

Tracing the conveyor belt in the Hamburg large-scale geostrophic ocean general circulation model

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Abstract. The flow which constitutes the conveyor belt in the Hamburg large-scale geostrophic ocean general circulation model has been investigated with the help of a particle tracking method. In the region of North Atlantic Deep Water formation a thousand trajectories were calculated backward in time to the point where they upwell from the deep ocean. Both the three-dimensional velocity field and convective overturning have been used for this calculation. Together, the trajectories form a representative picture of the upper branch of the conveyor belt in the model. In the Atlantic Ocean the path and strength (17 Sv) of the conveyor belt in the model are found to be consistent with observations. All trajectories enter the South Atlantic via Drake Passage, as the model does not simulate any Agulhas leakage. Large changes in water masses occur in the South Atlantic midlatitudes and subtropical North Atlantic. Along its path in the Atlantic the water in the conveyor belt is transformed from Antarctic Intermediate Water to Central North Atlantic Water. On the average the timescale on which the water mass characteristics are approximately conserved is only a few years compared to the timescale of 70 years for the conveyor belt to cross the Atlantic. The ventilation of thermocline waters in the South Atlantic midlatitudes is overestimated in the model due to too much convective deepening of the winter mixed layer. As a result the fraction of the conveyor belt water flowing in the surface layer is also overestimated, along with integrated effects of atmospheric forcing. The abnormally strong water mass transformation in the South Atlantic might be related to the absence of Agulhas leakage in the model.

1. Introduction

Warm salty water is advected into the northern North Atlantic by poleward flowing currents. At high latitudes the net buoyancy flux is negative. As a result an increase in density takes place until the stratification becomes unstable and convection occurs, the surface water sinking to the deep ocean as North Atlantic Deep Water (NADW). The formation rate of NADW has been estimated with a variety of methods [Broecker, 1979; McCartney and Talley, 1984; Rintoul, 1991; Dickson and Brown, 1994] to lie between 12 and 20 Sverdrups ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$).

The study of the formation of a water mass involves the question of where the water originates. This question is related to the large-scale circulation that ventilates and replaces the waters of the deep ocean by surface waters. The large-scale circulation is determined by a number of processes which are still poorly understood, particularly vertical motion and the associated mixing. Direct measurement of these processes on the relevant spatial and temporal scales is not (yet) feasible. As a result it has not been possible to establish the sources of the water replacing NADW in a conclusive way. Up to now, answers have been sought by combining various observations from hydrographic data, current observations, and tracer measurements [e.g., Gordon, 1986; Broecker, 1991; Rintoul, 1991; Schmitz and McCartney, 1993; Schmitz, 1995]. In addition, modeling studies have been carried out to examine the processes of the water mass formation involved [e.g., Bryan and

Lewis, 1979; Cox, 1989; England, 1993; Döös, 1995]. However, it is still under debate whether this water upwells globally or more locally in the Southern Ocean as part of a pole-to-pole circulation in the Atlantic. It is also not clear whether the dominant transport into the Southern Atlantic is through Drake Passage or from the Indian Ocean.

Observational studies can only give partial insight into the thermohaline circulation associated with NADW formation, because the data sets are restricted. General circulation models of the World Ocean yield a more complete data set, but they may suffer from a lack of realism. One approach is to examine the conveyor belt in a coarse resolution model. The advantage of coarse resolution models is that they can be spun up to a steady state. Such models aim at a first-order simulation of the integral properties of the large-scale flow field. At present they do a fairly good job in describing both the overall thermohaline circulation and the gross distribution of the principal water masses [e.g., England, 1993; Maier-Reimer et al., 1993]. One disadvantage in these models is that (parts of) the large-scale circulation may be critically influenced by small-scale motions which are not resolved. These models do not represent details of ocean currents such as the structure of boundary currents, currents in passages, the equatorial currents, upwelling regimes, or eddies. All these details might play a critical role in determining the pathway of the conveyor belt.

A second approach is to study the conveyor belt in global scale eddy-resolving ocean models [e.g., FRAM Group, 1991; Semtner and Chervin, 1992]. These models have the advantage of resolving the flow in much more detail. They are very useful for evaluating physical processes and the dynamical balance of the flow field. However, integrations can only be carried out

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for a restricted time. One has to rely on a robust diagnostic technique [Toggweiler *et al.*, 1989] to accelerate the convergence in the deep ocean, and it is well known that this method severely degrades the resulting circulation. Eddy-resolving models can also be run in a prognostic mode. In that case the limited integration time does not permit the deep water mass properties to reach equilibrium with the surface forcing. As a result the transport of fresh water and heat is in neither case compatible with the observed atmospheric forcing. At present, observational studies and analyses using eddy-resolving and steady state coarse resolution models are complementary in contributing to our knowledge of the thermohaline circulation.

In this study we use the steady state of a low-resolution global ocean general circulation model (OGCM), the Hamburg large-scale geostrophic (LSG) model [Maier-Reimer *et al.*, 1993], to trace the path of the NADW return flow. One objective is to give a Lagrangian view of the large-scale flow field in the model, which is very well suited to tracing the water flow in the conveyor belt. A second objective is to analyze the integrated thermal and saline budget for the conveyor belt along its path in the Atlantic. The amount of water mass transformation and the role of convection in this process will be emphasized. In particular, we focus on the question of how, in the absence of Agulhas leakage, in the model the conveyor belt is consistent with both the observed atmospheric forcing and a realistic distribution of water masses [Maier-Reimer, 1993; Maier-Reimer *et al.*, 1993].

To this end the modeled three-dimensional velocity field is used to calculate the water movement. We consider water to consist of many water parcels and calculate the large- (relative to turbulent motion) scale movement of the water by tracking them. A similar method has been used previously, e.g., by Böning and Cox [1988], Fujio *et al.* [1992], and Döös [1995]. In addition, convective mixing is taken into account when calculating trajectories. The convection algorithm applied in the model introduces a random process with respect to individual water parcels and hence to the calculation of trajectories.

After a short review of the OGCM in section 2 an outline of the particle tracking method is given in section 3. A general description of the conveyor belt in the model is presented in section 4 and compared with existing observations. In section 5 we present the budgets of salinity and potential temperature for the conveyor belt along its path in the Atlantic. In section 6 the results of section 5 are interpreted, and the role of diabatic forcing and associated water mass transformation is discussed within the context of existing theories on the global conveyor belt. In section 7 the main results are summarized and we present our conclusions.

2. The Ocean Circulation Model

The Hamburg LSG ocean circulation model has evolved from a concept originally proposed by Hasselmann [1982]. This model is designed for climate studies. The concept is based on the observation that for the large-scale ocean circulation, the characteristic spatial scales are large compared with the internal Rossby radius. In addition, the characteristic timescales are large compared with the periods of gravity modes and barotropic Rossby waves. Since it is generally believed that these waves are not important for climate research studies, they have been filtered out of the model. The nonlinear advection of momentum has been excluded in the set of primitive equations, and the equations are integrated with an implicit method

which permits a time step of 30 days. This makes the model numerically efficient and permits integrations over thousands of years. The free surface is treated prognostically, without invoking the rigid lid approximation. The numerical scheme is unconditionally stable and can be applied uniformly to the entire globe.

Convective overturning is introduced whenever the stratification becomes unstable. The water column is stabilized with the minimum mixing compatible with the conservation of heat and salt: water in the thinnest layer is exchanged with an equal volume of water from the thickest layer. The water newly injected into this cell is then completely mixed with the residual water. This algorithm is applied successively to all pairs of layers, beginning at the top.

The standard global version of the model has a horizontal resolution of $3.5^\circ \times 3.5^\circ$ and 11 vertical layers centered at depths of 25, 75, 150, 250, 450, 700, 1000, 2000, 3000, 4000, and 5000 m. Topography is included, as is a thermodynamic sea-ice model with viscous rheology. For more details about the model we refer to Mikolajewicz and Maier-Reimer [1990] and Maier-Reimer *et al.* [1993].

The model is forced at the upper boundary by the monthly mean wind stress climatology from Hellerman and Rosenstein [1983]. The temperature in the surface layer is driven by a Newtonian type of coupling to prescribed monthly mean air temperatures derived from the Comprehensive Ocean-Atmosphere Data Set (COADS) [Woodruff *et al.*, 1987], with a coupling coefficient of $40 \text{ W m}^{-2} \text{ K}^{-1}$ (in the absence of sea ice). This yields a time constant of about 2 months for an upper layer thickness of 50 m. The salinity is driven with a similar Newtonian coupling to the observed climatological annual mean surface salinity from Levitus [1982] with a time constant of 40 days. (The seasonal data set appeared less reliable with respect to salinity due to the sparseness of the original data.) After 10,000 years of integration the model has reached an almost cyclostationary steady state. This standard run with the Hamburg LSG model described by Maier-Reimer *et al.* [1993] will be referred to as ATOS1. It should be noted that for ATOS1 the thermohaline surface forcing is tuned to reproduce the observed amount of NADW outflow and the distribution of passive tracers such as radiocarbon. This tuning consisted of varying the temperature boundary condition and the relaxation time constants for temperature and salinity.

General features of the resulting flow are presented in Figures 1a and 1b. The North Atlantic Current is clearly recognizable from the horizontal velocity field at 250-m depth (Figure 1a). It transports water poleward to the overturning regions. This figure also shows that such a circulation is absent in the Pacific. The velocity field at 2000 m shows the Deep Western Boundary Current, which transports NADW from the North Atlantic to the Southern Ocean (Figure 1b). A complete retroreflection of the Agulhas Current occurs (Figure 1a), which prevents any direct transport between the Indian and South Atlantic Oceans. Agulhas leakage is absent in all layers. A more detailed discussion of the absence of a direct link between the Indian and Atlantic Oceans and the consequences for the global thermohaline circulation will be given in section 4. The outflow of NADW at 30°S is a typical measure for the strength of the thermohaline circulation in the Atlantic. In the model this outflow is 17 Sv (see Figure 2), which lies well within the range of observational estimates.

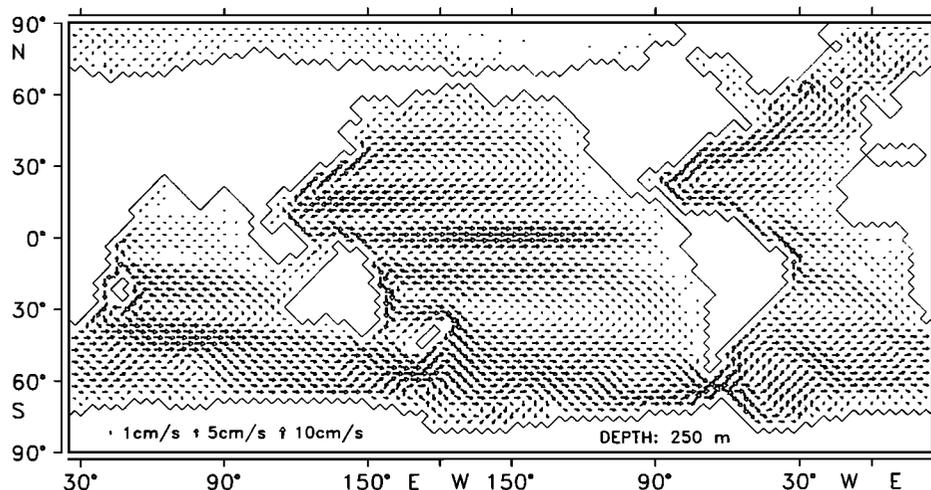


Figure 1a. Annual mean horizontal velocity at 250 m. The vector scaling is indicated within the figure.

3. Tracking Method

To calculate water movement, we consider it to consist of many individual parcels. Such a parcel has a finite volume, small enough to resolve the spatial variability of the large-scale circulation but large enough to average over small-scale turbulence. This means that we are not interested in the motion of individual water molecules but in the advective path of the large-scale circulation that is explicitly resolved in the OGCM.

The advective path or Lagrangian displacement of water parcels can be deduced from the three-dimensional velocity field by interpolating the velocity along the trajectories. The method has been described by Böning and Cox [1988], Fujio *et al.* [1992], and Döös [1995]. For the present model we have used the characteristics of the Arakawa E grid [Arakawa and Lamb, 1977] for the interpolation scheme. This grid configuration can be thought of as two separate, overlapping grids (“odd” and “even”) (Figure 3). Each odd grid box is partly overlapped by four even grid boxes and vice versa. At any given point a parcel can be located within both one odd and one even grid box. After interpolation within each grid box the velocities within the odd and even box are weighted by accounting for the

distance of the parcel to the respective centers of the boxes. A linear interpolation scheme has been used. The interpolation is performed within a scalar grid box, with the velocities positioned at the boundaries such that the scheme is consistent with the continuity equation applied in the OGCM.

Apart from linear interpolation, more sophisticated interpolation methods can be employed. The sensitivity to the interpolation method depends on the specific path of a trajectory, e.g., whether the trajectory transects a region where streamlines strongly diverge. However, we must emphasize that we are not interested in how well an individual trajectory is calculated but rather in how representative the statistics for an ensemble of trajectories will be. In general, the statistics for an ensemble of trajectories are much less sensitive to the interpolation method, and linear interpolation gives reliable results when the integration method used is sufficiently accurate [Böning and Cox, 1988].

When calculating the water movement from the three-dimensional velocity field, the water parcel is considered to be advected by the larger-scale motions only, its characteristics constantly being modified by diffusive mixing. In the OGCM

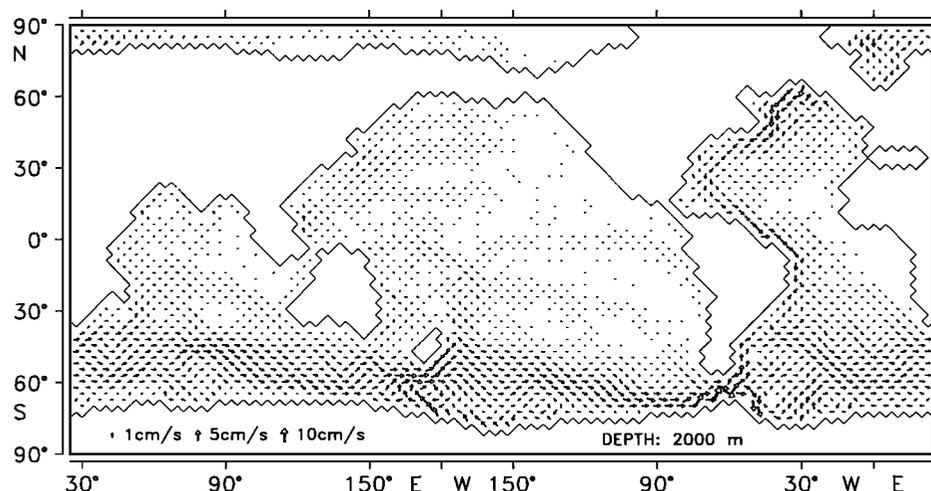


Figure 1b. Annual mean horizontal velocity at 2000 m.

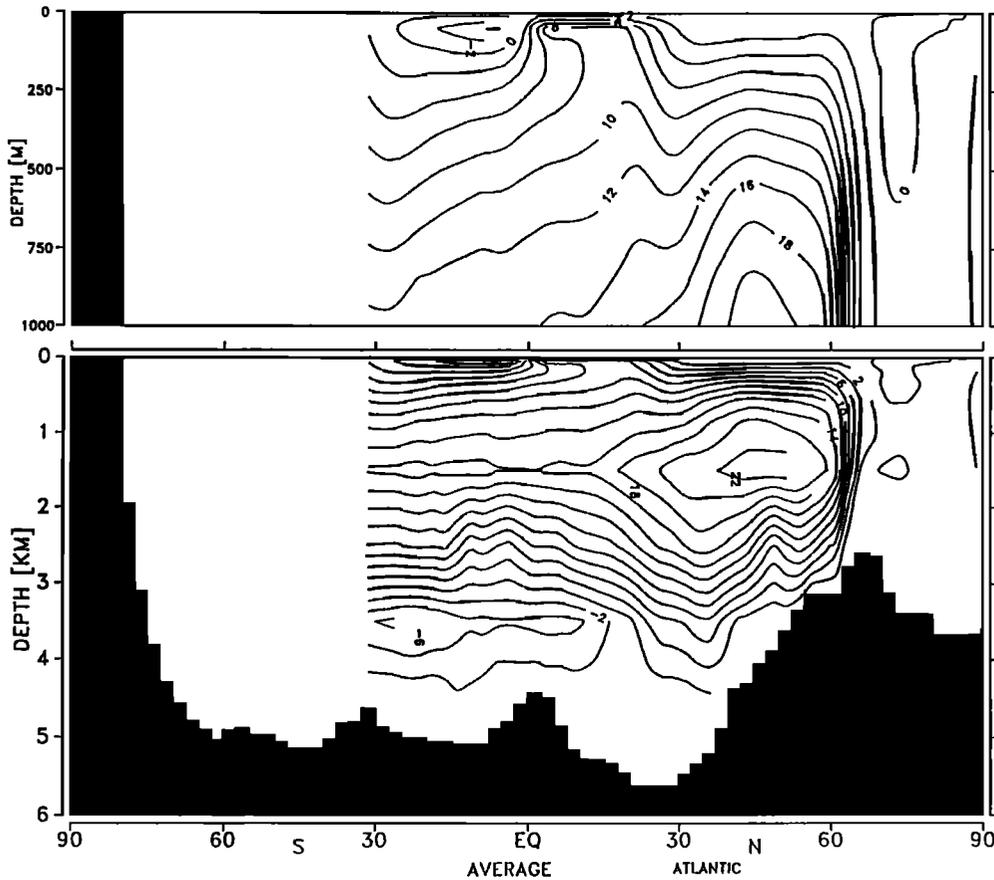


Figure 2. Zonally averaged Atlantic meridional overturning circulation in Sverdrups. Contour interval is 2 Sv.

one can separate the upwind scheme for tracer advection into an advective and a diffusive flux divergence with diffusion coefficient $|u| \delta x$ [Bacastow and Maier-Reimer, 1990]. The particle tracking method is consistent with the advective flux of the upwind scheme.

In addition to the displacement caused by the three-dimensional velocity field, computation of trajectories also includes convective overturning. In the particle tracking model we hypothesize that diffusive mixing influences the motion of

individual water molecules, but the center of gravity of a finite volume of water large enough to average over small-scale turbulence is advected by the averaged large-scale motion. Convection, however, may introduce a systematic displacement of an ensemble of water parcels. When a water parcel enters the top (bottom) of a convective column, it will be displaced on the average downward (upward), apart from the mean vertical velocity. As a result, convection adds an effective vertical motion to the three-dimensional velocity field on the scale of the basin depth. This motion is part of the large-scale thermohaline circulation. Considering vertical velocities only would lead to a severe underestimation of vertical water exchange, especially in regions of overturning. In the model the convection algorithm is a parameterization of various processes, i.e., large-scale subduction, deep water and mode water formation, and deepening of the winter thermocline and winter mixed layer. It can occur on horizontal scales of $O(1000 \text{ km})$.

When applied to water parcels, the convection algorithm used in the OGCM consists of a random process. As the resolution in the vertical is variable, convective mixing consists of the exchange of water between grid boxes with different volumes. For a parcel residing in the larger (lower) grid box and subject to convective mixing with the smaller grid box above, the chance of being displaced one grid box upward is equal to the volume ratio of the two grid boxes. A second element of chance results from the E grid configuration with overlapping grids. Since a parcel resides in two different grid boxes at the same time, it is possible that only one of the two boxes is subject to convective mixing. In that case the chance

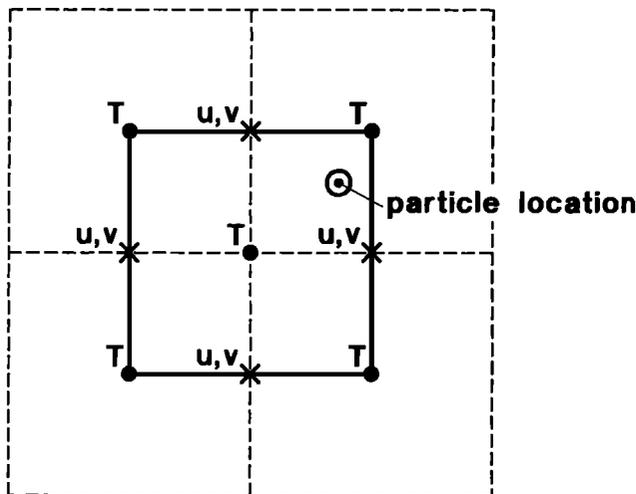


Figure 3. The grid configuration.

that the parcel will be subject to the random process of convective mixing is determined by its relative position to the centers of the two grid boxes involved. The random process is effectuated by letting the computer draw a number between zero and one. If the number is smaller than the calculated chance, a convective displacement is added. If the number is larger, the parcel remains where it is.

In our model the particle tracking method consists of backtracking trajectories from parcels that undergo deep convection in the northern North Atlantic. This backtracking implies a reversal of the three-dimensional velocity field and the convection algorithm. The velocity field is interpolated with a much smaller time step than the OGCM time step of 1 month. Convection is applied once per month rather than being interpolated in time. The reversal of the convection algorithm means that once per month it is checked whether convective mixing has occurred with the grid box above. If so, the random process described above determines whether the parcel will be displaced to the grid box above. If the computer draws a number which is larger than the chance for displacement, the procedure stops. If the computer draws a smaller number, the procedure is repeated with the next pair of grid boxes. Ultimately, within one convection event the parcel may end in the uppermost grid box of the convectively unstable column. If the parcel is not displaced upward at all, it is checked whether convective mixing has occurred with the grid box below. In that case the parcel is moved one grid box downward. After a convective displacement the relative position in the vertical within the new grid box is chosen at random by letting the computer again draw a number between zero and one. The parcel keeps its position in the horizontal plane. This is motivated by the observation that convection in the ocean takes place on a much smaller scale than the grid size of the model ($5^\circ \times 5^\circ$).

In the computation of trajectories we resolve the seasonal cycle with respect to velocity fields and the occurrence of convection. This makes an analytical integration method for calculating trajectories impracticable. Such a (far more efficient) method is only feasible if one uses a constant (averaged) velocity field and neglects convective displacements [Döös, 1995]. In the present model the time integration has been carried out with a finite difference method: a fourth-order Runge-Kutta scheme.

We have tested the integration method by calculating trajectories for an idealized two-dimensional velocity field with closed (circular) trajectories. The velocity was chosen to have a typical value of 0.1 m s^{-1} ; the spatial scale was 1000 km. The time step for the velocity interpolation in the trajectory model was reduced from 1 month to successively smaller values until no difference could be discerned between the trajectory after 100 loops and after one loop. For this velocity field it turned out that an interpolation time step of a half day is sufficient for the integration of the trajectories.

Since a no-slip condition has been imposed on the lateral wall, parcels slow down as they approach the boundary. This can result in parcels that stop just on the boundary. In addition, the Runge-Kutta method can "overshoot" the parcel displacement in the normal direction when the curvature is large, resulting in parcels ending on land. The problem can be solved by defining a small boundary layer near coasts where the velocity component normal to it is set at zero in the case of converging streamlines [Fujio *et al.*, 1992]. In the present model the boundary layer has been chosen to be as small as

possible without trajectories ending at land-sea boundaries. A boundary layer with a width of 2% of the grid box size ($O(10 \text{ km})$) is sufficient to prevent this.

4. The Conveyor Belt

Introduction

The conveyor belt is a schematic concept for the thermohaline circulation associated with NADW formation. This concept was developed by Broecker [1991], partly inspired by the work of Gordon [1986], who first proposed a pathway for the NADW return flow in which all oceans communicate with each other. Gordon [1986] argued that the heat transport from the Southern Ocean into the Atlantic would originate from the Indian Ocean, following the "warm water path." Broecker [1991] adopted the warm water path into his concept, but in the same article he questioned the importance of the warm water path and developed arguments for the importance of the "cold water path," a heat and freshwater transport from the Southern Ocean into the Atlantic which originates from Drake Passage and the Antarctic Circumpolar Current (ACC). Subsequently, a debate has developed on the importance of the cold water path versus the warm water path [e.g., Rintoul, 1991; Gordon *et al.*, 1992].

Part of the interest in this controversy resides in the fact that coarse resolution OGCMs and coupled climate models do not simulate the warm water path, since the Indian-Atlantic transfer is determined by smaller-scale processes. On the other hand, global eddy-resolving models suggest a major contribution of the Indian-Atlantic exchange to the global conveyor belt [Semtner and Chervin, 1991; Döös, 1995]. Mesoscale eddies play a large role in Indian-Atlantic exchange [Lutjeharms and Gordon, 1987]. Model calculations show that individual eddies might transport significant amounts of heat and salt [Drijfhout, 1990]. This is confirmed by observations on Agulhas eddies [Gordon, 1985]. However, the water mass transport by the eddies might be offset by the transport of an eddy-induced mean circulation [Drijfhout, 1994]. This would result in a circulation which connects the Indian and South Atlantic basins but has only a net weak exchange of water mass characteristics. Such a circulation has been proposed by Rintoul [1991] and Gordon *et al.* [1992]. It is also in agreement with model calculations [De Ruijter and Boudra, 1985; Cai and Greatbatch, 1995].

It should be stressed that the present study is not aimed at closing the discussion on the importance of the Indian-Atlantic exchange. Since both the amount of NADW formation and NADW outflow from the Atlantic into other oceans are realistic in the model, it is suggested that the cold water path in the model is consistent with both the observed atmospheric forcing and a realistic distribution of the principal water masses [Maier-Reimer, 1993; Maier-Reimer *et al.*, 1993]. However, if the warm water path is important in reality, in coarse resolution OGCMs a conveyor belt following the cold water path would be associated with too much water mass transformation in the Atlantic. If, on the other hand, the cold water path is dominant, the transformation of conveyor belt water would be consistent with observations.

With the particle tracking model we expect to visualize and to describe in more detail the Atlantic conveyor belt hypothesized within an Eulerian framework by Maier-Reimer *et al.* [1993]. In the Atlantic its main characteristic is expected to be the dominant role of the western boundary currents. Since the

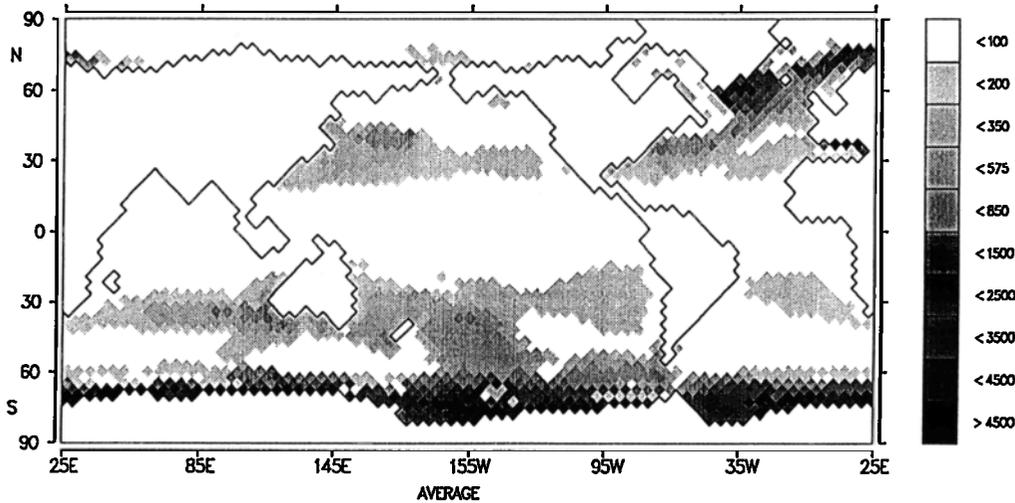


Figure 4a. Depth (meters) of the convective column produced by surface cooling, defined as the greatest depth connected with the surface by at least one convection event within an annual cycle.

ratio of NADW outflow at 30°S to the maximum production rate is roughly 70%, about 70% of the trajectories should originate from oceans other than the Atlantic. In particular, we will compare the pathway of the model conveyor belt within the Atlantic to the scheme of Broecker [1991] and evaluate to what extent the model conveyor belt implies a communication between the various oceans and how the associated upwelling is distributed. A discussion on the thermal and saline balance of the conveyor belt is presented in the next section.

Method of Calculation

To start the numerical integration of trajectories, a set of initial positions which is representative for the model’s conveyor belt is required. To obtain such a set, parcels were released in the North Atlantic Current between the surface and 575-m depth. Trajectories were calculated forward in time. Some trajectories remained in the surface layers; others followed a closed loop in the Atlantic, and still others were advected out of the Atlantic by the Deep Western Boundary Current after joining deep convection in the north. Of these latter, 100 trajectories have been selected, from each of which

a point prior to the occurrence of deep convection was taken as the initial position for the backward integration.

Between inflow into a region of convection activity and outflow from that region a parcel “oscillates” many times between the surface and the deep ocean, subject to a long sequence of successive convection events. Starting the backward integration after convection occurred would only yield a set of trajectories representative for the model’s conveyor belt if the time integration would be prolonged to several thousands of years. As an alternative the initial positions were generated by randomly perturbing the 100 selected points just prior to the occurrence of deep convection by a maximum value of 0.5° in the horizontal and 1/10 of the depth of the grid box. After randomizing, 1000 initial positions were chosen. All were located in the upper 600 m near the southern boundary of the region of deep convection in the North Atlantic (compare Figure 4a and 4b).

Starting from these initial positions the trajectories were integrated backward in time until the depth of the parcel was larger than 1500 m. This was assumed to be the point where

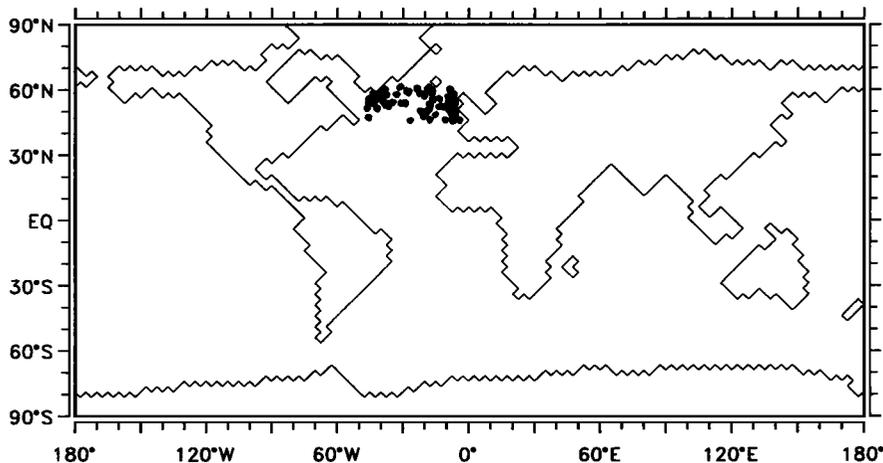


Figure 4b. Initial positions for (backward) trajectory calculation.

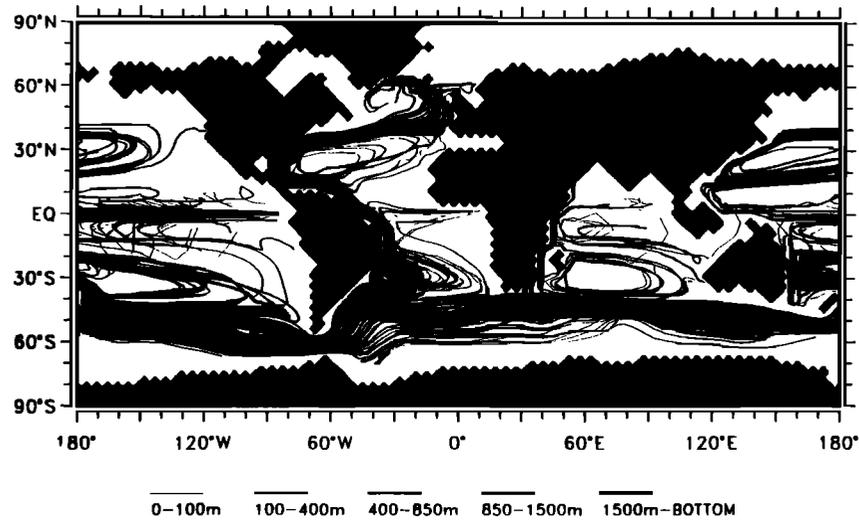


Figure 5. Forty trajectories representing the upper branch of the conveyor belt in the model. The thickness of the line indicates the depth of the trajectory.

the water parcel upwells from the deep ocean. Since middepth convection occurs in large parts of the North Atlantic, nearly all water parcels occasionally resided beneath the 1500 m. For this reason the “upwelling criterion” was relaxed to a level of 2500 m there. When after 500 years of integration time the parcel still resided in the upper 1500 m, the source region was attributed to the end point after 500 years. This was the case for 20% of the parcels. With respect to the source regions the distribution appeared to be insensitive to whether the parcel actually upwelled or not.

Results

From the computed set of trajectories, 73.1% enter the Atlantic from the south via Drake Passage and 26.9% are upwelled within it or enter from the north. We conclude the set of trajectories obtained to be representative for the model's conveyor belt. A representative subset of 40 trajectories is presented in Figure 5. This figure shows a far more complicated circulation than the visual concept of Broecker [1991]. Both circulation patterns, however, have features in common. Most important is the presence of a global thermohaline cir-

ulation that connects the various ocean basins, with the dominant pathway along the western boundary in the Atlantic and the large-scale sinking associated with NADW formation. The most important differences are the cold water path and the prominent roles of the wind-driven gyres and the ACC in constituting the conveyor belt. The wind-driven gyres and equatorial currents and countercurrents make some of the trajectories in Figure 5 rather complicated. However, a main path of the model's conveyor belt is easily distinguished.

The upwelling sites where the return flow originates are found throughout the World Ocean (Figure 6); larger concentrations appear at the northern edge of the Gulf Stream, north of the Weddell Sea, in Drake Passage, and in the Mozambique Channel. The upwelling is almost uniform. Dividing the World Ocean into the major basins and summing up the numbers of upwelling points yield the following distribution of the upwelling: 1% in the Arctic, 20.5% in the North Atlantic, and 5.4% in the South Atlantic. This percentage of NADW (26.9%) takes no part in the outflow into other ocean basins. The 73.1% of NADW that supplies the outflow can be divided into 13.6% that upwells in the Indian Ocean, 0.8% in the

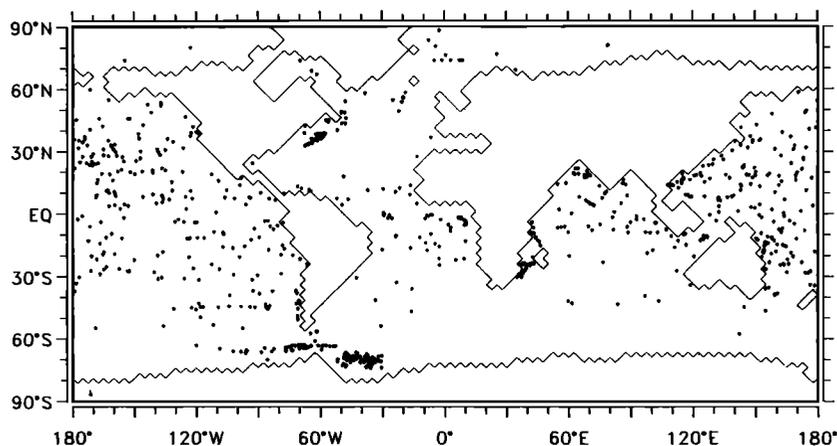


Figure 6. End positions for (backward) trajectory calculation.

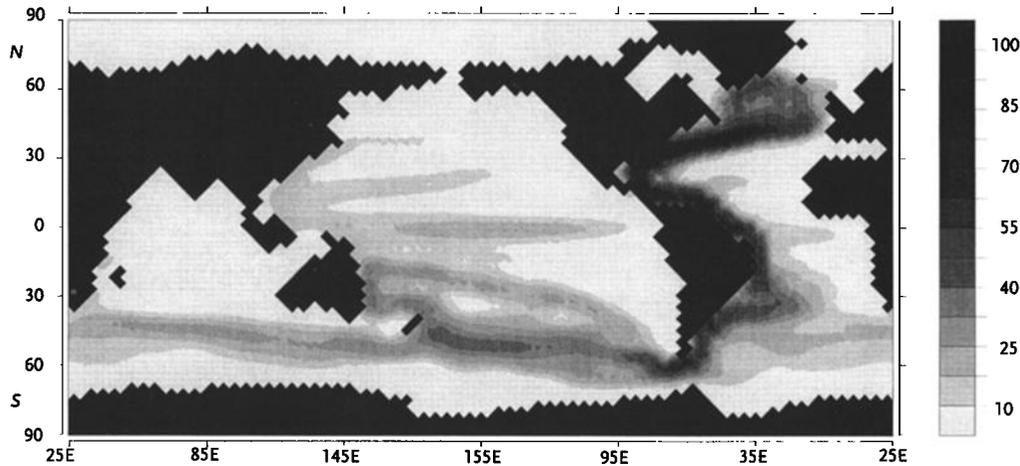


Figure 7. Percentage of trajectories passing through a grid box of $5^\circ \times 5^\circ$.

Indonesian Seas, 22.5% in the North Pacific, 20.6% in the South Pacific, and 15.6% in the Southern Ocean.

Figure 7 summarizes the model's conveyor belt by showing the number of trajectories that transect a grid box. It shows the existence of a main path for the conveyor belt. Although one might argue that the flow in coarse resolution models does not have enough degrees of freedom to organize itself otherwise than in well-defined circulation patterns, it can be deduced from Figure 5 that individual trajectories can be very sensitive to small changes in initial position and that it is not obvious beforehand whether an ensemble of such trajectories has an organized character. It is seen in Figure 7 that the western boundary currents in the Atlantic and the ACC play a major role in determining the main path. On the average, parcels that enter the South Atlantic have made one (1.1) extra loop around Antarctica, advected by the ACC. The average number of extra loops is independent of the region where the parcel upwells. The ACC acts as a blender mixing the various characteristics (S, T) to rather uniform values. Figures 5 and 7 also reveal that water entering the South Atlantic follows not only the western boundary but also the path of the anticyclonic subtropical gyre. The timescale associated with the advective path in the Atlantic is about 70 years. For the whole upper branch of the conveyor belt it is about 350 years.

Only 5.4% of the trajectories go through the Pacific-Indian Throughflow (PIT), 13% of the parcels that upwell in the Pacific. This is consistent with the weak transport through the Indonesian Seas, about 2 Sv. The PIT transport is thought to be larger in reality, although there is still much uncertainty about this feature [e.g., *Hirst and Godfrey, 1993*]. An unpublished version of the model with slightly altered topography in the Indonesian Seas yields a much larger PIT (15 Sv). In that case, about 50% of the parcels which upwell in the Pacific join the PIT (7 Sv, or 40% of the PIT). Eventually, they all merge with the Agulhas retroflection and, later on, with the ACC. An increase of the PIT causes an increase of the travel path of this water before it merges with the ACC. The salinity is increased by the exposure to evaporation in the Indian Ocean. However, the travel path within the ACC is also increased. This compensates the higher salinity of water entering the ACC by a longer exposure to regions with freshwater gain. In the model the PIT influences the path of the conveyor belt only between

the upwelling site and the merging point with the ACC. It increases the role of the ACC as blender.

5. Temperature and Salinity Budgets in the Atlantic

In this section we calculate the change in water mass characteristics of the model's conveyor belt in the Atlantic. The trajectories apply to water parcels with a finite volume. Along the advective path the parcel will change its characteristics by diabatic processes and by exchanging water molecules with its surroundings. Advection itself will not change the parcel's identity. As a result we can apply the model's budget equations for tracer concentration Q along the advective path of the water parcel:

$$\frac{dQ}{dt} = F_{\text{mix}} + F_{\text{con}} + F_{\text{air}} \quad (1)$$

where F denotes a forcing term which can be divided into three different contributions: one by small-scale mixing, denoted by mix; one by convection, denoted by con; and one by air/sea interaction, denoted by air.

It should be noted that in (1), F_{mix} is a residual term to close the budget; it is very difficult to calculate this term explicitly. F_{mix} consists of many separate contributions: explicit horizontal and vertical mixing, the effective diffusion of the upstream scheme, and numerical dispersion, as well as spatial and temporal interpolation errors in the trajectory calculation.

The budget calculations reveal that a significant part of the model's conveyor belt consists of intermediate water. At Drake Passage the conveyor belt flows at an average depth of 300 m. The average temperature and salinity are 4.4°C and 34.5 parts per thousand (ppt). Between the Drake Passage and 30°S the water gradually downwells. However, part of it remains in the upper layers (21% in the upper 200 m compared to 43% at Drake Passage). This water undergoes increases in temperature and salinity caused by atmospheric heating and evaporation (Figures 8a, 8b, 9, and 10). Temperature and salinity are also increased by diffusive mixing.

At 30°S the average temperature and salinity have increased significantly, 8.5°C and 35.0 ppt. The average depth is 500 m,

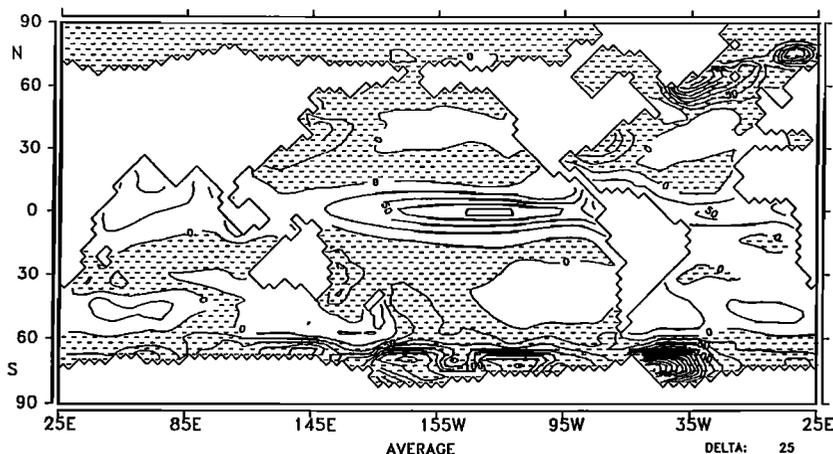


Figure 8a. Heat flux in watts per square meter as determined diagnostically from the temperature boundary condition. Positive values indicate a flux from the atmosphere into the ocean. Contour interval is 25 W m^{-2} .

and 75% of this water has a temperature between 9°C and 3°C and is intermediate water; 10% flows beneath the 850-m level. In the model the heat flux across 30°S is even less than Rintoul's [1991] estimate of 0.13 PW.

Between 30°S and 8°N the conveyor belt water is only subject to modest changes. At 8°N , temperature and salinity have increased to 8.8°C and 35.1 ppt. Although the proportion of surface water above 9°C has increased to 30%, the average depth of the conveyor belt remains the same. The conveyor belt is spread along the vertical. Evaporation increases salinity within the surface layer up to 3°S , after which the surface waters gain fresh water. This is largely compensated by mixing with more saline water (Figure 9). Atmospheric heating increases the temperature and compensates for the mixing with colder waters, especially between 10°S and 5°N (Figure 10).

A large change in water mass characteristics occurs in the subtropical North Atlantic. Between 8°N and 24°N the surface waters are heated and evaporation takes place. At 24°N , 70% of the conveyor belt water has a temperature above 9°C . The average depth has decreased to 300 m; temperature and salinity have increased to 12.2°C and 35.7 ppt. The largest contribution to the diabatic forcing in this region comes from diffusive mixing (Figures 9 and 10). The strong mixing is reflected

in the disappearance of the low-salinity tongue associated with Antarctic Intermediate Water (AAIW) [see Maier-Reimer *et al.*, 1993, Figures 7e and 7f].

North of 23°N the conveyor belt is cooled. The boundary of zero freshwater gain lies further north (Figures 8a and 8b). North of 30°N the conveyor belt also becomes fresher, but this is mainly accomplished by mixing. Up to 47°N , net evaporation partly counteracts mixing. The temperature decrease of the conveyor belt between 24°N and 58°N is 6.7°C . This is almost as large as the increase between Drake Passage and 24°N , 7.8°C . The salinity decrease between 24°N and 58°N , however, is much less, 0.4 ppt, compared to an increase of 1.15 ppt between Drake Passage and 24°N . The diabatic forcing increases the density of the conveyor belt along its path in the Atlantic.

The contribution of convection in changing water mass characteristics of the conveyor belt is rather small (Figures 9 and 10). However, convection plays indirectly a much larger role in the temperature and salinity budgets. Due to convection, a larger fraction of the conveyor belt flows in the surface layer and is subject to air/sea interaction. We have repeated the trajectory calculation with the convection shut off. In that case the fraction of the conveyor belt which flows in the surface

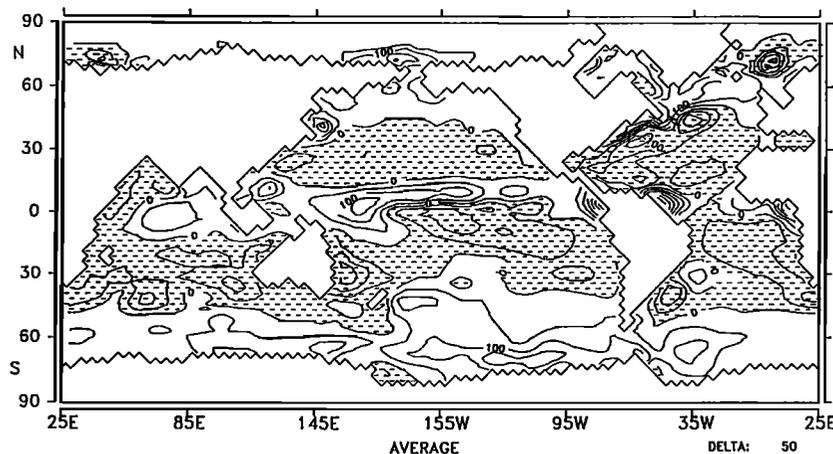


Figure 8b. Freshwater flux into the ocean in millimeters per month as diagnosed from the salinity boundary condition. Contour interval is 50 mm month^{-1} .

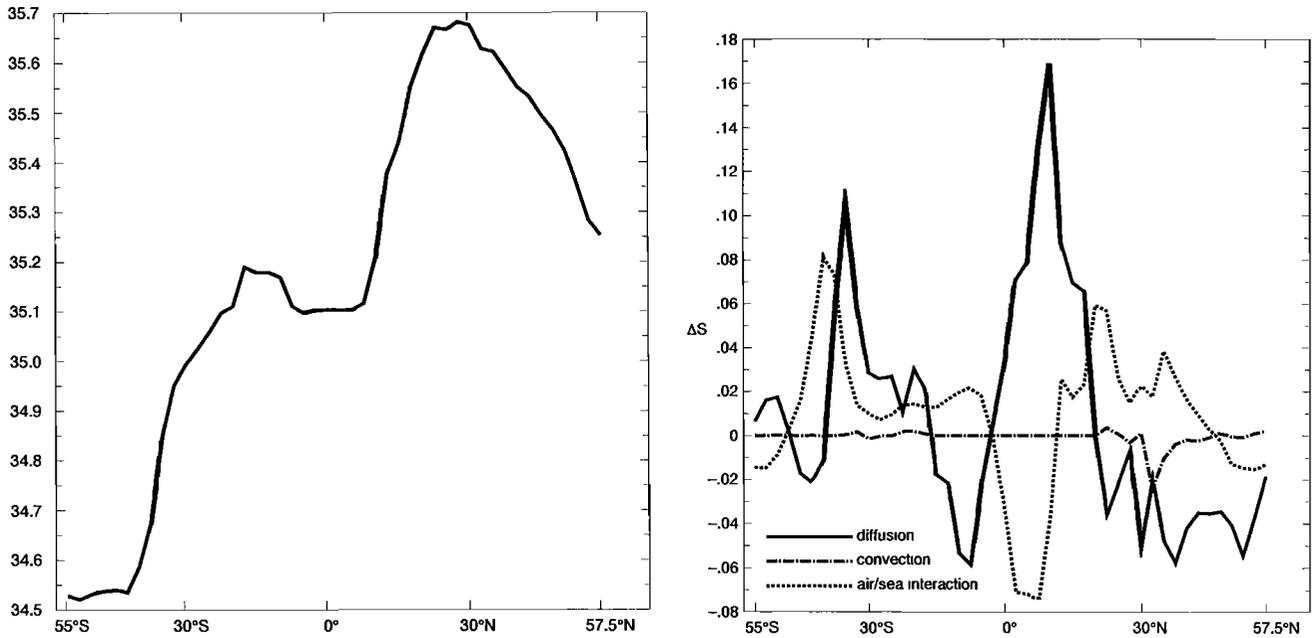


Figure 9. (left) Average salinity in parts per thousand for the conveyor belt in the Atlantic as a function of latitude and (right) the integrated salinity change within zonal bands of 2.5° latitude.

layer is decreased by 35%. Also, the percentage of the trajectories leaving the Atlantic decreases from 70 to 50%. Convection enhances the conveyor belt. It increases the amount of conveyor belt water flowing in the surface layer. As a result the conveyor belt is more concentrated within the fast western boundary currents. Also, forcing at the surface becomes more important. When it flows through the Atlantic, convection causes a larger increase in temperature and salinity of the conveyor belt water. The average number of convection events (with a time step of 1 month) for each trajectory between upwelling and entering the region of deep convection is 67.

This implies that the timescale on which diabatic effects can be neglected for the upper branch of the conveyor belt is only a few years, compared to a circulation timescale of $O(100)$ years).

6. Discussion

When *Gordon* [1986] proposed the warm water route as the dominant path for the conveyor belt, the conjecture was based on the mean temperature of the Sverdrup transport calculated from observations on mass and heat transport by the Brazil

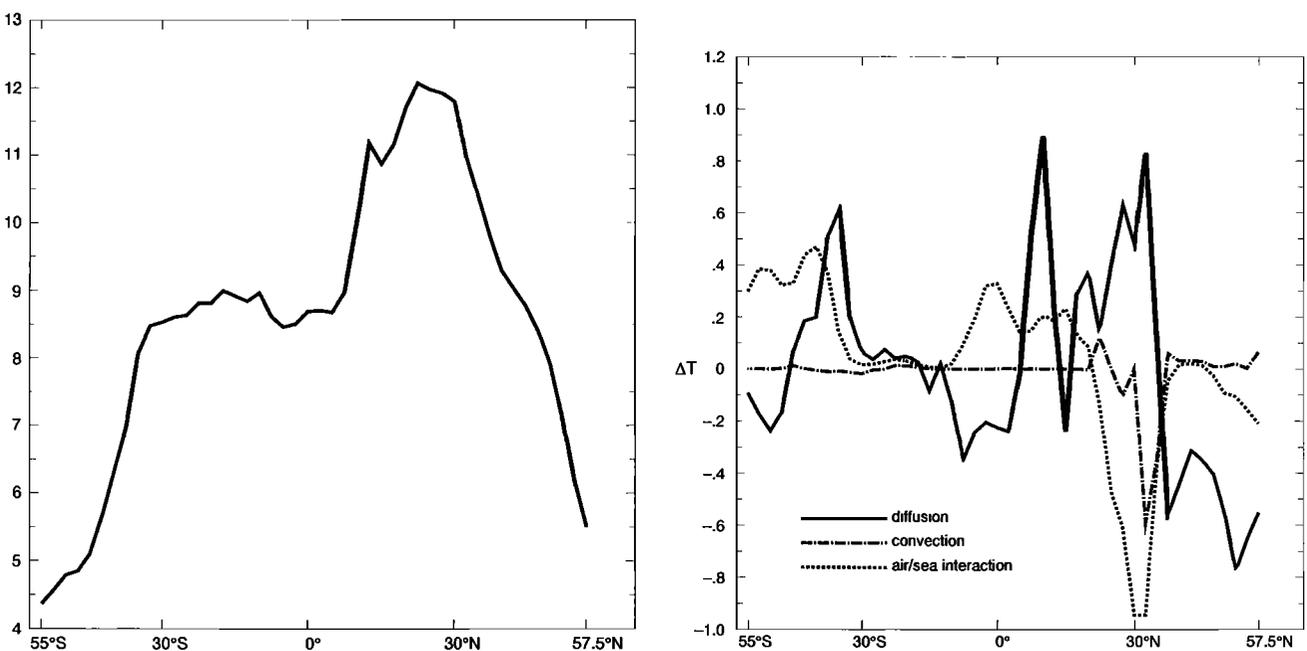


Figure 10. Same as Figure 9, but for potential temperature in degrees Celsius.

Current and NADW outflow and on measurements of the northward heat flux across 30°S in the Atlantic. He arrived at a mean temperature for the return flow of about 15°C and concluded that the conveyor belt resides in the thermocline waters. The temperature of the conveyor belt water was too high for it to have originated from Drake Passage. This conclusion was questioned by *Rintoul* [1991], who arrived at a much lower heat flux across 30°S: 0.25 PW compared to 0.69 PW [*Hastenrath*, 1982], with an inverse model calculation. *Rintoul* concluded that a significant amount of the conveyor belt at 30°S consists of intermediate water (colder than 9°C). This conclusion was confirmed by observations of *Gordon et al.* [1992].

In the model the conveyor belt follows the scheme proposed by *Rintoul* [1991], the cold water path. The differences between the model's conveyor belt and the scheme proposed by *Gordon* [1986], which involves the warm water path, can be attributed mainly to two features. First, in the model the heat flux at 30°S in the Atlantic is much less than the value used by *Gordon*. As discussed previously by *Rintoul* [1991], the value used by *Gordon* is probably too high. On the other hand, the heat transport in the LSG model is rather weak (see *Maier-Reimer et al.* [1993] for a discussion). The low heat transport in the model at 30°S in the Atlantic might be associated with the absence of an Agulhas leakage. Furthermore, *Gordon* might have underestimated the role of diabatic forcing. In the model the temperature difference between the upper branch of the conveyor belt and the NADW outflow has become twice as large at 30°S in the Atlantic compared to the entry point at Drake Passage. At least in the model the NADW return flow cannot be traced by the distribution of water mass characteristics in the thermocline.

If in reality the warm water path is important, the model has to compensate for the anomalously cold and fresh conveyor belt water from Drake Passage by an anomalously large diabatic forcing to increase the temperature and salinity of the conveyor belt water in the Atlantic. Prior to the occurrence of deep convection in the North Atlantic, a large increase in the temperature and salinity of the conveyor belt occurs in two regions in the model, in the South Atlantic midlatitudes between 25°S and 40°S and in the subtropical North Atlantic between 10°N and 25°N. In the following we will try to estimate how diabatic forcing of conveyor belt water is achieved in these two regions and whether or not the magnitude of the forcing is overestimated.

In the subtropical North Atlantic, diabatic forcing is mainly associated with diffusive mixing. Diffusion may be overestimated by the upstream scheme. However, due to the upstream scheme, gradients are underestimated. An enhanced role of diffusive mixing must be associated with too many trajectories crossing the front between cold and fresh intermediate water and warm and salty subtropical thermocline water.

The partition into different water masses of the conveyor belt in the Florida Straits has been estimated by *Schmitz and Richardson* [1991]. They argue that at 24°N about 55% of the conveyor belt resides in the upper 100 m and about 40% consists of water colder than 12°C. In the temperature range between 12°C and 24°C, nearly all the water should be of North Atlantic origin. In their analysis, *Schmitz and Richardson* assume that water mass characteristics are nearly conserved. The model estimates support *Schmitz and Richardson's* estimates. The main difference is that the model displays an additional 5 Sv in the temperature range between 12°C and 24°C, which is

of South Atlantic origin. This South Atlantic contribution to the Florida Current is masked by its TS characteristics through diffusive mixing while it recirculates in the subtropical gyre (Figure 5). The recirculating part of the conveyor belt can only be traced by a Lagrangian method.

If diabatic forcing of conveyor belt water in this region is too large in the model, it should be associated with the recirculating part in the North Atlantic subtropical gyre. We cannot directly judge whether 5 Sv of recirculating conveyor belt water in the model is too large. The increase in heat transport in this region, 0.2 PW, is, however, consistent with observations and eddy-resolving models [e.g., *Böning and Herrmann*, 1994].

The diabatic forcing of the recirculating part is enhanced by convection. The mean vertical velocity is downward here but overcompensated by a large convective upward displacement. Since the core of the recirculating water is at the bottom of the convective column, the net effect of convection is to move parcels upward. This can be interpreted as the process of capturing thermocline water into the deepening winter mixed layer. In the model the depth of convection within the subtropical gyre in March is in agreement with observations of the depth of the mixed layer in this region [*Levitus*, 1982]. There is no indication that convection in the model overestimates the ventilation of thermocline waters here. We conclude that there is no indication that diabatic forcing of the conveyor belt in the subtropical North Atlantic is strongly overrated in the model.

A second region where conveyor belt water is transformed prior to the occurrence of deep convection is the midlatitude South Atlantic. Here air/sea interaction plays a dominant role in changing the water mass characteristics of the conveyor belt. The air/sea interaction is enhanced by convection between the two upper layers in the model, which reflects the deepening of the winter mixed layer. Convection entrains a larger amount of thermocline water into the surface layer, where it is subject to atmospheric forcing. If there is too much convection (too much deepening of the winter mixed layer) in the model, the diabatic forcing of conveyor water belt will be too large.

The maximum mixed layer depth in this region is 50 m [*Levitus*, 1982], the depth of the model upper layer. In the model, there is frequent convection between the two uppermost layers in the winter season, so the equivalent maximum mixed layer depth in the model extends to 100 m here. In this region the model overestimates the ventilation of thermocline waters by convection by about 35%. This model result is also in agreement with the findings of *Matano and Philander* [1993], who concluded that the transformation of intermediate water into surface water in the South Atlantic is overestimated in the coarse resolution model of *Bryan* [1969]. The subsequently too large transformation of conveyor belt water in the South Atlantic midlatitudes is in agreement with the hypothesis that in reality the warm water path plays a role in constituting the global conveyor belt.

7. Summary and Conclusions

The large-scale water movement associated with the upper branch of the conveyor belt in the Hamburg LSG model has been investigated with the help of a particle tracking method. Initial positions were chosen in the upper ocean, just south of the region of deep convection (NADW formation). The origin of the water parcels is calculated by reversing the advective and convective components in time. Mean fields were derived from the run ATOS1 [*Maier-Reimer et al.*, 1993].

A thousand trajectories have been integrated backward in time until the parcels upwell from the deep ocean. Calculations were stopped after 500 years when the parcels still resided in the upper ocean. Together, the trajectories form a representative picture of the upper branch of the model's conveyor belt. Consistent with the ratio of NADW outflow from the Atlantic to the NADW formation rate, 73.1% of the parcels upwelled outside the Atlantic and 26.9% upwelled within.

All trajectories enter the South Atlantic via Drake Passage, in agreement with the model's flow field showing a complete Agulhas retroflexion. Before entering the South Atlantic, parcels made an average of one extra loop around Antarctica. The trajectories depict a thermohaline circulation in which all oceans communicate. The ACC acts as a blender in mixing the various water mass characteristics associated with the global upwelling of the NADW return flow with those of Antarctic Intermediate Water. Apart from the ACC the wind-driven gyres play a prominent role in the global conveyor belt. A large part of the conveyor belt recirculates in the subtropical gyres. When convection is excluded in the trajectory calculation, the fraction of conveyor belt water which flows in the surface layer is decreased with 35%. Due to convection a larger amount of conveyor belt water is subject to atmospheric forcing and enhanced diffusion in the mixed layer. As a result, convection increases the amount of water mass transformation of the conveyor belt water. Prior to NADW formation, large changes in water masses occur in the South Atlantic midlatitudes between 40°S and 25°S and in the subtropical North Atlantic between 10°N and 25°N.

In the South Atlantic midlatitudes the strong water mass transformation is associated with thermocline water entrained by the deepening of the winter mixed layer. The entrainment is too large in the model because of too much convection between the two uppermost layers. As a result the water mass transformation of conveyor belt water is too large. This supports the hypothesis that in reality part of the conveyor belt follows the warm water path.

A large increase in temperature and salinity also occurs in the subtropical North Atlantic between 10°N and 25°N. Diffusive mixing, and to a lesser extent air/sea interaction, increase the temperature and salinity. The strong mixing is associated with flow from the AAIW low-salinity tongue into the warmer and saltier subtropical thermocline. The mixing is achieved while the water recirculates in the subtropical gyre, where part of it is ventilated into the deepening winter mixed layer. Here the amount of deepening of the winter mixed layer in the model is consistent with observations.

The timescale on which diabatic effects can be neglected is only a few years for the upper branch of the conveyor belt, as compared to a timescale for the circulation of $O(100)$ years. It should be noted, however, that the latter result could be model dependent, in that it is determined by the parameterization of convection, the model time step (1 month), and the diffusive characteristics of the model (upstream advection). The extent to which the modeled conveyor belt is sensitive to model parameters cannot be determined until a comparison with other model runs has been made.

The water mass characteristics of the conveyor belt are mainly determined by local diabatic processes along its path in the Atlantic. This indicates that air/sea interaction is important for the conveyor belt in a much larger region than the area of NADW formation, which underlines the importance of the conveyor belt for global climate. North of 30°N, the conveyor

belt cools and freshens, caused both by air/sea interaction and diffusive mixing. The water mass characteristics, however, remain saline and warm relative to the surrounding waters. When atmospheric cooling increases further north, the conveyor belt is preconditioned for deep convection.

In conclusion we may state that our analysis of the conveyor belt in a coarse resolution model has shed light on the transformation of conveyor belt water along its path in the Atlantic. The model results suggest that a conveyor belt following the cold water path is not inconsistent with both the observed atmospheric forcing and a realistic simulation of the principal water masses. However, in the model, there is too much transformation of Antarctic Intermediate Water into surface water within the conveyor belt, which might be related to the absence of Agulhas leakage. Although eddy-resolving ocean models have not been spun up to a steady state, the performance of a similar analysis in a global eddy-resolving model and comparison with the results obtained here would shed more light on the role of Agulhas leakage in the global conveyor belt.

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