes negative again and the Gulf Stream returns to its relatively unstable state. At this point the cycle is ready to start again. The oscillation is found to exist for a wide range of forcing parameters consistent with those for the real North Atlantic. The strength and frequency of the oscillation can be strongly dependent on the forcing parameters; however, the period is approximately decadal (10–30 years) and the amplitude is zeroth order. The sensitivity of the amplitude and frequency to variations in forcing is as expected based on the understanding of the oscillator dynamics.

The oscillation has large consequences for many aspects of the basin-scale flow. In the stable state the Gulf Stream penetrates far into the basin and shifts northward with the variability concentrated in the eastern Gulf Stream, the west being stabilized by the entrained DWBC water. In the unstable state the Gulf Stream does not penetrate very far into the basin and shifts southward with the variability concentrated in the west. The transition between states can take place very rapidly, on the order of one to several years. The entrainment and detrainment of the DWBC results in large variations in the properties of the upper DWBC south of the crossover. This can be readily observed in fluctuations in the effective age of the DWBC, which may vary between 5 and 20 years with the period of the oscillation. Again, the transition between a "young" and an "old" DWBC can occur very rapidly, on the order of 1 year, as the DWBC switches from an entrained path to a southward flowing path.

Several assumptions have been made here that deserve some discussion. First, the low number of model layers explicitly filters out all but the first two vertical modes. This is important because, while the entrained DWBC stabilizes the mode of instability that derives its energy from the baroclinicity of the main thermocline, higher vertical modes may continue to be unstable and sustain the eddy-driven recirculation gyres. This influence may mitigate the primary oscillation discussed here, but it will probably not override it. The presence of realistic bottom topography may inhibit the spinup of the eddy-driven recirculation gyres and weaken the entrainment at the crossover. Finally, the model forcing here is very idealized: the addition of time dependence, feedbacks, and buoyancy forcing is likely to introduce additional variability in the system.

It is also possible that the inclusion of diapycnal mixing may erode the low potential vorticity signature of the entrained DWBC water, thus reducing its influence on the stability of the Gulf Stream. A scaling estimate of the change in potential vorticity anomaly due to diapycnal mixing under the Gulf Stream gives

$$\frac{\delta q}{Q} = \frac{KL}{H^2U'}, \tag{3}$$

where δq is the change in potential vorticity in layer 2 due to diapycnal mixing, Q the potential vorticity of layer 2, K the diapycnal mixing coefficient, L the downstream length scale of the recirculation gyres, H the thickness of layer 2, and U is the downstream velocity in layer 2. Assuming typical values of $K = 10^{-4}$ $m^2\ s^{-1}$, $L = 10^6$ m, $H = 10^3$ m, and $U = 10^{-1}$ cm s^{-1} gives $\delta q/Q = 10^{-3}$. This is two orders of magnitude less than the potential vorticity anomaly of the entrained DWBC water, which is about 10% of Q. Thus, it would take very strong ($K = 10^{-2}\ m^2\ s^{-1}$) diapycnal mixing at 1000-m depth under the Gulf Stream for this to significantly influence the stabilizing effects of the entrained DWBC water.

The primary aim of this study is to identify fundamental mechanisms of interaction between the upper and lower limb of the thermohaline circulation at the crossover point and to demonstrate the influences of these local, internal physics on several important aspects of the larger-scale general circulation. With this knowledge of the crossover dynamics, these results may be applied to the study of more complete ocean models and observational data to determine if similar processes are active in these more complex systems.

There have been several basin-scale eddy resolving models of the North Atlantic that include more complete physics and more realistic forcing than the present model. A series of multidecadal, eddy-resolving wind- and buoyancy-forced models of the North Atlantic basin have been carried out under the World Ocean Circulation Experiment (WOCE) Community Modelling Effort (CME) (Bryan et al. 1995). However, there is no evidence of an entrainment of the upper DWBC or the existence of a low-frequency oscillation in those calculations (F. Bryan 1995, personal communication). This may be explained by the lack of a coherent LSW component in the DWBC between Cape Hatteras and the Grand Banks in those model calculations (Bryan et al. 1995; and Böning et al. 1995). Such a DWBC structure would eliminate the source of low potential vorticity water to the interior and the mechanism of the low-frequency oscillations. A study by Ezer and Mellor (1992) reports results from a high resolution, wind and buoyancy forced regional primitive equation model. The oscillation would not be expected (and is not found) in their model for a number of reasons. First, there was no baroclinic deep western boundary current in their model, so the source of low potential vorticity was not present. In addition, the strength of the recirculation gyres was partially imposed through the lateral boundary conditions, limiting any feedbacks with the Gulf Stream stability, and the model integrations were relatively short in duration (5–10 years). Both of the above-mentioned models could, in principle, support the low-frequency oscillation discussed in the present paper. The absence of the oscillation is at least consistent with the mean flow structure along the western boundary of those models lacking the vertical structure of the DWBC imposed in the present model, and observed by Pickart (1992) and PS93.

If such an oscillation does exist in the real ocean, it would have tremendous consequences for our ability to monitor and model the thermohaline circulation. The influence of the time-dependent exchange between the DWBC and interior recirculation gyres stresses the idea that the DWBC does not provide a simple, or steady, connection between high-latitude source regions and the rest of the oceans basins. Large, low-frequency fluctuations may be present in the system independent of variability in the source regions. While this greatly complicates the interpretation of observed variability in the DWBC at a single location, it also provides the opportunity to use the effective age at a series of western boundary locations to identify regions of enhanced exchange with the interior.

The oscillator also impacts the general problem of representing the thermohaline circulation in numerical models, both on the predictive timescale of weeks to months and on the low-frequency climate timescale of years to centuries. On the short timescales, the potential vorticity profile of the Gulf Stream controls its stability characteristics and hence the growth rate (predictability) of small perturbations. It was shown here that the entrainment of upper DWBC water significantly influences the stability characteristics of, and the pattern of variability in, the Gulf Stream. Modeling the thermohaline circulation on long timescales requires a reasonable representation of the Gulf Stream separation for both the proper exchange with the atmosphere and representation of water masses along the western boundary. The present results reemphasize the importance of the DWBC to Gulf Stream separation, as discussed by Thompson and Schmitz (1989), but also introduce the complicating factors of time dependence and feedback. The meridional transports of heat, fresh water, and passive tracers carried by the thermohaline mode are of great importance to climate problems. These properties are strongly influenced by the amount of mixing experienced by parcels, which in turn will be strongly influenced by the parcel pathways and can be expected to be quite different for the limit states identified here. Thus, the correct representation of the DWBC, particularly the upper core, and its entrainment at the crossover has large consequences on a number of modeling issues from the synoptic to climate timescales. As the explicit representation of the entrainment process requires model resolution not presently achievable in climate models, a parameterization of this entrainment may be necessary in order to incorporate these physics into low resolution models.

Acknowledgments. This work was supported by the National Science Foundation Grant OCE-9301323 and NOAA Grant NA36GP0309 through the Atlantic Climate Change Program. Many calculations were carried out on the Cray YMP at the National Center for Atmospheric Research. Don Olson at the University of Miami also provided some computing resources. Harley Hurlburt and Tammy Townsend are thanked for providing the original model code and for answering many questions about the model.

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