An Overlooked Problem in Model Simulations of the Thermohaline Circulation and Heat Transport in the Atlantic Ocean

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(Manuscript received 28 December 1993, in final form 12 May 1994)

ABSTRACT

Many models of the large-scale thermohaline circulation in the ocean exhibit strong zonally integrated upwelling in the midlatitude North Atlantic that significantly decreases the amount of deep water that is carried from the formation regions in the subpolar North Atlantic toward low latitudes and across the equator. In an analysis of results from the Community Modeling Effort using a suite of models with different horizontal resolution, wind and thermohaline forcing, and mixing parameters, it is shown that the upwelling is always concentrated in the western boundary layer between roughly 30° and 40°N. The vertical transport across 1000 m appears to be controlled by local dynamics and strongly depends on the horizontal resolution and mixing parameters of the model. It is suggested that in models with a realistic deep-water formation rate in the subpolar North Atlantic, the excessive upwelling can be considered as the prime reason for the typically too low meridional overturning rates and northward heat transports in the subtropical North Atlantic. A new isopycnal advection and mixing parameterization of tracer transports by mesoscale eddies yield substantial improvements in these integral measures of the circulation.

1. Introduction

In an analysis of an early, coarse resolution model of the wind- and thermohaline-driven circulation in a simple ocean basin, Veronis (1975) pointed out a discrepancy of the vertical velocity field at middepths, compared to that expected on the basis of theoretical ideas advanced by Stommel (1958). Instead of a broad upwelling at the base of the thermocline, the numerical model showed a downward motion in the interior and an intense upwelling in the western boundary layer of the subtropical gyre. From a scale analysis of the terms in the dynamical equations, Veronis concluded that the horizontal diffusion of heat across the strong front associated with the model's Gulf Stream must represent a dominant term in the heat budget of that region, balanced only by vertical advection. That conclusion is supported by more recent results from sensitivity studies with a similar model by Cummins et al. (1990): their analysis of the heat budget indicates that the vertical advection pattern near the western boundary closely reflects the pattern of horizontal diffusion.

Spurious upwelling inshore of the western boundary current in the North Atlantic was identified as one of the major flaws in the global ocean model used by Toggweiler et al. (1989) to simulate the uptake and distribution of bomb-produced carbon 14. About one-third of the model's North Atlantic deep water (NADW) production returned to the surface in that area, resulting in a glaring deficit in the tracer-inventory distribution. The model used by Toggweiler et al. (1989) was built on the Geophysical Fluid Dynamics Laboratory (GFDL) ocean general circulation model with 12 vertical levels and a $4.5° \times 3.75°$ horizontal grid, a configuration similar to that typically used for coupled modeling studies of the greenhouse effect [e.g., Washington and Meehl (1989); Manabe et al. (1990)]. Strong upwelling in midlatitudes, which significantly reduces the amount of deep water that is carried from the deep water formation regions toward low latitudes and the equator, is evident in almost all recently published results from (level) models of the large-scale ocean circulation: it can be found not only in studies based on the GFDL model (Sarmiento 1986; Suginohara and Aoki 1991; England 1993; Washington et al. 1994) but also in the Hamburg large-scale geostrophic model of Maier-Reimer et al. (1993), an exception being the high-resolution global model of Semtner and Chervin (1992) that used a robust diagnostic technique for the layers below 700 m.

While the deficits in the tracer simulation carried out by Toggweiler et al. (1989) raised the question of

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the ability of ocean models to accurately simulate the penetration and pathways of anthropogenic changes in greenhouse gases and tropospheric heat, the relevance of the artificial shortcut in the meridional overturning pattern for the poleward heat transport in the North Atlantic seems to have slipped the attention of the modeling community. The failure of many ocean models to reproduce the observed northward heat transport of 1.2 ± 0.3 PW in the subtropical North Atlantic near 25°N (Hall and Bryden 1982; Roemmich and Wunsch 1985), corresponding to a too weak overturning in that area, is commonly thought to be the result of a too weak deep-water formation rate in the subpolar North Atlantic. Since model studies show a strong sensitivity of NADW production to the specifications of the poorly known freshwater and heat fluxes in the subpolar North Atlantic [e.g., Maier-Reimer et al. (1993); Döscher et al. (1994)], it is tempting to "improve" model performances by increasing the production rate.

In this paper, we wish to direct attention to the fact that the spurious upwelling in midlatitudes can significantly contribute to an overturning deficit in the subtropical North Atlantic. We shall use results from various runs with the North Atlantic model that has been developed as a "Community Modeling Effort" (CME) for the World Ocean Circulation Experiment (Bryan and Holland 1989), differing in horizontal resolution (1° × 1.2°; 1/3° × 2/5°; 1/6° × 1/5°), wind and thermohaline forcing, and mixing parameterization. Previous examinations of the CME results dealt with aspects of the circulation in the tropical and subtropical North Atlantic and its seasonal variation (Spall 1990, 1992; Böning et al. 1991; Schott and Böning 1991; Didden and Schott 1992; Böning and Hermann 1994; Bryan et al. 1994) and the generation of mesoscale eddies (Treguier 1992; Stammer and Böning 1992; Beckmann et al. 1994a,b). Here we will examine the midlatitude upwelling pattern, compare it with the observational picture of the thermohaline circulation in the North Atlantic, and discuss its implication on the northward transport of heat. A broader discussion of the mean wind-driven and thermohaline circulation in the CME and its dependencies on the forcing, resolution, and mixing parameterizations is presented in Holland and Bryan (1993), Bryan et al. (1995), and Böning et al. (1994). In addition to these many results with the conventional use of horizontal tracer diffusion, we briefly report on some encouraging results with a new isopycnal mixing and eddy advection parameterization (Gent and McWilliams 1990; hereafter G—M) showing a strong and beneficial effect on the overturning pattern and heat transport in the midlatitudes.

2. Model parameters

For a thorough description of the numerical model and the various experiments that have been carried out, the reader is referred to the papers cited above. At this point it may suffice to give a brief review of the model configuration and of those model factors that will be shown to have an effect on the upwelling pattern. The model is based upon the GFDL ocean circulation model (Bryan 1969; Cox 1984) and spans the Atlantic Ocean with realistic topography between 15°S and 65°N. All model versions use 30 levels in the vertical, with a vertical viscosity of 10 $cm^2 s^{-1}$ and vertical diffusivity of 0.3 $cm^2 s^{-1}$.

The high resolution cases (lat × long grid sizes of 1/3° × 2/5° and 1/6° × 1/5°) use biharmonic diffusivity (A_h) and viscosity (A_m) for horizontal mixing. There are a number of model runs, differing mainly in the wind stress forcing [see, e.g., Bryan et al. (1995)] that show little differences in the overturning pattern. Therefore, for the purpose of the present study, we shall focus mainly on one 1/3° case (N13-2) and one 1/6° case (N16-1), both forced with the monthly mean wind stresses given by Hellerman and Rosenstein (1983; hereafter HR). The mixing coefficients for these cases are $A_m = -1.0 \times 10^{19}$ $cm^4 s^{-1}$, $A_h = -2.5 \times 10^{19}$ $cm^4 s^{-1}$ and $A_m = -0.4 \times 10^{19}$ $cm^4 s^{-1}$, $A_h = -0.4 \times 10^{19}$ $cm^4 s^{-1}$ for the 1/3° model and 1/6° model, respectively.

The coarse resolution (1° × 1.2°) version of the model has been run with different horizontal mixing schemes. In the basic configuration, traditional Laplacian diffusion was used. The standard parameters for the horizontal viscosity and diffusivity were $A_m = 1 \times 10^8$ $cm^2 s^{-1}$ and $A_h = 1 \times 10^7$ $cm^2 s^{-1}$. In this configuration, several experiments were carried out with some artificial changes in the wind and thermohaline forcing that will allow us to examine the effect of changes in the Gulf Stream transport on the upwelling at its inshore edge. The reference case (N1-12.0) used HR wind stresses and buffer zones at the northern and southern boundaries with a relaxation of temperature and salinity to the climatology of Levitus (1982); in N1-12.2 the northern buffer zone, and in N1-12.4 both buffer zones were shut off, decreasing the maximum overturning and, hence, the poleward transport of upper-layer water by the Gulf Stream. In N1-12.5, the wind stress was increased by 30%, and in N1-12.7, the wind forcing was set to zero.

Additional experiments, using the same forcing as the reference case N1-12.0, have been carried out to shed some light on the effect of the horizontal mixing. In N1-25.0, the horizontal viscosity was increased to $A_m = 4 \times 10^8$ $cm^2 s^{-1}$ {corresponding to an increase of the frictional boundary layer width $[(A_m/\beta)]^{1/3}]$ from about 80 km to 130 km}, and diffusivity was decreased to $A_h = 0.5 \times 10^7$ $cm^2 s^{-1}$ in an attempt to reduce the effect of horizontal mixing across the Gulf Stream front.

A more drastic effect on the upwelling pattern in coarse resolution models may be expected by adopting an isopycnal mixing scheme. It is widely held that mixing in the ocean interior is more nearly aligned with surfaces of constant density than surfaces of constant depth (McDougall and Church 1986). One option in a z-coordinate model is the diffusion formulation described by Redi (1982) based on a rotation of the diffusive flux vector to a coordinate locally tangent to an isopycnal surface. As demonstrated by Cox (1987), this scheme has a strong impact on the spreading of passive tracers in the main thermocline, leading to property distributions more similar to eddy-resolving calculations than those obtained with the horizontal mixing scheme. However, while the diapycnal component of diffusion associated with horizontal diffusivity in regions with sloping isopycnals is reduced, it is not eliminated in the Redi–Cox scheme. For numerical reasons, a "background" horizontal diffusivity is still necessary to control numerical noise on the grid scale. Several cases were carried out where the Redi–Cox scheme was implemented with depth-dependent isopycnal diffusivities (typically 2×10^7 cm^2 s^{-1} in the upper ocean, decreasing to 5×10^6 cm^2 s^{-1} in the deeper layers) and a background horizontal diffusivity of 1×10^6 cm^2 s^{-1}.

A new parameterization of the effect of mesoscale eddies on the transport of tracers was presented by Gent and McWilliams (1990). It includes mixing along isopycnal surfaces of the thickness between isopycnal surfaces that implies the dynamical effect of depletion of available potential energy and vertical transfer of momentum in ways that mimic the adiabatic processes of baroclinic instability and isopycnal form stress. This term appears as an eddy-induced transport velocity in the tracer equations for potential temperature and salinity. In addition, there is tracer diffusion along isopycnals, as in the Redi–Cox scheme. In contrast to the Redi–Cox scheme, however, no background horizontal diffusivity is needed in this parameterization scheme. Gent et al. (1994) show that, applied to a strong, sloping front, the G—M parameterization leads to a final state in which the front is flat, without a change in the vertical stratification. McWilliams and Gent (1994) demonstrate its effects in a nearly adiabatic wind-driven midlatitude ocean basin. Danabasoglu et al. (1994) show its effects on a coarsely resolved equilibrium global ocean solution; these include improvements in the distributions and fluxes of tracers and the locations of deep-water formation. Here we will present some preliminary results from a 1° CME run (N1-26.0) that incorporates this parameterization, with an isopycnal diffusivity coefficient of 1×10^7 cm^2 s^{-1}.

3. Results

Figure 1a shows the streamfunction of the zonally integrated volume transport for the 1° standard case (N1-12.0). It exhibits the familiar pattern of shallow overturning cells associated with the divergence (at the equator) and convergence (near 30°N) of the wind-driven currents in the surface Ekman layer. In the deep ocean there are two overturning cells associated with the formation and southward flow of North Atlantic Deep Water (NADW) (between 1200- and 3000-m depth) and northward flow of Antarctic Bottom Water (AABW) (below 4000 m). The relative strengths and depths of the two cells depend on the thermohaline boundary conditions at the surface and at the northern and southern boundaries (Holland and Bryan 1993; Böning et al. 1994; Döscher et al. 1994). Here we are not concerned with these dependencies but focus on the latitudinal structure of the NADW cell.

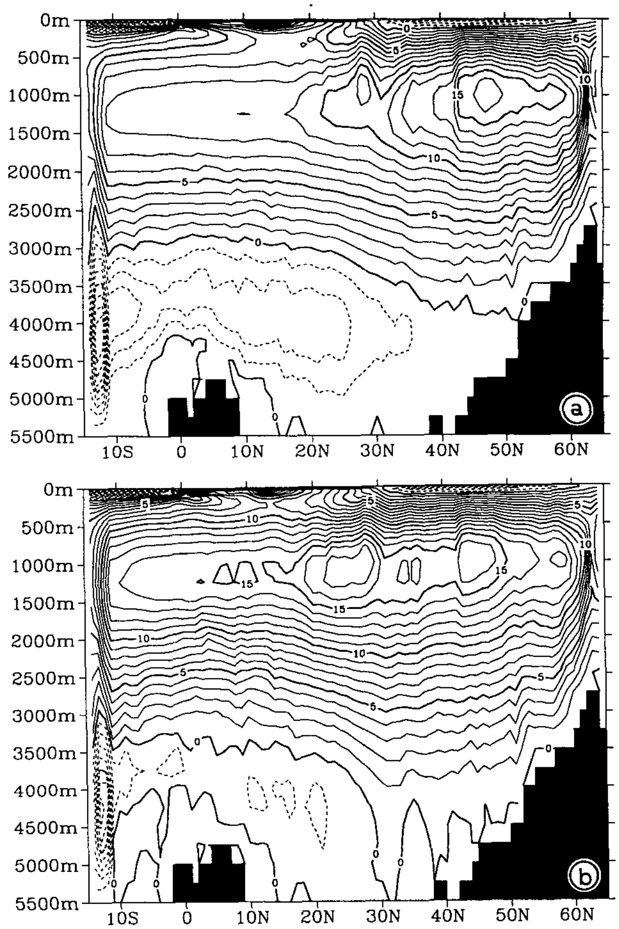

FIG. 1. Streamfunction (in Sv; 1 Sv = 10^6 m^3 s^{-1}) of the zonally integrated, time-mean flow in the latitude-depth plane for the 1° version of the CME: (a) reference case N1-12.0 with horizontal mixing, (b) case N1-26.0 using the parameterization scheme of Gent and McWilliams. Contour interval is 1 Sv.

The total production of deep water in this model case is 16.6 Sv. However, due to the strong upwelling in the midlatitudes between 30° and 40°N, only about half of the deep water formed in the subpolar North Atlantic finds its way across the equator to be transformed to upper intermediate and thermocline water in the restoring zone at the southern boundary. In the recent account of Schmitz and McCartney (1993), the formation rate of NADW in the subpolar North Atlantic, as inferred from a number of different observations, is 13–14 Sv (1 Sv = 10^6 m^3 s^{-1}); augmented by an upwelling and modification of AABW ($\sim$ 3 Sv), the net transport of deep water leaving the subpolar gyre is thought to be 16–17 Sv. An important aspect of their observational scheme is the close correspondence between the deep-water formation rate in the subpolar North Atlantic and the transports of the NADW cell farther south; that is, about 13 Sv of the Florida Current transport are of South Atlantic origin according to Schmitz and Richardson (1991), close to the meridional transport inferred by Rintoul (1991) for the South Atlantic at 30°S. Hence, the main discrepancy between the model result (Fig. 1a) and that observational scheme is not in the deep-water formation rate but in the meridional pattern of the NADW cell: the large upwelling found in the model, mainly between 30° and 40°N, leads to a significantly lower than observed overturning in the subtropical North Atlantic.

The outflow of the deep water formed in the subpolar North Atlantic takes place almost exclusively in the western basin, with the bulk of the transport in the deep western boundary current. The upwelling seen in Fig. 1a is also concentrated near the western boundary. Figure 2 shows the vertical velocity field at 1000-m depth. The vertical-velocity field reveals patches of strong upward motion exceeding 10^{-3} cm s^{-1}, especially in the vicinity of Cape Hatteras, while the velocities in the interior are small. Zonal cross sections near 35°N are given in Fig. 3, showing that the upwelling extends over a great depth range along the inshore edge of the Gulf Stream. The patterns shown in Figs. 2 and 3 are very similar to the model results discussed by Veronis (1975) and Cummins et al. (1990). Their analysis of the heat budget in the western boundary layer indicated a primary balance between hori-

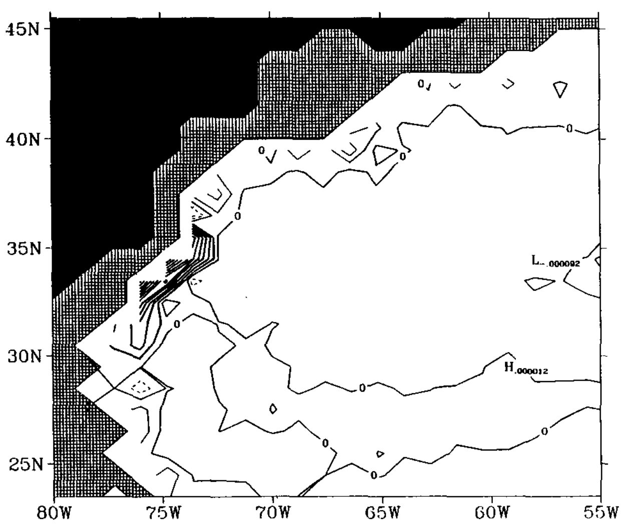

FIG. 2. Vertical velocity at 1000-m depth for model case N1-12.0. Negative (downward) contours are dashed. Contour interval is 10^{-3} cm s^{-1}.

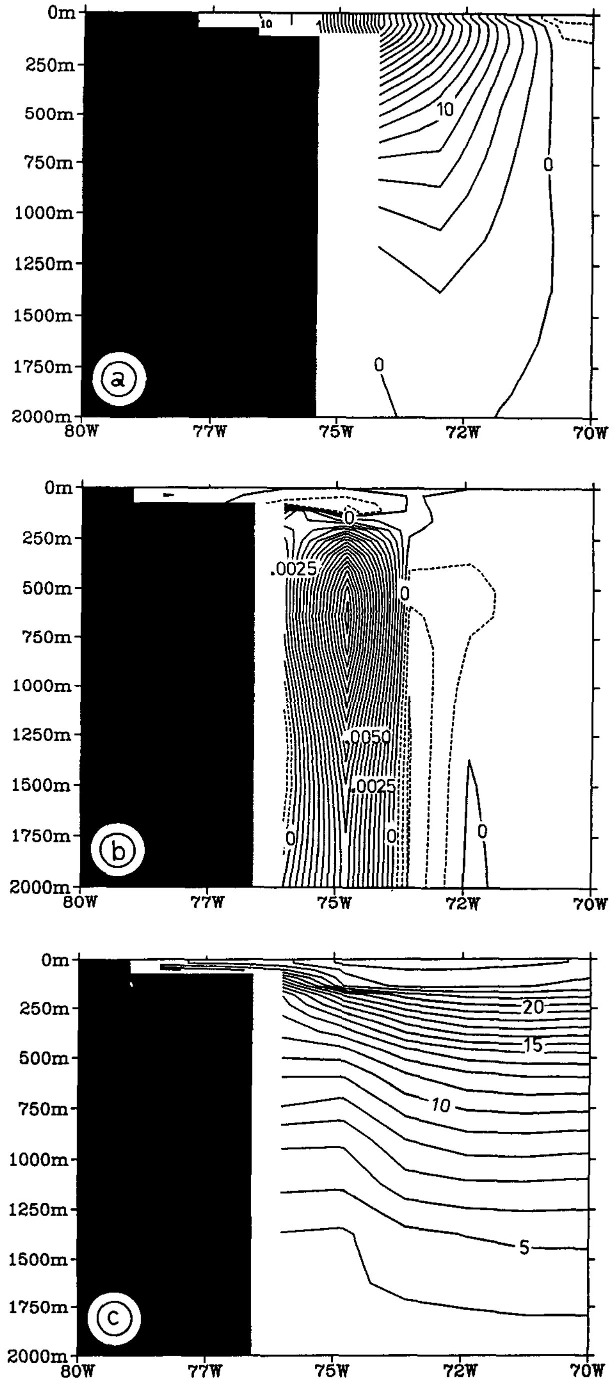

FIG. 3. Cross sections of (a) meridional velocity, (b) vertical velocity, and (c) potential temperature along 35°N near the western boundary. Model case N1-12.0.

zontal (i.e., cross-frontal) diffusion and vertical advection, $w\partial T/\partial z \sim K_h \partial^2 T/\partial x^2$. An estimate of the magnitude of terms in the advection–diffusion balance from Fig. 3 would give a vertical velocity scale ($W \sim K_h H/L^2$) somewhat exceeding 10^{-3} cm s^{-1}, indicating that even in this model case with a substantially higher resolution than in the models of Veronis and Cummins et al., horizontal diffusion across the Gulf Stream front must be a dominant term in the dynamical balance.

If the upwelling pattern in the meridional overturning is related to the local dynamics of the western boundary layer, one should expect to find a dependence on the magnitude of the Gulf Stream (corresponding to the frontal slope) and model factors that influence the cross-frontal diffusivity, that is, resolution and mixing parameters. In the following, we will compare different model cases in terms of their net vertical transports across 1000 m in the western boundary layer, obtained along a roughly 10° wide strip parallel to the boundary between 25° and 47°N. In all model cases, this measure captures the bulk of the midlatitude upwelling that is seen in the respective zonally integrated overturning patterns; that is, the vertical transports outside of this strip are rather small.

In Fig. 4, the upwelling for the different model cases is displayed against the northward transport of the Gulf Stream at 25°N, which, for a given resolution and mixing, may be thought of as a rough measure of the frontal slope in the western boundary layer. A striking feature of the model cases based on the same grid resolution and diffusion coefficients (i.e., the 1° cases with the horizontal mixing scheme, the 12.n cases) is that the upwelling appears to be well correlated with the northward transport of the Gulf Stream. This result holds regardless of whether the change in the northward transport is brought about by different thermohaline forcing (giving changes in deep-water formation and maximum meridional overturning) or by different wind forcing (and, thus, changes in the subtropical-gyre transport): it is the local strength of the Gulf Stream and, thus, apparently the local properties of the frontal region near the western boundary that appear to control the upwelling behavior.

The effects of changes in the model diffusivity and/or horizontal resolution are indicated by the dashed lines in Fig. 4. A substantial reduction of the horizontal diffusion coefficients can be achieved by an increase of the grid resolution. In the present high-resolution cases, a biharmonic scheme was adopted that allows a significantly smaller frictional influence on scales larger than the grid (Semtner and Mintz 1977). However, in any case the explicit diffusivity must be a term of leading order in the tracer balance on length scales comparable to the grid. Thus, two competing effects have to be considered when going to higher horizontal resolution: for a given length scale, say, the width of the front in the coarse-resolution reference case, the horizontal diffusivity becomes much smaller. However,

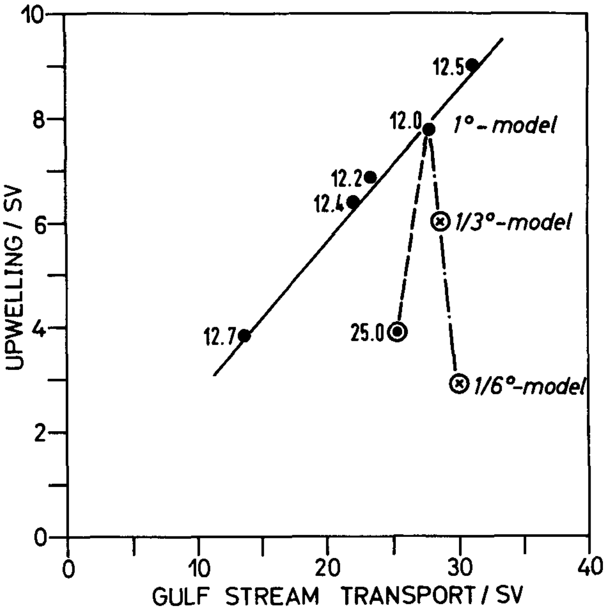

FIG. 4. Vertical transports at 1000-m depth in an approximately 10° wide strip along the western boundary between 25° and 47°N for different model cases plotted against the northward transport of the western boundary current at 25°N.

the simultaneously reduced horizontal viscosity leads to a sharpening of frontal features (as long as the grid width is not small compared to the local Rossby radius). The front in the 1/3° case is, as in the 1° case, not much wider than the grid spacing, and horizontal diffusion is still likely to play a dominant role on that scale. A drastic reduction of the upwelling is only achieved in the 1/6° case: in that case, where the grid spacing becomes clearly smaller than the local Rossby radius, the front is much better resolved by the model grid, and horizontal diffusivity may no longer be the controlling factor in the tracer budget on the frontal scale.

Two different ways of reducing the horizontal diffusion in the frontal region without changing the grid resolution are examined here: the use of a smaller ratio of A_h to A_m, and the use of isopycnal mixing schemes. In case N1-25.0, the horizontal viscosity A_m was increased from 1×10^8 to 4×10^8 cm^2 s^{-1}. The smaller horizontal diffusion due to weaker horizontal gradients, together with the somewhat reduced value of A_h (0.5×10^7 instead of 1×10^7 cm^2 s^{-1}), gives only a small change of the western boundary current transport but leads to a significant reduction of the upwelling compared to the reference case N1-12.0.

Only a small effect on the upwelling is found in the model versions that use the Redi–Cox isopycnal mixing scheme. As shown by Cox (1987), this scheme significantly improves the distribution of passive tracers in the main thermocline. Its usefulness in the western boundary layer is not clear, however. The model analysis of Cummins et al. (1990) indicated no qualitative change in the heat budget of that region compared to a horizontal mixing case. In the CME experiments we do not find a substantial influence of the Redi–Cox scheme on the pattern of meridional overturning.

A drastic change of the processes near the western boundary does take place, however, when the G—M parameterization (N1-26.0) is used. A thorough description of the model results, with that mixing scheme and an analysis of the mechanisms responsible for the different behavior, is beyond the scope of this paper, and we will restrict the discussion to changes in the zonally integrated streamfunction and its effect on the northward heat transport. Figure 1b shows only a minor change in the net deep water production in the subpolar North Atlantic compared to the reference case (Fig. 1a). However, in contrast to the horizontal mixing version, there is almost no loss of deep water due to upwelling in the midlatitudes, leading to a much larger overturning rate in the subtropics and Tropics. Almost 13 Svs of NADW reach all the way to the southern boundary, compared to only 8 Svs for the standard case.

The midlatitude upwelling and associated southward decrease in the overturning rate have a strong influence on the latitudinal structure of the poleward heat transport in the North Atlantic. Previous model evaluations have mainly been focused on comparisons with the observed heat transports from the transoceanic sections at 24.5°N (1.2 ± 0.3 PW), probably the most accurate direct estimate in the ocean because of the well-constrained western boundary current and low level of eddy energy at that latitude (Hall and Bryden 1982; Roemmich and Wunsch 1985). However, when looking for causes of the typically too low heat transport in ocean circulation models at that latitude, we feel it is important to also examine the latitudinal dependence.

Figure 5a shows that near 25°N, a large fraction of the net heat transport H is associated with the effect of the meridional overturning H_{OT}. As evident from Fig. 1, part of H_{OT} is due to the shallow Ekman cells, part due to the NADW cell. (The AABW cell can have only a minor effect due to the very weak vertical temperature differences in the deep ocean.) There is no simple way to separate the thermohaline and Ekman H_{EK} contributions to H_{OT}. However, we may note that H_{EK} must be zero at the equator and near 30°N; hence, the H_{OT} values of 0.40 and 0.47 PW at these latitudes can be considered as the pure effect of the deep thermohaline overturning cell. If we assume that there is only a slow latitudinal change in the effect of the deep overturning, its effect near 25°N should be close to the value at 30°N, that is, about 0.45 PW in the model case N1-12.0.

Figure 5b shows the net heat transport in several model cases in comparison to different observational studies: the results of Isemer et al. (1989) based on the surface heat flux, with an inverse model calculation of

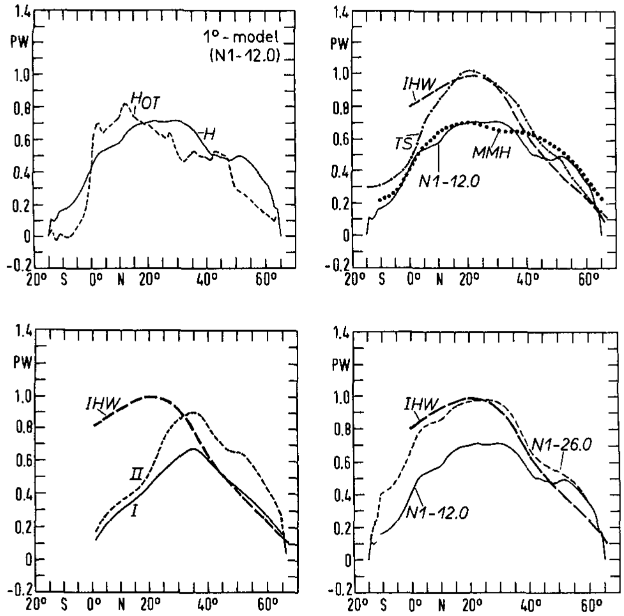

FIG. 5. Northward transport of heat for different model cases, compared to the observational patterns of Trenberth and Solomon (1994, denoted TS) and Isemer et al. (1989; denoted IHW). (a) Net heat transport H and contribution from overturning H_{OT} for the 1° standard case N1-12.0. (b) Net heat transports for N1-12.0 and the large-scale geostrophic model of Maier-Reimer et al. (1993; denoted MMH). (c) Net heat transports for 1/3° model cases using relaxation to climatological data near 65°N (I), and relaxation to actual section data (II) (from Döscher et al. 1994). (d) Net heat transports for the 1° standard case with horizontal mixing (N1-12.0) and a 1° case using the isopycnal mixing scheme of Gent and McWilliams (1990) (N1-26.0).

the bulk coefficients that took into account the directly observed heat transport near 25°N, and the recent results from Trenberth and Solomon (1994) based on the radiation budget at the top of the atmosphere and the atmospheric energy divergence. Despite the considerable error bars on these indirect calculations [± 0.4 PW at 25°N as given by Trenberth and Solomon (1994)], the latitudinal patterns are very instructive. Case N1-12.0 seems to have a fairly realistic net deep water production rate in the subpolar North Atlantic, and there are rather small differences in the northward heat transport at higher latitudes. However, the model–data deviation grows between 40° and 30°N, that is, in the latitude range where the spurious upwelling decreases the meridional overturning rate. A similar distribution can be seen, for example, in the model of Maier-Reimer et al. (1993). The result included in Fig. 5b is from a version of their model that has a maximum overturning of 22 Sv and, therefore, slightly higher heat transports in the subpolar area. But again, the model transports begin to diverge from the observed curves in midlatitudes to yield significantly lower than observed heat transports in the subtropical North Atlantic.

Sensitivity studies with the CME (Holland and Bryan 1993; Döscher et al. 1994; Böning et al. 1994) and other large-scale circulation models (e.g., Maier-Reimer et al. 1993; Washington et al. 1994) have demonstrated a strong dependence of the maximum overturning rate and the maximum northward heat transport on the specification of the thermohaline fluxes at the surface or open northern and southern boundaries. Those results imply that it should certainly be possible to obtain a better agreement between the simulated and observed overturning rates and heat transports at 25°N by increasing the deep-water formation rate in the subpolar North Atlantic. However, given the possibility of a deteriorating effect of the midlatitude upwelling pattern, it seems important to not just focus model validations on a single latitude but to examine the latitudinal pattern of the heat transport. Increasing the maximum overturning necessarily increases the northward transport of warm upper-layer water by the Gulf Stream system; hence, in a model with a given mixing parameterization, one should expect an increase in the western boundary upwelling, and thus an effect on the latitudinal pattern of the heat transport. An extreme example of such a tendency may be seen in the model solutions shown by Washington et al. (1994) in which a (artificial) change in the surface heat flux forcing produced a vigorous deep-water formation of more than 60 Sv in the subpolar North Atlantic; however, a large fraction of that (> 30 Sv!) upwells in the midlatitudes. As another less extreme example, Fig. 5c compares the heat transport patterns from two solutions with the 1/3° CME model: a case with restoring to smoothed climatological data in the northern buffer zone, and a case with a buffer zone based on hydrographic data from actual oceanic sections near 65°N (Döscher et al. 1994). In the latter case, the forcing and transport of the lower NADW is enhanced (the maximum overturning increases from about 12 to 15 Sv) and so is the maximum northward transport of heat in the basin. However, while the heat transport now exceeds the observational results for the subpolar ocean, it still exhibits a sharp decrease in midlatitudes.

The foregoing results suggest that an alternative way of improving the heat transport in the subtropical North Atlantic would be to decrease the artificial upwelling in the western boundary layer, as does occur with the G—M parameterization. The zonally integrated streamfunction of model case N1-26.0 showed about the same net deep water production in the north as the reference case N1-12.0 but very little zonally averaged upwelling in the midlatitudes (Fig. 1b). The heat transport of that case is shown in Fig. 5d. The changed overturning pattern leads to a significant enhancement of poleward heat transport south of 40°N and a quite close correspondence to the observed heat transport pattern in the North Atlantic.

4. Conclusions

While it has been demonstrated in a number of model studies that the maximum rate of meridional overturning and northward heat transport in the Atlantic Ocean are very sensitive to details of the thermohaline forcing in the high latitudes, the effect of spurious midlatitude upwelling of deep water on the meridional transport patterns has not received much attention. As was first noted by Veronis (1975), parameterization of eddy mixing of tracers by a horizontal diffusivity in ocean circulation models based on Cartesian coordinates in the vertical inevitably leads to diapycnal diffusion in the frontal region near the western boundary, inducing a strong upwelling of cold water and thus affecting the large-scale distribution of vertical velocity over the basin. Almost all z-coordinate models of the thermohaline circulation in the Atlantic Ocean exhibit strong zonally integrated upwelling in midlatitudes that significantly reduces the amount of deep water transported toward lower latitudes and across the equator. The main purpose of this paper is to direct attention to the fact that this shortcut in the ocean's "conveyor belt" affects not only the simulation of tracer distributions [as noted by Toggweiler et al. (1989)] but also the northward transport of heat across the equator and through the subtropical North Atlantic, and has to be considered as one of the main reasons for the typical failure of ocean circulation models to simulate the observed heat transport near 25°N.

The results from the CME show that the upwelling concentrated at the inshore edge of the Gulf Stream depends on a number of model factors. For given horizontal mixing parameters in a non-eddy-resolving model version, the upwelling almost linearly increases with the transport of the western boundary current, that is, due to an increased overturning rate. This makes it almost impossible to improve the simulation of poleward heat transport in the Tropics and subtropics by changes in thermohaline forcing. A larger deep-water formation rate certainly increases the maximum heat transport in the basin, but since a significant fraction of that water upwells in the midlatitudes, it also tends to distort the latitudinal pattern of the heat transport compared to the observed pattern.

In recent years there have been some advances in the development of ocean models based on layers of constant potential density (isopycnic coordinates) that, in principle, should not suffer from the problem described here (Bleck et al. 1992; Oberhuber 1993a). Their potential for the simulation of the large-scale thermohaline circulation has yet to be demonstrated, however, since conceptual difficulties arise due to the nonlinear equation of state and the presence of two dynamically active tracers (temperature and salinity). First results from two different isopycnal models for the North Atlantic show a too weak meridional overturning in the deep ocean (Oberhuber 1993b; New et al. 1994), but due to the unknown dependencies on various model factors (e.g., the unrealistic nature of the closed southern boundary), it appears still too soon to make definite statements about the relative merits of the different model types.

There seem to be basically two ways to reduce the artificial numerically induced upwelling of deep water in z-coordinate models. An obvious avenue is to use higher horizontal resolution, allowing smaller horizontal diffusivities. The disturbing lesson from the CME results presented here is that a very fine grid (1/6°) seems to be necessary to resolve the Gulf Stream front to an extent that diffusive processes become less important. At present it is not clear, however, which role the ratio of horizontal diffusivity to horizontal viscosity plays in these solutions. In particular, the 1/3° cases examined here had a A_h/A_m ratio of 2.5, while for the 1/6° case the ratio was 1. It is, therefore, hard to say how much of the improvement at 1/6° comes from the increase in resolution and how much comes from the reduction of the ratio. As in the 1° model (case N1-25.0 vs N1-12.0), the upwelling might be significantly reduced if one would choose a smaller ratio. These and related questions regarding the mechanism of the upwelling and the dynamical balance in the western boundary layer have to be examined further.

The second, at present probably the more promising avenue in the context of climate modeling is to adopt parameterization schemes for the effect of mesoscale eddies on tracer transport that eliminate, or at least minimize, diapycnal diffusivity induced in areas of sloping fronts. For reasons not understood yet, the isopycnal mixing scheme developed by Redi (1982) and Cox (1987) had only a small effect on the upwelling in the CME experiments. On the other hand, the preliminary results from the CME experiment using the parameterization developed by Gent and McWilliams (1990) indicated an elimination of the shortcut in the meridional overturning, leading to a dramatic increase of the northward heat transport in the lower latitudes. While the actual mechanisms of that effect have yet to be studied, the result clearly demonstrates the deleterious impact of the midlatitude upwelling pattern on the heat transport in the North Atlantic. Without the artificial upwelling, a realistic heat transport in the subtropics can be obtained without the need to excessively increase the deep-water formation rate in the subpolar North Atlantic.

Acknowledgments. We would like to thank Julianna Chow and Peter Herrmann for their assistance in the integration and analysis of the CME model. GD was supported by NOAA under Grant NRAZ000020115. CB was supported by the Deutsche Forschungsgemeinschaft. WH, FB, and JM were supported by the National Science Foundation through its sponsorship of the National Center for Atmospheric Research.

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