Influence of the Meridional Overturning Circulation on Tropical–Subtropical Pathways

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ABSTRACT

An ocean GCM is used for idealized studies of the Atlantic circulation in a square basin. The subtropical, the tropical, and the equatorial gyres are produced by forcing the model with a wind stress profile having only latitudinal dependence. The goal is to understand the effect of the meridional overturning circulation (MOC) on the Atlantic intergyre exchanges. The MOC is imposed by prescribing an inflow all along the southern boundary and an outflow at the northern boundary. The results indicate that the northward flow of the MOC has a crucial effect on the subtropical–tropical pathways. In this idealized configuration the North Atlantic wind field creates a basinwide potential vorticity barrier. Therefore, the water subducted in the North Atlantic has to flow to the western boundary before turning equatorward. This is shown by the trajectories of floats injected in a band of northern latitudes. The warm water return flow of the MOC inhibits this pathway and reduces the inflow of North Atlantic waters into the equator from 10 Sv in the purely wind-driven case to 2 Sv (Sv $\equiv 10^6$ m^3 s^{-1}). Thus, the equatorial thermocline consists mainly of water from the South Atlantic. The analysis of synthetic float trajectories reveals two distinct routes for the return flow of the MOC, the first one occurring in the intermediate layers along the western boundary and the second all across the basin in the surface layer. The surface path starts with water subducting in the South Atlantic subtropical gyre, flowing within the North Brazil Current to the equator, entering the Equatorial Undercurrent (EUC), becoming entrained into the tropical mixed layer, and finally flowing northward in the Ekman layer. The contribution of thermocline water to the MOC return flow is negligible.

1. Introduction

For a long time, the subtropical gyres, the tropics, and the meridional overturning circulation have been treated independently [see Pedlosky (1996), Philander (1990), and Warren (1981) for reviews]. Then Pedlosky (1987) developed a theory for the Equatorial Undercurrent (EUC) that shows that the Tropics and the subtropics must exchange a significant amount of water [see also Liu (1994) and McCreary and Lu (1994) for a more detailed discussion of the pathways]. Evidence for this connection can be found in hydrographic measurements [see Tsuchiya et al. (1989) for the Pacific and Arhan et al. (1998) for the Atlantic]. Other measurements indicate that the water of the meridional overturning circulation (MOC) returns through the upper layers of the tropical Atlantic (Schmitz and McCartney 1993).

This raises the question: to what extent do those three dynamically very different systems influence each other? The physical mechanisms of the tropical–subtropical exchange are based on conservation of potential vorticity and were first described by Pedlosky (1987) in his adiabatic theory of the EUC. Because of the highly diabatic nature of the tropical upwelling system, it was not possible to close the circulation in the framework of this theory. But idealized numerical studies by Liu et al. (1994) and McCreary and Lu (1994) confirmed Pedlosky's results. Those models were symmetric about the equator and focused on the relation between the strength of the wind field and the tropical–subtropical exchange (Liu and Philander 1995).

Further up the hierachy of complexity there are only high-resolution OGCMs with realistic topography and wind fields. Their complexity makes it difficult to analyze the impact that basin geometry, topography, wind field and the MOC have on the tropical–subtropical circulation. To our knowledge only Fratantoni (1996) ever tried to single out the effect of the MOC on the wind-driven circulation. But he too used a very complex model and focused on equatorial variability.

We tried to fill this gap between highly idealized models and very complex models. We took the experiment of Liu et al. (1994) as a starting point and increased the model complexity step by step. It turned out that the influences of bottom topography and basin geometry

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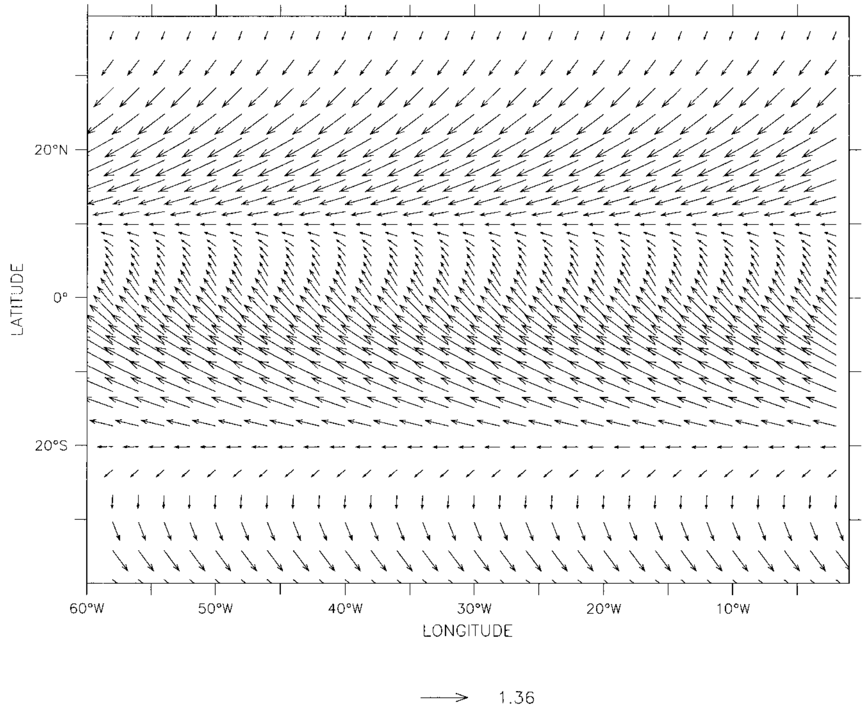

FIG. 1. The wind stress in dyn cm^{-2}.

are either negligible or easily understood compared to the influence that a superimposed MOC has. Thus, we will focus here on the comparison between the the purely wind-driven circulation and the circulation resulting from wind and thermal forcing. The paper is arranged as follows. Section 2 introduces the model and the experiments to be performed; section 3 briefly discusses the flow fields. In section 4, float trajectories are used to study the pathways between the subtropics and the Tropics as well as the cross-equatorial pathways. Section 5 analyzes shortcomings and summarizes the results.

2. The model and the experiments

The model used is the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM, version 2b). The domain is an idealized rectangular basin from 40°S to 40°N in latitude and from 0° to 60° in longitude, with a flat bottom at 3000 m. The longitudinal resolution is 1°; the latitudinal resolution is ⅓° within 6° of the equator and increases linearly to 1° at the domain boundaries at 40°. There are 30 levels in the vertical with a 10-m resolution in the top 100 m.

Horizontal mixing is the constant coefficient scheme with the eddy viscosity and diffusivity of 2000 and 1000 m^2 s^{-1}, respectively. In two sponge layers poleward of 36°, temperature is restored to prescribed values with a restoring timescale of 40 days at 36°. This value decreases to 4 days at 40°. In the vertical a Richardson-number-dependent vertical mixing scheme is used. Unstable temperature gradients are eliminated by mixing heat vertically to a depth that ensures a stable density gradient.

The initial condition is a state of rest. Salinity remains a constant value of 35 psu. The wind stress (Hellerman and Rosenstein (1983), smoothed and averaged in time and longitude) and the wind curl are shown in Figs. 1 and 2. The initial temperature is prescribed as shown in Fig. 1 of Liu and Philander (1995). This profile is also used at the surface to restore the surface temperature with a 40-day relaxation time and as boundary restoring values in the sponge layers (expt 1). Both experiments were integrated for 18 years before any analysis was performed or floats were injected [for a justification see Liu and Philander (1995)]. In the case of the experiment with the superimposed MOC (expt 2) the sponge layers are replaced by open boundary conditions (OBC). It is known that OBC render the problem of solving the primitive equations ill posed (Oliger and Sundstrom 1978). Nevertheless one can make progress if the errors introduced by the OBC are small enough and do not grow in time [see Spall and Robinson (1989) for a detailed discussion].

At the open boundaries temperatures and meridional velocities need to be specified. Furthermore, the tangential velocites of the inflow points have to be defined [on outflow points it can be calculated by assuming conservation of relative vorticity (Spall and Robinson 1989)].

The code used (MOM 2b) provides subroutines for OBC that were modified for our purposes. At the open boundaries the temperature and the barotropic stream-

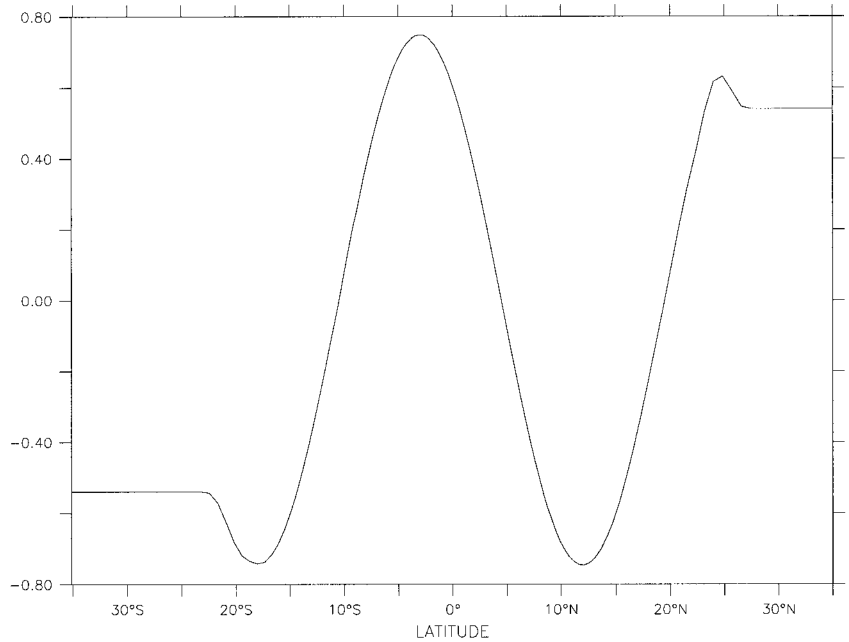

FIG. 2. The wind curl in 10^{-8} dyn cm^{-3}.

function are specified. With the help of the Sverdrup balance and geostrophy the model is supposed to calculate the velocity field (Stevens 1990). As we will see, this does not work perfectly because the inflow is too barotropic.

The open boundaries were put at 35°N/S to allow the inflow to be in agreement with the Sverdrup balance. The barotropic streamfunction and temperature at 35°N/S are taken from the steady-state solution of the purely wind-driven circulation. To simulate the throughflow of the MOC return flow, the barotropic streamfunction is modified so that there are 15 Sv (Sv $\equiv 10^6$ m^3 s^{-1}) flowing into the South Atlantic all along the southern boundary and leaving the North Atlantic in the northwest corner through a western boundary current. These 15 Sv are roughly consistent with numbers from the literature (Schmitz and McCartney 1993) and the Sverdrup transport across 35°S. We did not try to simulate a deep western boundary current because we wanted to focus on the interaction between the thermocline of the tropical Atlantic and the warm water return flow of the MOC.

3. The flow field

Given the difficulties using OBC, we compared the temperature, velocity, and stream-function fields of the two experiments. The first two fields did not reveal any major differences in the interior so that we will just present the barotropic and the overturning streamfunctions for a comparison. Figures 3 and 4 show the barotropic streamfunctions for experiments 1 and 2, respectively. In both pictures we can clearly distinguish the subtropical, the tropical, and the equatorial gyres, which are separated by the zero wind curl lines.

The major differences in these two figures are—except for the 15 Sv offset in Fig. 3—close to the western boundary. In experiment 2 the northward western boundary currents increased their strength by 15 Sv, whereas the southward western boundary currents became weaker by the same amount. This was to be expected from a simple linear superposition of the wind-driven flow and the warm water return flow. We do not expect any changes in the interior because the wind field is the same in both experiments. A nonlinear effect of this superpostion can be seen at 8°N where the North Brazil Current (NBC: for the sake of argument we will name the currents after the corresponding real ocean currents) overshoots its original separation latitude and the North Equatorial Counter Current (NECC) meanders back to the zero wind curl line. Fratantoni (1996) showed that this overshooting is caused by the MOC and is responsible for most of the mesoscale variability in this area.

Of course, those figures show the barotropic streamfunction and not the velocity, but the mentioned currents control most of the transport in that area. The baroclinic structure of the western boundary currents in experiment 2 is shown in Fig. 6. We can see the western boundary currents of the four gyres and a continuous northward flow below 700 m. The northward flow in the northern Tropics is blocked by the western boundary current of the tropical gyre. As we will see later, this has important implications. Away from the boundaries, the interior velocity fields of the two experiments are almost identical and are not shown here [see Philander (1990) for

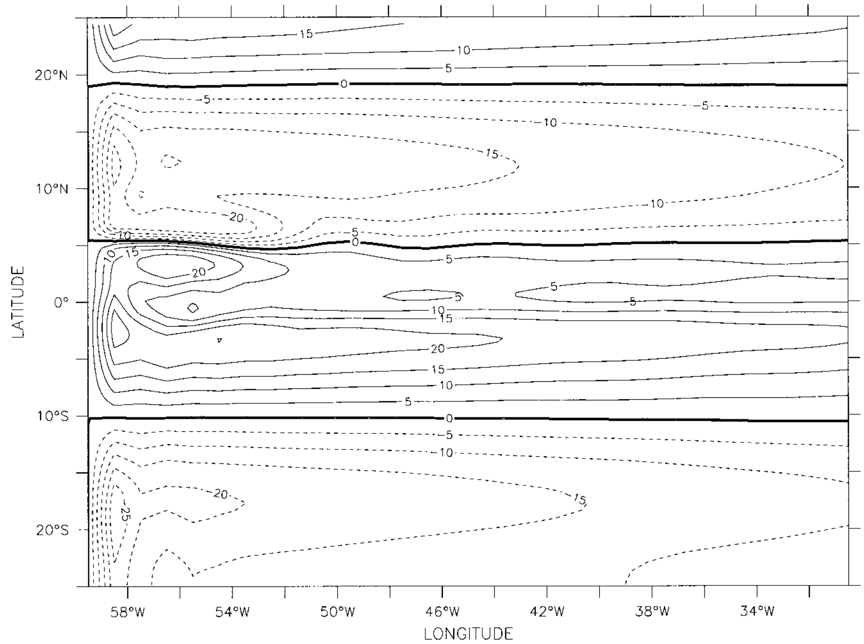

FIG. 3. Barotropic streamfunction in the western tropical part of the basin for expt 1.

their description]. However, there are changes in the transition region between the western boundary and the interior. As an example, the flow field along the boundary between the northern subtropical gyre and the tropical gyre is shown in Fig. 5. In experiment 1 the interior has to feed a southward western boundary current and gains a southward component at the gyre boundaries in the transition region. In experiment 2 (broken arrows), the interior feeds a northward western boundary current and therefore has a northward component in the transition region.

The overturning streamfunction (Fig. 7) indicates that a significant amount of intermediate water is entrained into the equatorial thermocline. It also shows that due to the lack of a deep western boundary current in our model, the return flow of the MOC has an abyssal com-

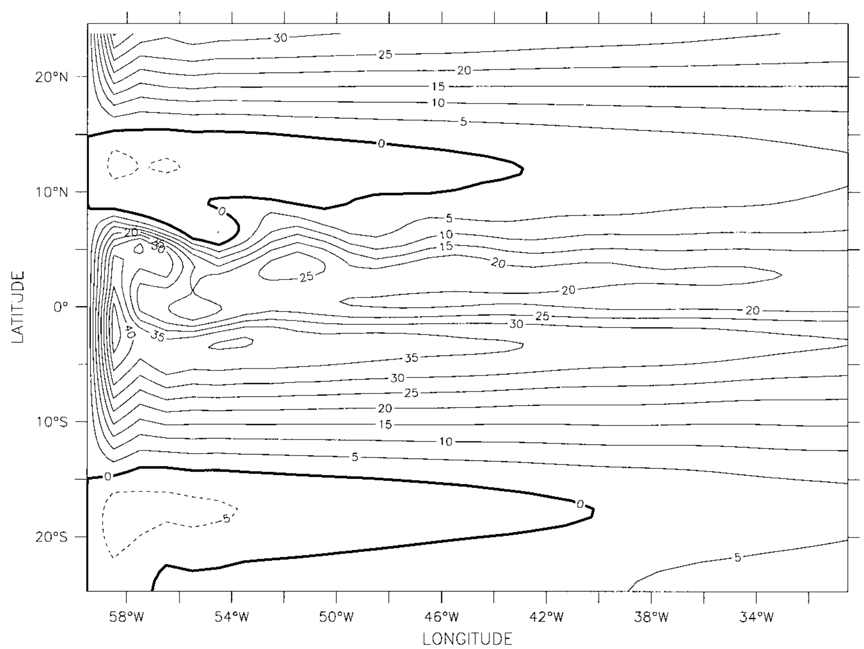

FIG. 4. As in Fig. 3 but for expt 2. Note the intensification of the NBC and its overshooting at 8°N.

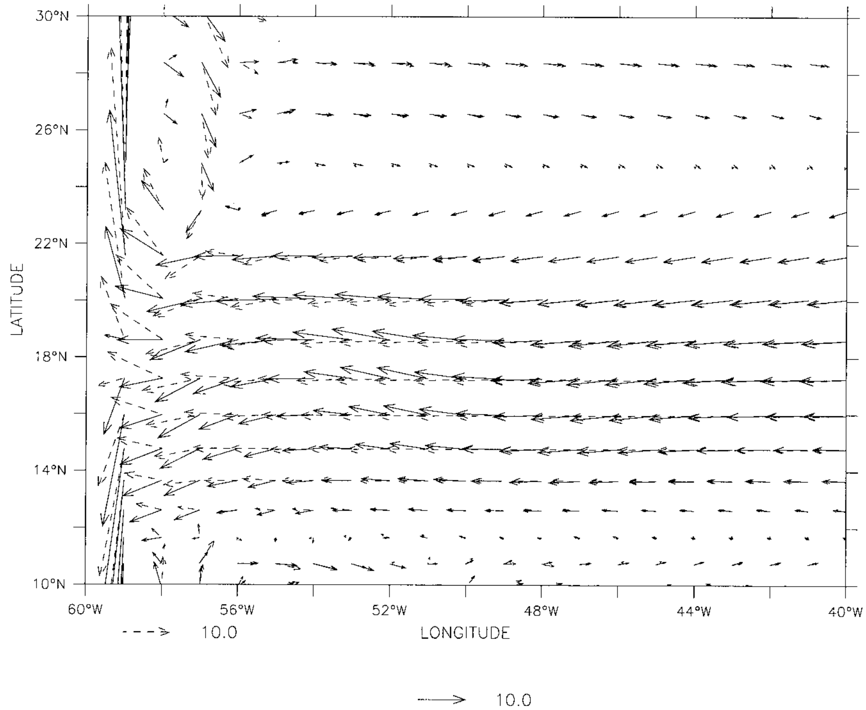

FIG. 5. Velocity in 100-m depth (cm s^{-1}). The broken arrows show the velocity field in expt 2. the other ones show the velocity field in expt 1. In most of the shown domain the fields are almost indistinguishable, but where the gyre boundary approaches the western boundary, the field of expt 1. has a southward component, the field of expt 2 has a northward component.

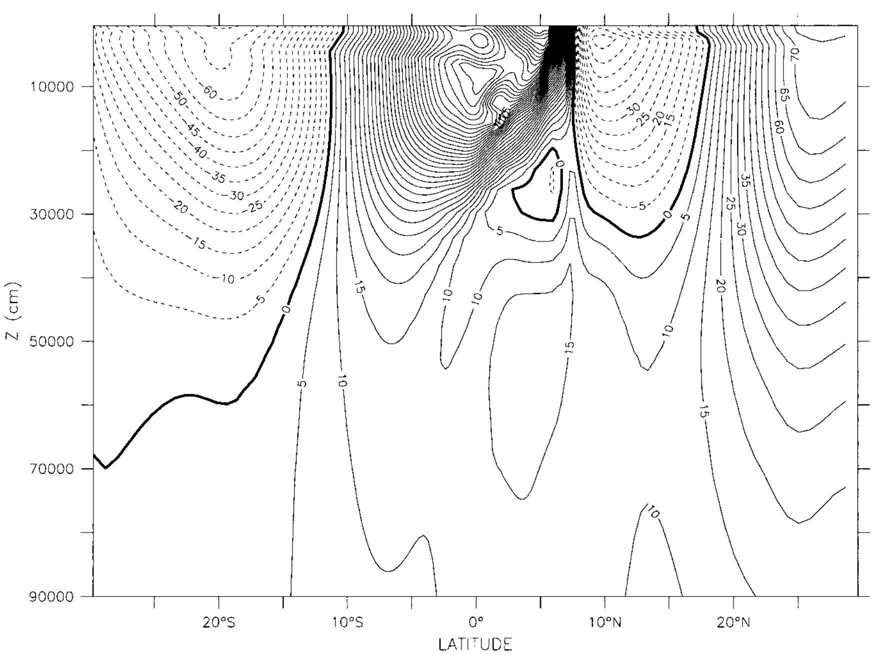

FIG. 6. Meridional velocity in the upper 900 m along the western boundary for expt 2 (countour interval: 5 cm s^{-1}). The maximum velocity is reached in the NBC with 200 cm s^{-1} at 5°N.

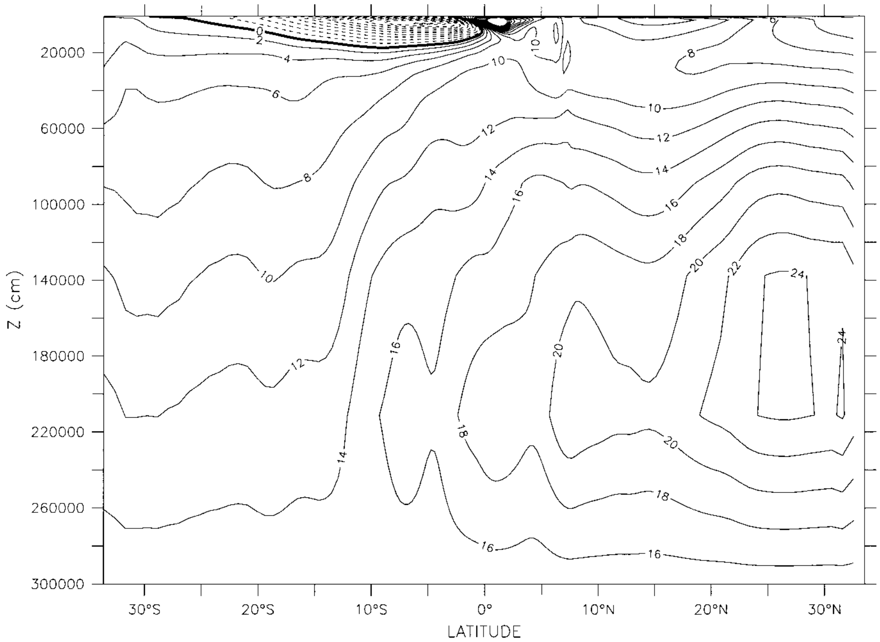

FIG. 7. Overturning streamfunction for expt 2 in Sv.

ponent in the Southern Hemisphere and is not restricted to the upper or intermediate levels as it should be. This spurious deep flow is due to the OBC, and we will assume that its only effect is to weaken the upper-level return flow in the Southern Hemisphere.

At the equator we can see the two tropical cells that are induced by the Ekman upwelling system at the equator. The cross-equatorial winds make them nonsymmetric. We can also see the southern subtropical cell, which apparently merged with the dynamically different southern tropical cell [see Liu et al. (1994) for their dynamics]. The northern subtropical cell is barely noticable, a point which will be discussed in the next section.

4. The tropical–subtropical pathways

After subduction, water from the subtropical gyres has three different possibilities to continue. The water from the westernmost part of the basin recirculates in the subtropical gyre. Water from regions farther east enters the Tropics either through the interior or in a lower western boundary current. This water feeds the EUC, becomes entrained in the equatorial mixed layer, and returns poleward in the Ekman layer (Pedlosky 1987; Liu et al. 1994).

This complex three-dimensional structure of the flow field is best analyzed with floats (virtual particles in our model). We released the floats in different areas of the subtropical gyres. Out of all the resulting trajectories we picked those that highlight the interaction between the MOC and the wind-driven circulation.

Figure 8 shows the trajectories of floats in experiment 1. The floats were released below the Ekman layer in the eastern part of the northern subtropical gyre (NSG). The first thing to notice is that the Ekman suction in the tropical gyre effectively blocks the interior pathway in the north. After subduction down to about 150 m, the water flows west in the North Equatorial Current, flows southward in a lower western boundary current until it leaves the boundary again to flow east in the NECC. From there it flows southward to feed the EUC, upwells, and makes it poleward again in the Ekman layer. It takes about 8 to 10 years to complete this journey.

In experiment 2 the floats were released in exactly the same area but only a minority of them enter the Tropics (Fig. 9). Most of them recirculate in the subtropical gyre. Apparently the return flow of the MOC changed the flow field at the western boundary in a way to inhibit the connection between the northern subtropical gyre and the Tropics. Figure 10 shows that the return flow of the MOC shifted the latitude of vanishing meridional velocity (bifurcation latitude) south by approximately 1–2 degrees in the relevant depths (relevant means that a significant amount of subducted water arrives at those depths at the western boundary). Futhermore, as pointed out already in the last section, the MOC does not only change the flow field at the western boundary, but to a lesser degree changes the field within several degrees of it. A comparison of the endpoints of the vectors nearby the bifurcation latitude shows that this effect is about as big as the effect of shifting the bifurcation latitude. Because the position of the subtropical gyre did not change, this means that some of

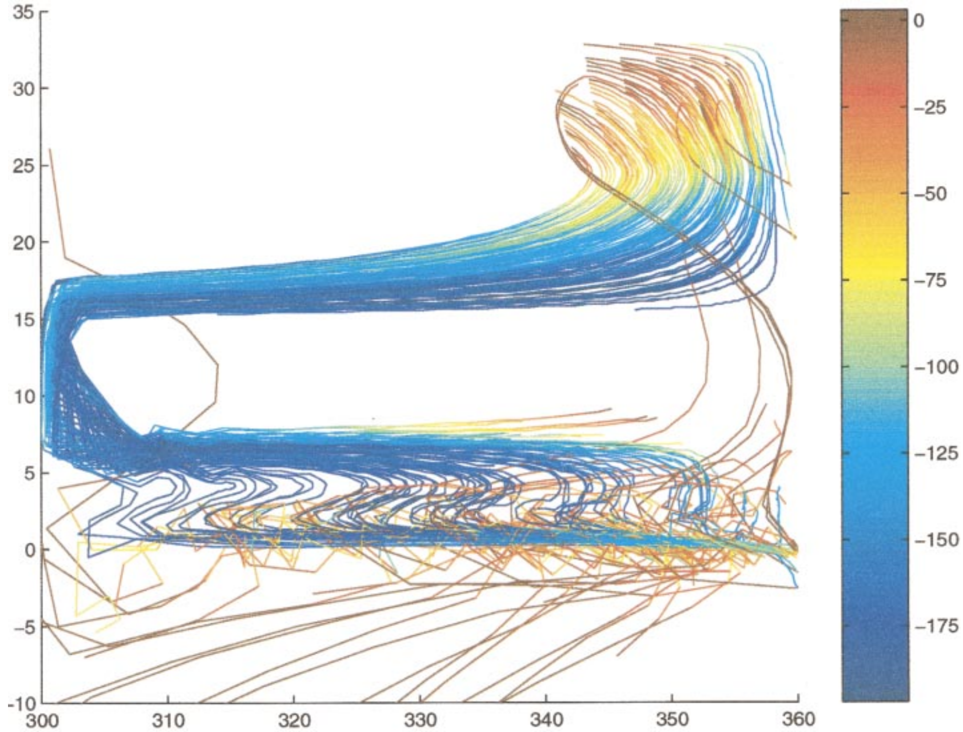

FIG. 8. Trajectories for floats released below the Ekman layer in the northern subtropical gyre of expt 1 (the colorbar indicates the depth in meters).

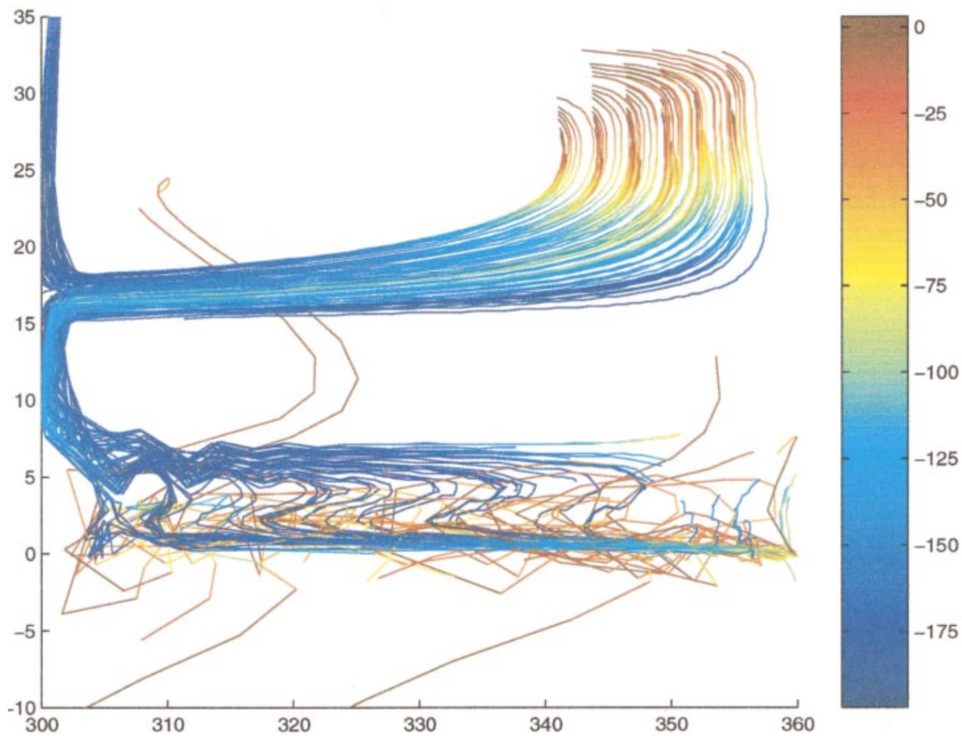

FIG. 9. As in Fig. 8 but for expt 2.

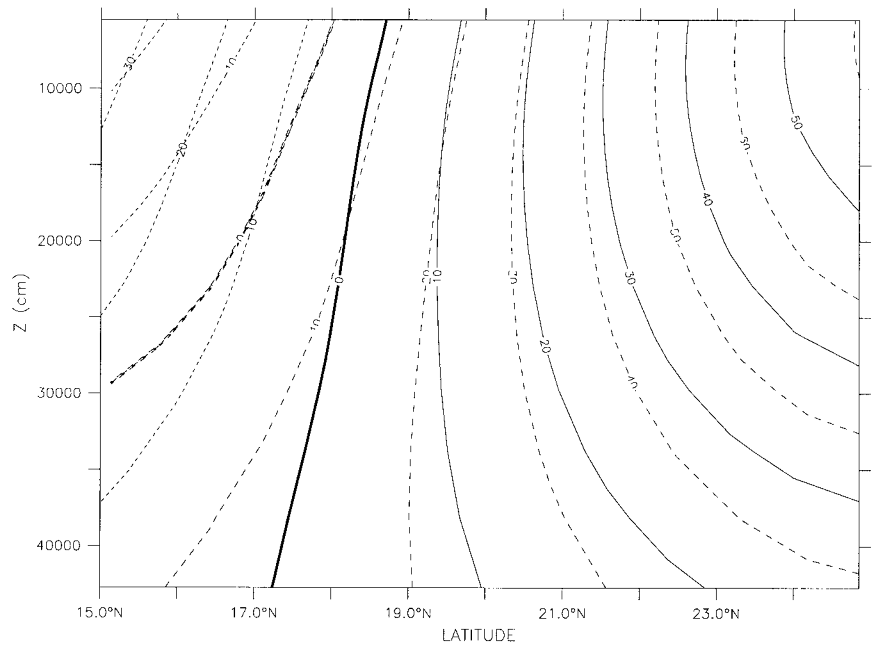

FIG. 10. Velocities of the area where the floats enter the western boundary currents (in cm s^{-1}). The straight zero line indicates the bifurcation latitude for expt 1, the broken zero line the bifurcation latitude for expt 2.

the water that turned southward in experiment 1 has to turn northward in experiment 2.

To quantify those changes in the cross-equatorial flow, it is easiest to look at the overturning streamfunction in both experiments. As Fig. 11 shows, there are 17 Sv of water upwelling into the equatorial thermocline from the southern subtropical gyre and 10 Sv from the northern subtropical gyre. This asymetry is due to the wind field. By looking at Fig. 12 we see that in experiment 2 25 Sv come from the southern and only 2 Sv from the northern subtropical gyre (the equatorward gyre boundaries are at 10°S and at 18°N, respectively).

What are the exact pathways of the MOC return flow? The analysis of float trajectories shows that there are just two possiblities in our model. The first one is a flow along the western boundary in intermediate layers (Fig. 13). The second one is a complicated journey through the thermocline and Ekman layers of the equator. Because of the southward motion in the interior of the equatorial gyre the water has to flow north either in the Ekman layer or along the western boundary to reach the northern midlatitudes. Two typical trajectories are displayed in Fig. 14. The water subducts in the southern subtropical gyre to about 200 m and enters the NBC, which feeds the EUC. The EUC transports the water east where it upwells into the mixed layer. After another upwelling and downwelling cycle in the equatorial upwelling system it finally escapes north in the Ekman layer.

Without tracers it is difficult to tell exactly how much of the MOC return flow passes through the Ekman layer or the intermediate western boundary current (IWBC). So we carried out an additional run identical to experiment 2 but without wind. Apart from the nonlinear overshoot region of the NBC/NECC the results in the flow field were almost identical to the difference of experiments 2 and 1 (not shown). Below the thermocline and above the abyss, we interprete this as the intermediate layer (400–1500 m) return flow of the MOC that is constrained to the western boundary. An integration of the flow at the western boundary yields an IWBC transport of about 7 Sv across the equator and next to nothing in the abyss. Due to lateral influx and upwelling of abyssal waters at the western boundary, the IWBC increases its transport to about 11 Sv at 20°N. Thus, we can estimate that about one-half of the MOC return flow is carried by the IWBC; the other half has to pass through the Ekman layer.

5. Summary and discussion

A highly idealized numerical study of the tropical–subtropical pathways has been performed. Neither are the results revolutionary nor do we claim to give a realistic picture of the circulation in the tropical Atlantic, but our experiments do highlight one aspect of the interaction between the MOC and the wind-driven gyres. As such they are one step in a series of process studies that will eventually lead to a high-resolution time-dependent OGCM of the tropical/subtropical Atlantic.

In reality the tropical gyre and its induced western

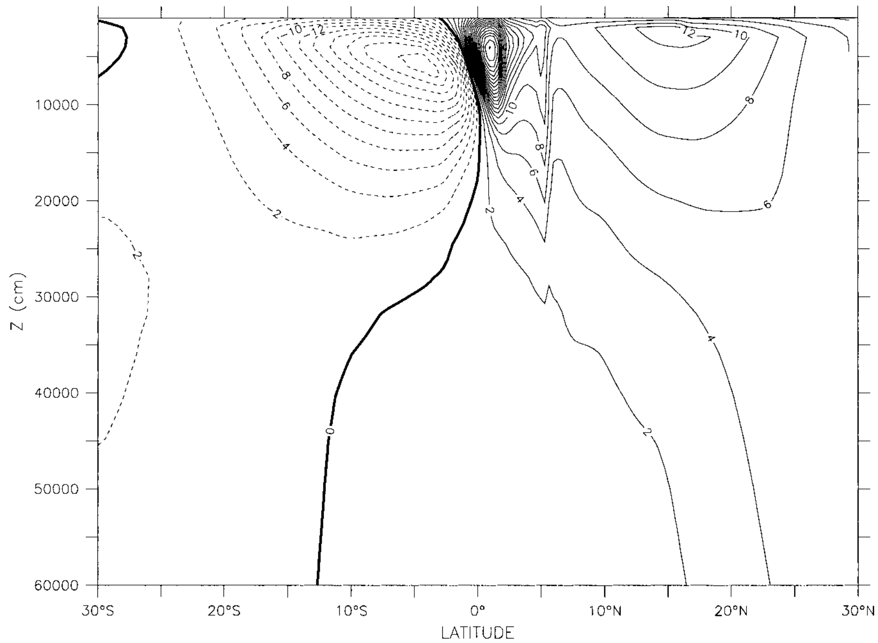

FIG. 11. Overturning streamfunction for expt 1 in Sv. Just the upper 600 m are shown to emphasize the structure in the thermocline.

boundary current are about half as strong as in our model (Wilson and Johns 1997). This is due to our idealized basin, which has a constant width of 60° whereas the Atlantic at the latitude of the tropical gyre has a width of only 40°. Furthermore there is one more possible pathway that has not been considered yet, simply because our model does not resolve it: the vorticies shed by the NBC retroflection (Fratantoni et al. 1995). They have estimated transports of about 2–4 Sv, but it is not clear how much of this water escapes the tropical gyre and makes it to the NSG or the Caribbean Sea. The wind field is another critical issue. It is known (Schott et al. 1995; Johns et al. 1998) that the circulation in the tropical Atlantic has a strong seasonal component. Also, the zonal variations of the wind field are important, especially for the tropical gyre [Inui (1999, personal

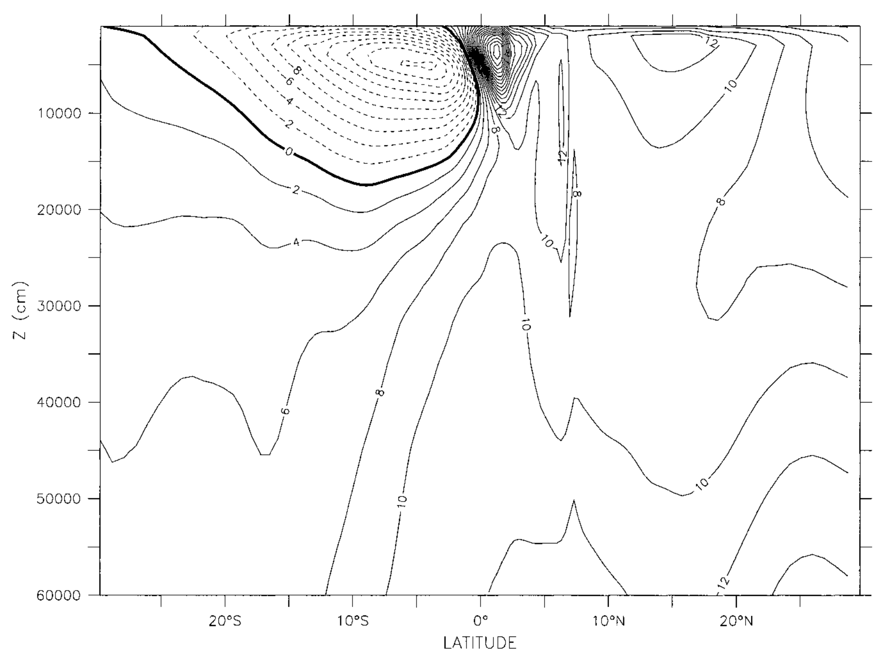

FIG. 12. As in Fig. 11 but for expt 2.

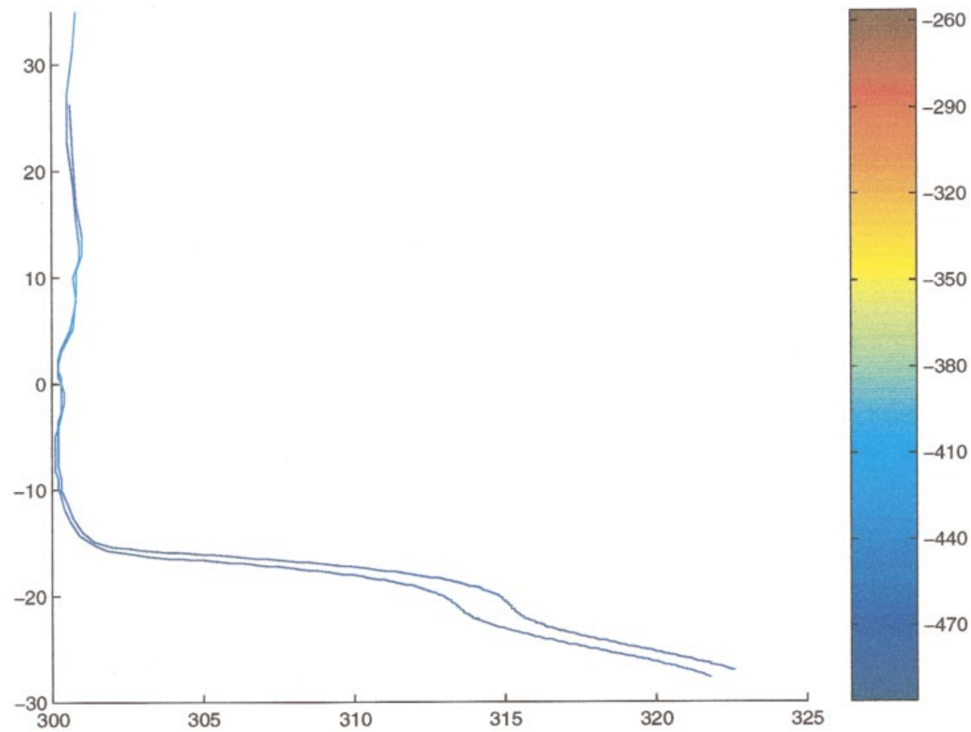

FIG. 13. Float trajectories in the intermediate level.

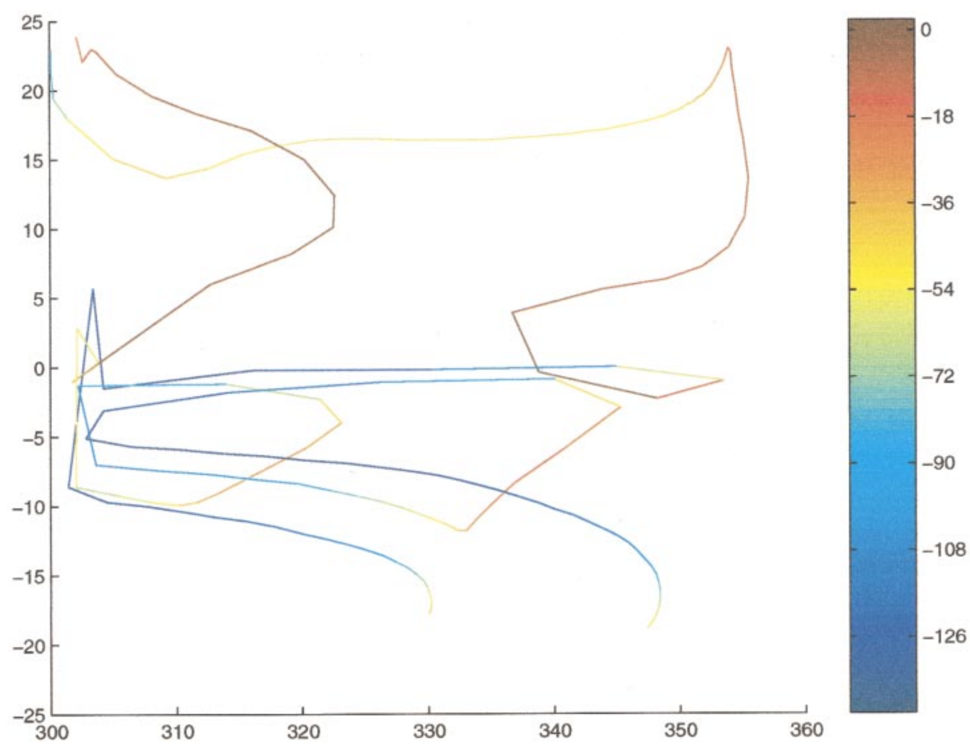

FIG. 14. Two floats that take the surface path to the north. They move to the EUC, join the tropical cell, and after another upwelling–downwelling cycle they escape to the north in the Ekman layer.

communication) showed that the da Silva wind field allows for an interior pathway, whereas the Hellerman-Rosenstein wind field does not]. Nevertheless our experiments do offer dynamical explanations for some of the observed features of the tropical Atlantic. That the equatorial thermocline consists mainly of South Atlantic water (Kawase and Sarmiento 1985) can now be explained by the specific shape of the wind field in the tropical Atlantic and the way the MOC interacts with the resulting flow at the western boundary. The wind field in the North Atlantic strongly inhibits an interior pathway and allows only a lower western boundary current as a path from the NSG to the Tropics. However, at the western boundary the MOC return flow forces this water north and blocks this path.

Second, Wilson and Johns (1997) showed during a carefully designed series of cruises, that not more than 8 Sv of South Atlantic water enters the Caribbean Sea through the southern passages. This water flows there in the surface layer and in the intermediate layer, but no thermocline water from the South Atlantic enters the Caribbean Sea. Our experiments show how the thermocline water of the South Atlantic feeds the NBC, which retroflects into the EUC (only the surface part of the NBC flows into the NECC). In the EUC the water is entrained into the mixed layer and continues northward in the Ekman layer (Fig. 14). This not only shows why there is no South Atlantic thermocline water found in the North Atlantic. Moreover it may explain why only half of the MOC return flow enters the Caribbean Sea through the southern passages. The other half might flow north in the Ekman layer all across the basin and can enter the Caribbean Sea anywhere or not at all.

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