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Heat and Mass Balances of the South Atlantic Ocean Calculated From a Numerical Model

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The general circulation model of Bryan (1969), modified by the introduction of open boundary conditions at the Drake Passage and between Africa and Antarctica, has been used to study the mass and heat budgets of the South Atlantic Ocean. The model was initialized with the climatological annual mean values of temperature and salinity of Levitus (1982) and forced at its surface with the climatological wind stress data of Hellerman and Rosenstein (1983). After 3 years of integration the model reached a quasi-stationary state. A heat balance shows that the model transports 0.19 PW of heat toward the north across 30°S. While a large part of this heat is supplied by the atmosphere and involves the conversion of intermediate waters into surface waters, a comparison with climatological data of atmospheric heat fluxes suggests that an extra source of heat is necessary to maintain the northward heat flux.

1. INTRODUCTION

Because of its complex geometry the oceanic circulation of the South Atlantic Ocean has a number of distinctive features (Figure 1). In the northern half of the basin the surface circulation is characterized by an anticlockwise, wind-driven gyre closed to the west by the Brazil Current, the western boundary current of the South Atlantic Ocean. The Brazil Current flows south along the continental slopes of Brazil, Uruguay, and Argentina to a point near 40°S where it encounters a branch of the Antarctic Circumpolar Current (ACC), the cold, northward flowing Malvinas Current. After this confluence both currents turn eastward and flow offshore in a series of large-scale meanders [*Gordon and Greengrove*, 1986; *Reid*, 1989]. The surface circulation of the eastern part of the basin is dominated by the eastward inflow of the Agulhas Current, the western boundary current of the Indian Ocean. The Agulhas Current flows south along the eastern coast of Africa to 35°S, where it branches near the southern extreme of that continent. One branch passes over the Agulhas Bank and turns to the northwest, while the major part, probably under the influence of the Agulhas Plateau, retroflects and joins the ACC [*Lutjeharms and van Ballegooyen*, 1988]. In the southern half of the South Atlantic basin the upper circulation is characterized by the eastward flow of the ACC. In general, the ACC is not very fast but is quite deep. Its average volume transport in the Drake Passage has been estimated to be around 120 Sv (10^6 m^3 s^{-1}) [*Whitworth and Peterson*, 1985], with approximate changes of 20% during the year [*Nowlin and Klinck*, 1986].

The meridional circulation in the South Atlantic Ocean is characterized by a flux of heat that goes from the pole toward the equator. This is the result of a circulation cell in which warm waters flow equatorward and cold waters flow poleward. The cold waters are produced in the North Atlantic Ocean. There, warm thermocline waters, flowing northward beneath a cool atmosphere, experience a strong heat loss, mainly through evaporation. This results in the production of cool, saline North Atlantic Deep Water (NADW) in the subarctic regions. After sinking in the North Atlantic the NADW flows south along the western margin of the American continent. At the middle latitudes of the South Atlantic these waters turn eastward and flow into the Indian and Pacific oceans, where they upwell into the thermocline [*Warren*, 1981]. The origin of the warm waters that maintain the equatorward heat flux through the South Atlantic Ocean is controversial. *Gordon* [1986] suggested that the heat deficit of the Atlantic Ocean can be maintained by the entrainment of warm Indian waters at the Agulhas retroflection. *Rintoul* [1991] pointed out that the equatorward heat flux can also be sustained by heat transfer from the atmosphere. Cold intermediate waters entering through the Drake Passage may gain heat from the air above and then flow northward as thermocline waters. Because of the scarcity of data it is difficult to confirm either of these hypotheses. The dominance of one of these two sources of upper flow, one warm and salty from the Indian Ocean and the other cold and fresh from the Pacific, will have a large impact on the thermohaline properties of the Atlantic.

The purpose of this study is to recalculate the heat and mass balances of the South Atlantic using the climatological data set of *Levitus* [1982] and a numerical model. Analyses of oceanographic data are usually made by the dynamical method, which combines hydrographic data and thermal wind equations to estimate the vertical gradient of velocities. The major drawback of this method is the need to choose an arbitrary reference level to determine absolute velocities. Better results may be obtained with the beta spiral method of *Stommel and Schott* [1977] or the inverse technique of *Wunsch* [1978], both of which take into account the conservation of various physical properties. Although these techniques give more consistent results than the dynamical calculations, they do not fully exploit the equations that govern oceanic motion. These drawbacks can be overcome through the use of a numerical model, for which it is necessary to specify the temperature, salinity, wind stress forcing, and other surface fluxes as initial and boundary conditions. By solving the full set of primitive equations a numerical model, in principle, permits a realistic estimate of

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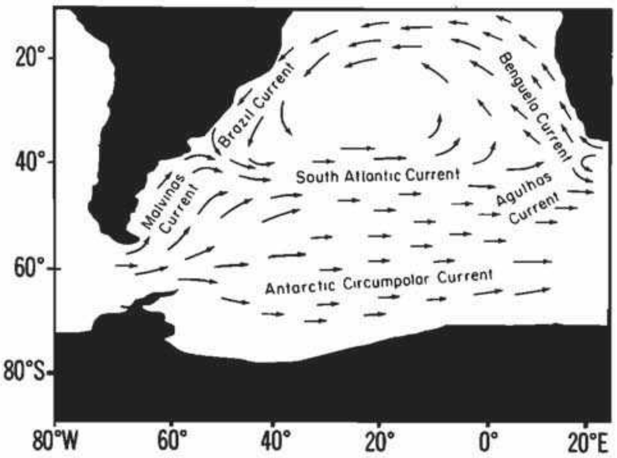

Fig. 1. Schematic representation of the surface circulation of the South Atlantic Ocean.

the transports and has the advantage of permitting a dynamical adjustment of the density field to the wind forcing and bottom relief. This adjustment is not allowed by the other methods mentioned above.

When using a numerical model the integration time must be specified. Integrations of approximately 1000 years are necessary to reach an equilibrium state when starting from an arbitrary initial state. If a realistic density field is specified initially, then the integration time required to reach a dynamical equilibrium can be far shorter (often only the few years required to generate the currents associated with the density gradients). In some calculations, those of *Rintoul* [1991] for example, the density is held fixed. We permit changes in the density resulting from a dynamical adjustment, but by limiting the simulation to a short span of time we avoid changes in the density caused by diffusive processes associated with a thermodynamic adjustment. (We believe that the parameterization of mixing processes is a weak point of the model used.)

This paper has four sections. Following this introduction, section 2 describes the model and its boundary conditions, section 3 presents the results, and section 4 gives a summary and some concluding remarks.

2. Background

For the sake of completeness, a brief description of the model and its boundary conditions will be given here. The reader interested in a more complete description is referred to *Matano* [1991].

2.1. *The Model*

The model used in this study is the multilevel numerical model described by *Bryan* [1969] and *Cox* [1984]. Figure 2 shows the domain of the model. It extends from 20°S to 70°S and from 72°W to 15°E. Solid walls are placed at 20°S and 70°S. In the interior of the basin the bottom topography and coastal distribution are realistic except that the Antarctic Peninsula and the islands of the South Atlantic Ocean are under 60 m of water. The horizontal resolution varies from 0.5° near South America to 1° elsewhere. The model has 15 vertical levels, and it is initialized from the climatological annual mean fields of temperature and salinity taken from *Levitus* [1982].

2.2. *The Boundary Conditions*

At solid boundaries both velocity components and the normal gradients of temperature and salinity are zero. Near the zonal walls at the northern and southern extremes of the domain the model equations for temperature T and salinity S have the additional terms $\gamma(T - T^*)$ and $\gamma(S - S^*)$, where T^* and S^* are the prescribed annual mean climatological temperature and salinity, respectively, for the region under consideration and γ is the Newtonian cooling coefficient. Its value is 0.5 d^{-1} near the wall and decreases smoothly to zero

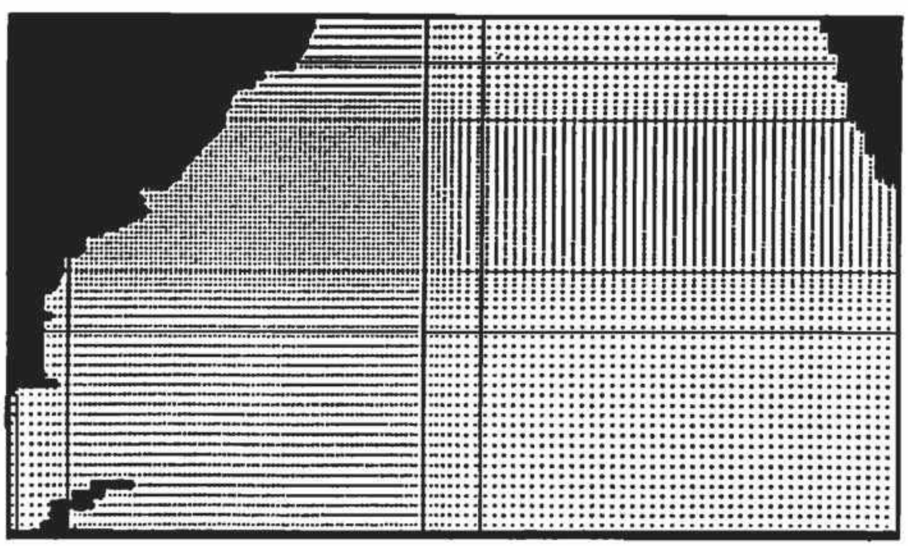

Fig. 2. Domain and grid point distribution of the numerical model. It extends from 20°S to 70°S and from 72°W to 15°E. In the interior of the basin the bottom topography and coastal distribution are realistic except that the Antarctic Peninsula and islands of the South Atlantic Ocean are under 60 m of water. The horizontal resolution varies from 0.5° near South America to 1° elsewhere. The model has 15 vertical levels.

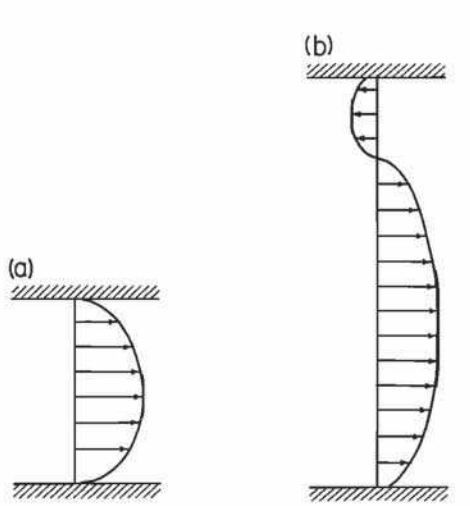

Fig. 3. The inflow/outflow conditions imposed on the barotropic component of the transport in the numerical model (*a*) at the Drake Passage and (*b*) at the eastern boundary. The total transport in both cases is 120 Sv.

southward. This device mitigates the effect of the artificial solid wall and forces the solution toward the climatology in those regions. The temperature and salinity in the first layer of the model are also restored to climatic annual mean temperatures and salinity with a Newtonian coefficient γ of $1/30\ d^{-1}$. This restoration to the climatological data provides the surface heat and water fluxes that maintain the thermohaline circulation. The model is forced at the surface with the climatological annual mean wind stress taken from *Hellerman and Rosenstein* [1983].

At the lateral open boundaries it is necessary to specify boundary conditions for the mass stream function, baroclinic velocities, temperature, and salinity. The mass stream function determines the total transport of the ACC and its horizontal distribution. Since our goal is to reproduce the mean circulation, it is appropriate to specify conditions that represent the average state of the ocean. Figure 3 shows the latitudinal distributions of the barotropic velocities that result from the mass stream function distributions chosen for the eastern and western boundaries of the model. At the Drake Passage the total mass transport is 120 Sv to the east [*Whitworth and Peterson*, 1985]. Since the grid size of the model does not permit the resolution of the jetlike structure of the ACC, we specified a current with a maximum transport in the middle of the channel (where the deepest regions are located) that weakens toward the sides. Observations indicate that in the region between Africa and Antarctica the ACC still conserves the jetlike structure that is characteristic in the Drake Passage [*Sarukhanyan*, 1985]. This structure is not present in the climatological data of *Levitus* [1982], nor can it be resolved by the present model. Instead, the barotropic velocities specified at the eastern boundary were guided by the numerical experiments of *Cox* [1975] and *Semtner and Chervin* [1988] and the observations of *Peterson and Stramma* [1991]. Since mass must be conserved, the total mass outflow between Africa and Antarctica must equal the inflow at the Drake Passage. The latitudinal distribution of this mass transport (Figure 3*b*) includes a westward flow of 8 Sv in the Agulhas region, representing the entrainment of Indian waters into the South Atlantic Ocean, and a broad eastward flow further south.

Once the barotropic component of the transport has been imposed, boundary conditions for the baroclinic variables must be chosen. At an open boundary there can be inflow and outflow. When there is inflow, temperature and salinity are prescribed, and the baroclinic velocities are adjusted geostrophically. At boundaries with outflow it is necessary to impose a condition that allows phenomena generated in the interior domain to pass through without distortion and without affecting the interior solution. The one used in this study is the Sommerfeld radiation condition [*Sommerfeld*, 1949].

One issue of importance is the sensitivity of the present results to changes in the open boundary conditions. At the Drake Passage the mass and heat balances to be discussed are more sensitive to changes in the total transport of the ACC than to changes in its meridional distribution. The small width of the passage and the strong steering effect of the bottom topography drive the ACC flow to the northernmost part of the channel, regardless of the initial distribution of the current. At the eastern boundary, significant differences appear only when considerable changes are introduced in the meridional structure of the imposed stream function. Such changes, however, lead to flow parameterizations that are in conflict with both observations and the climatological data with which the model is initialized. In the region of the Agulhas Current it is necessary to increase the entrainment of Indian Ocean waters into the South Atlantic from 8 Sv to 40 or 50 Sv to develop significant differences with the reported results.

2.3. *The Spin-Up*

A measure of the oceanic adjustment after the calculations are started is provided by the integral of the absolute value of the temperature tendency,

$$E(t) = \sum_x \sum_y \sum_z \frac{|\Delta T(x, y, z, t)|}{\Delta t} \tag{1}$$

where ΔT is the change in temperature and Δt is the time interval. Figure 4, which shows the time behavior of (1), reveals a rapid initial adjustment in the first year followed by a slow trend that is attributable to diffusive processes. Since the diffusive processes in the model are likely to distort the density field, we terminated the calculations before diffusion became important but after the dynamical adjustment had taken place. In what follows, the term "equilibrium" will be used to refer to the state of the model after 3 years of numerical integration.

3. THE RESULTS

3.1. *Temperature and Velocity Fields*

The set of hydrographic variables obtained by integrating the model has some advantages over the original *Levitus* [1982] data set. This is noticeable, for example, in the formation of strong fronts in the upper ocean and in the sharpening of western boundary currents. As an illustration, Figure 5 shows the instantaneous temperature field of the upper layer of the ocean from the Levitus data (Figure 5*a*)

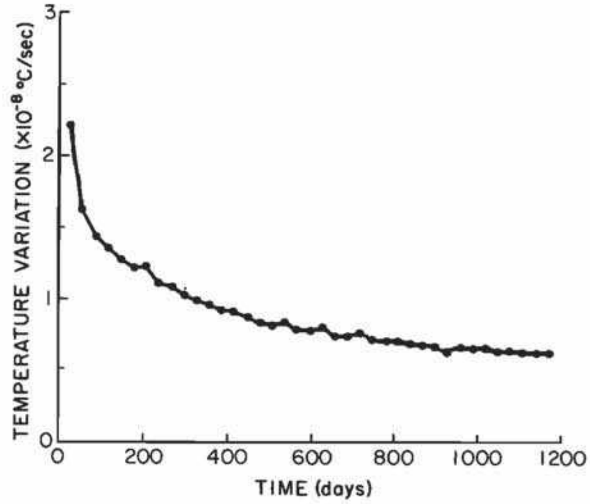

Fig. 4. The global integral of the absolute value of the temperature variation. This integral represents the temporal response of the model.

and from the model after equilibrium (Figure 5*b*). While there is good agreement between model and data over a large portion of the basin, there is a considerable sharpening of frontal structures in the temperature field of the model relative to the Levitus data. This is particularly noticeable in the southwestern Atlantic where the strong advective effects associated with the Brazil/Malvinas Confluence produce a marked displacement of the isotherms. The model also plays

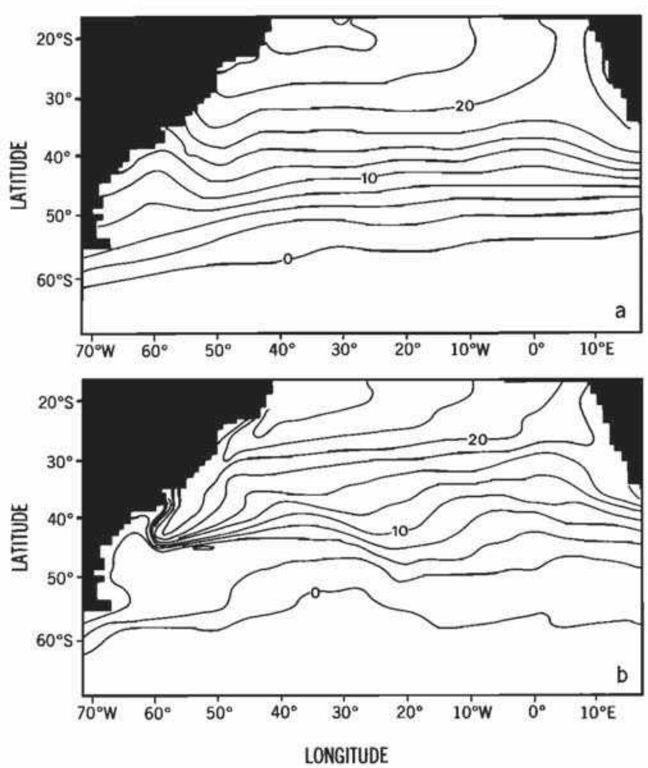

Fig. 5. Instantaneous temperature field at the sea surface (*a*) from the *Levitus* [1982] climatological data and (*b*) from the model after 3 years of integration.

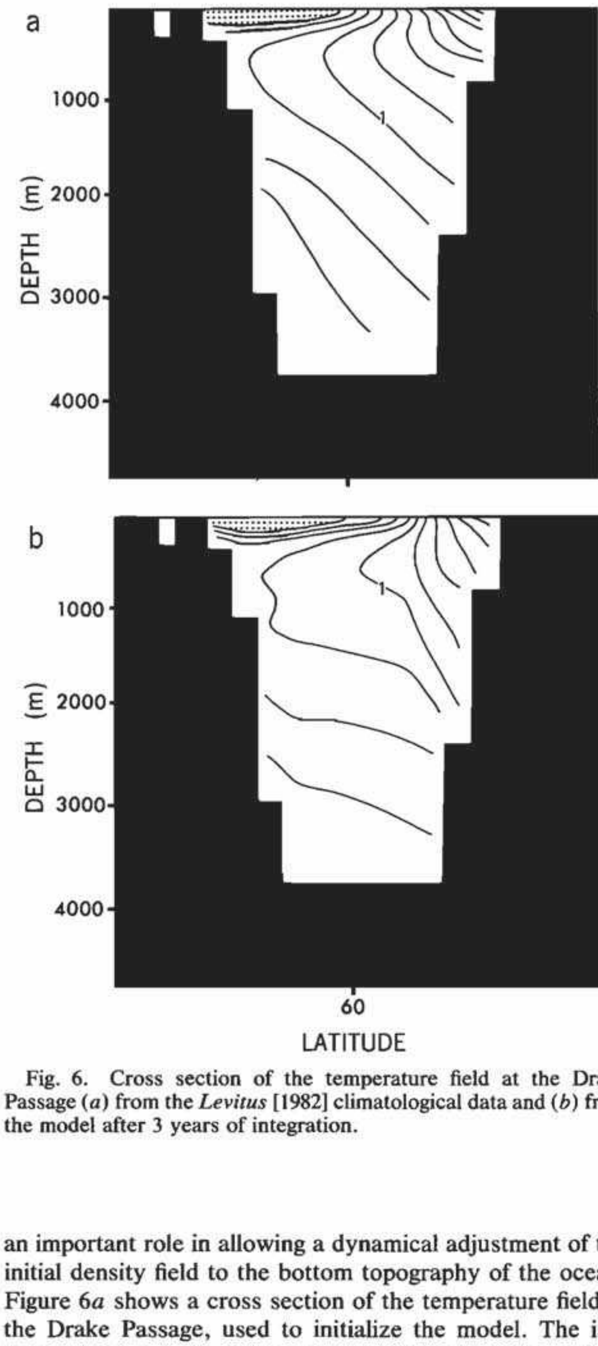

Fig. 6. Cross section of the temperature field at the Drake Passage (*a*) from the *Levitus* [1982] climatological data and (*b*) from the model after 3 years of integration.

an important role in allowing a dynamical adjustment of the initial density field to the bottom topography of the ocean. Figure 6*a* shows a cross section of the temperature field at the Drake Passage, used to initialize the model. The isotherms have a more or less constant slope from the surface to the bottom, implying that the transport is distributed uniformly with depth. Figure 6*b* shows the isotherms at the equilibrium state of the model. Because of a pronounced decrease in the ocean depth to the east of this section, a large portion of the ACC is deflected northward and confined to the top 2500 m of the passage. This information was presumably eliminated by the interpolation and smoothing process applied to the Levitus data but was recovered by the dynamics of the model.

In the interior of the ocean the thermal structure of the deep flow remains close to its initial value. Figure 7 shows the latitudinal distribution of temperature with depth for the midbasin region at 30°W. The thermocline, which is evident north of 50°S, has a bowllike shape in mid-latitudes and is almost flat at low latitudes. In general, it is confined to the

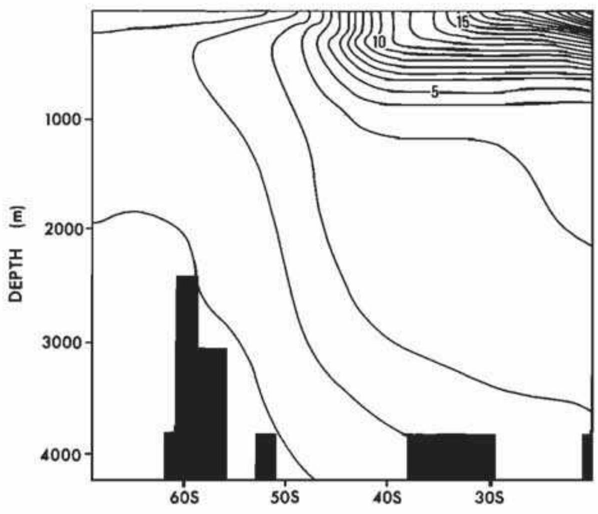

Fig. 7. Latitudinal distribution of temperature with depth for the midbasin region at 30°N.

upper 1000 m, while the deep part of the basin is filled with water of a potential temperature of less than 3°C.

Figure 8 shows the equilibrium stream function for the vertically integrated transport. After leaving the Drake Passage the ACC flows eastward and bifurcates at about 55°S, where it forms the Malvinas Current. The Malvinas Current flows northward until it converges with the Brazil Current where both turn offshore. The northern portion of the basin is dominated by an anticlockwise gyre with a clear western intensification. The top-to-bottom transport of this boundary current, which includes both the wind-driven Brazil Current in the upper layers and the thermohaline-forced NADW in deep layers, is about 60 Sv. This value agrees with those calculated previously by other authors. *Mellor et al.* [1982], using a diagnostic model with wind stress and real bottom topography, found a value of 60 Sv. *Rintoul* [1991], applying inverse methods to hydrographic data, estimated a value of 63 Sv at 32°S, which is greater than the 50 Sv calculated by the present model at that latitude. Considering the differences in the data sets and in the methods of calculation, the disagreement is small. On the other hand, *Cox* [1975], using the same model and initial conditions as described here, but in a global domain, estimated the transport of the western boundary current as 30 Sv. It is likely that the difference in the estimated transport between the two calculations is related to differences in the horizontal resolution. It is possible that the coarse resolution of *Cox*'s [1975] model (2°) tends to smooth the bottom topography, thereby eliminating the contribution to the total transport of the baroclinic-topographic torque. Note that 30 Sv is the transport predicted by the Sverdrup theory from the wind stress curl alone [*Hellerman and Rosenstein*, 1983]. More recently, *Semtner and Chervin* [1988] simulated the thermocline circulation of the world oceans using an eddy-resolving version of Bryan's model [*Bryan*, 1969]. In their results the poleward transport near the coasts of Brazil was about 40 Sv. Although this value is greater than the one calculated by *Cox* [1975], it is still smaller than the one obtained in this study. The differences with Semtner and Chervin's model are probably related to the longer time integration of that model.

Figure 9 shows the depth distributions of the zonal velocities at 30°W. The highest values are found at the subtropical confluence where it is possible to distinguish two eastward jets (the Malvinas and Brazil currents) with maximum surface velocities of approximately 10 cm s^{-1}. The confluence is not restricted to the upper layers but reaches depths of 2000 m. North of the confluence, the westward flow is dominated by the South Equatorial Current, which has maximum velocities of only 2 cm s^{-1}. In the southern ocean it is possible to distinguish the barotropic structure of the ACC, which favors a strong topographic control of its trajectory [*Matano*, 1991].

3.2. *Mass and Heat Balances*

The results of the present experiment allow us to make heat and mass balance calculations for the whole South Atlantic basin. The water masses were classified according to their potential temperature σ_θ and depth z as shown in

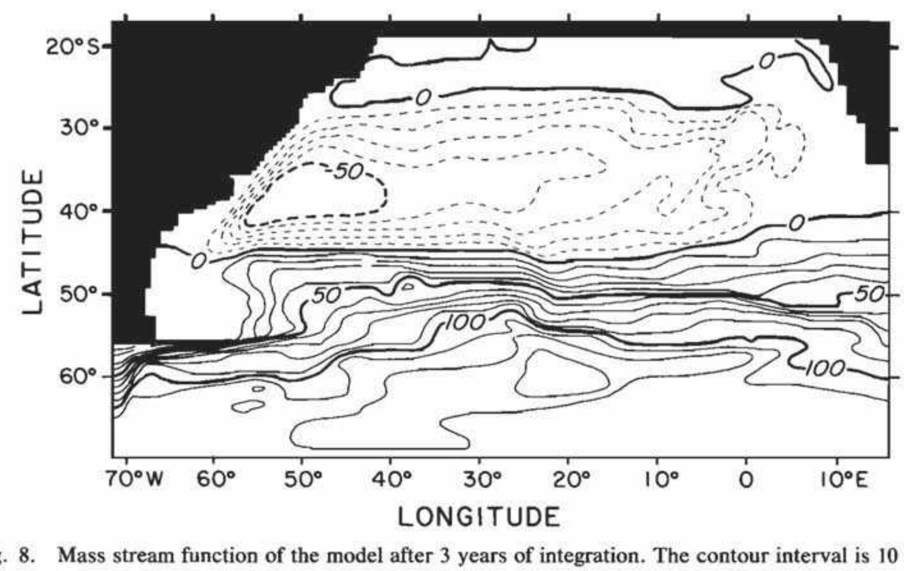

Fig. 8. Mass stream function of the model after 3 years of integration. The contour interval is 10 Sv.

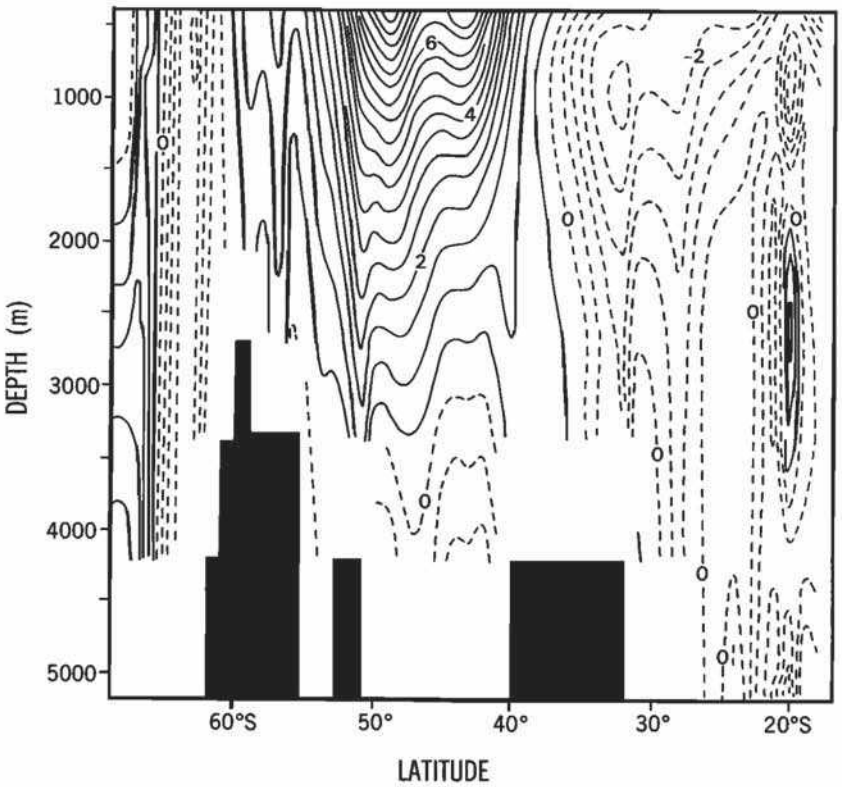

Fig. 9. Depth distributions of the zonal component of the velocity field at 30°W.

Table 1, using definitions adapted from *Rintoul* [1991]. Figure 10 summarizes the results for the flows through the Drake Passage and across 12°E and 30°S. In this figure the net transport of each water mass is plotted as a bar whose length is proportional to the transport. At 30°S the zonal integral of the circulation consists of an equatorward flow of surface, intermediate, and bottom waters and a poleward flow of deep water. The northward flux of warm water and the southward flow of cold water result in a net northward flux of heat of 0.19 petawatt (PW). At the Drake Passage the zonal transports are to the east in all layers. The transport of surface waters is small in comparison with that of intermediate and deep waters. The relatively small depth of the passage restricts the flow of the highest-density waters from the Pacific Ocean. The pattern near the eastern boundary is similar to that in the Drake Passage except that the transport of intermediate waters has decreased by about 8 Sv and there is a net outflow of bottom water and an increase in the eastward transport of deep water. The total mass balance is presented in Figure 11. The South Atlantic as a whole gains intermediate and deep waters from the other oceans. A portion of these waters is then converted into surface and bottom waters and exported back. It is thought that the production of bottom waters occurs primarily near the continental shelf of the Weddell Sea where Antarctic waters are made sufficiently cold through heat loss to the atmosphere and sufficiently saline through salt gain from the ice formation to become dense enough to reach the floor of the ocean [*Gill*, 1973].

As *Rintoul* [1991] previously noted, apparently, there is a net production of surface waters in the South Atlantic Ocean (Figure 11). The hypothesis is that the strong westerlies blowing over the southern ocean produce an equatorward transport of surface waters that are then replaced by an upwelling of deeper waters. These waters, in turn, are colder than the air above and extract heat from the atmosphere, changing their physical characteristics to those of surface waters. These positive contributions from the atmosphere can be appreciated qualitatively in the maps of heat flux of *Bunker* [1988], in which the region to the south of the subtropical convergence is shown as a zone of net heat gain from the atmosphere.

Since mass and heat fluxes are strongly related, further insight into the mass balance can be gained by considering the heat balance. To that end, net temperature fluxes were calculated near the eastern and western boundaries of the model and at 30°S. The net heat divergence in the South Atlantic is approximately 0.3 PW, of which about 0.1 PW is lost by the ACC and 0.19 PW is transported northward across 30°S. As this figure is larger than the value that can be obtained from *Bunker*'s [1988] data, a preliminary conclusion may be that the present calculation overestimates the production of surface waters by the atmosphere and that alternate sources of heat (and surface waters) are necessary. It seems likely that an increase in the entrainment of waters

TABLE 1. Classification of Water Masses

Water Mass	Classification
Surface water	$\sigma_\theta < 26.2$
Intermediate water	$\sigma_\theta > 26.2$ and $z < 1839.44$ m
Deep water	$1839.44 \text{ m} < z < 3783.76$ m
Bottom water	$3783.76 \text{ m} < z$

Here σ_θ represents potential temperature, and z represents depth.

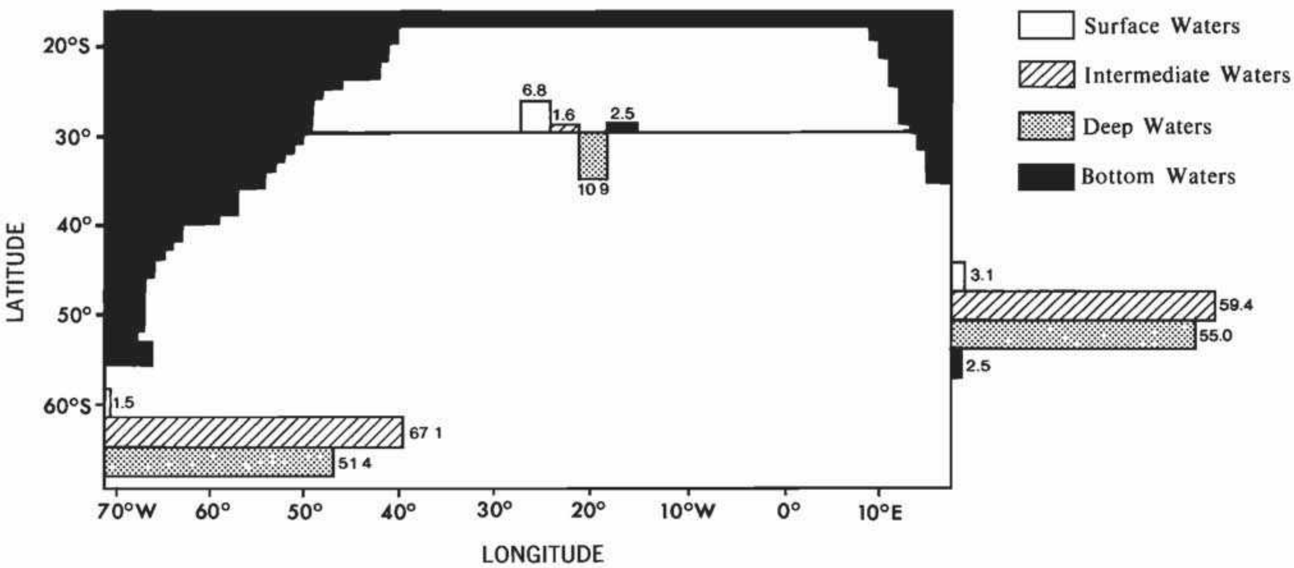

Fig. 10. Mass balance of the South Atlantic model. The mass definitions are given in Table 1. The length of each bar is proportional to the transport (in sverdrups).

from the Indian Ocean at the Agulhas retroflection may close the balance.

Rintoul [1991], using synoptic data collected during the International Geophysical Year, carried out an analysis similar to the present one but employed an inverse method to calculate the velocities from the density field. In this model the northward heat transport at 32°S was 0.24 PW. According to Rintoul, this heat deficit can be maintained by a positive heat flux from the atmosphere at high latitudes of the southern ocean. Although the results of the present study are similar to those of Rintoul, the uncertainties associated with the estimation of the heat transfer from the atmosphere make it difficult to conclude that the waters coming from the Drake Passage are the main supply that balances the Atlantic deficit of thermocline waters. Maps like those of *Bunker* [1988] allow only a qualitative estimate of the sign of this transfer because of the lack of data. Nevertheless, the apparent conversion of intermediate into surface waters in the present calculations indicates that at least a large fraction of the thermocline waters in the Atlantic Ocean may gain heat from the atmosphere.

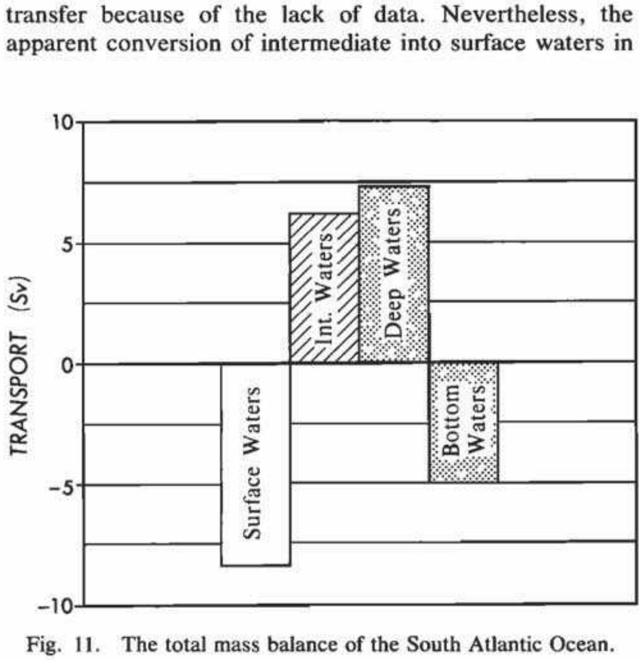

Fig. 11. The total mass balance of the South Atlantic Ocean.

4. Summary and Discussion

An ocean general circulation model was used to calculate the mass and heat balances of the South Atlantic Ocean. The model was initialized and forced with climatological mean values of temperature, salinity, and wind stress. After 3 years of numerical integration the model reached a dynamical equilibrium. A heat balance shows that the model exports 0.19 PW of heat to the north across 30°S. A mass balance shows that this heat flux is the result of a meridional cell in which warm waters flow northward and cold waters flow southward. Although a portion of the surface waters may have been formed by a conversion of intermediate waters into surface waters in the southern ocean [*McCartney*, 1977; *Rintoul*, 1991], a comparison between the model results and the estimated heat fluxes from the atmosphere suggests that an extra source of surface waters is necessary to maintain the northward heat flux. As originally suggested by *Gordon* [1986], this extra source may be an increased entrainment of Indian Ocean waters in the Agulhas retroflection region.

If the Atlantic draws surface waters from both the Indian and the Pacific oceans, it is possible that the temporal dominance of one source over the other is controlled by the latitude of the Subtropical Convergence. A southward position of the Subtropical Convergence allows a major penetration of the warm waters of the Subtropical Gyre into high latitudes, with a consequent reduction of the global amount of heat gained by the ocean. Conversely, if the Subtropical Convergence is farther north, the cold waters of the southern ocean can reach lower latitudes and increase the amount of heat gained from the atmosphere. *Matano* [1992] and Matano et al. (Seasonal variability in the southwestern Atlantic, submitted to *Journal of Geophysical Research*, 1992) showed that there is a direct connection between the meridional displacements of the Subtropical Convergence and the transport of the ACC, which therefore could control the thermohaline properties of the Atlantic Ocean. When the ACC transport is high and the Subtropical Convergence is at its northernmost position, the main supplier for thermocline waters could be the intermediate waters introduced at the Drake Passage and transformed into surface waters by gaining heat from the atmosphere [*Rintoul*, 1991]. At those times the Atlantic may become fresh and cool. Alternatively, when the ACC transport is low, the Agulhas eddies could have a dominant role, and the Atlantic may become warmer and saltier [*Gordon*, 1986]. A simple calculation provides an estimate of the importance of this mechanism. Suppose that the Subtropical Convergence moves meridionally 5° and that as a result of this shift there is a change of 2°C in the average temperature. Then, a water column of 100-m depth would provide a change in the heat flux of about 25 W m^{-2}. For the South Atlantic this means a change in the heat flux from the atmosphere of 0.1 PW, which is of the same order as its northward heat flux.

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REFERENCES

Bryan, K., A numerical method for the study of the circulation of the world ocean, *J. Comput. Phys.*, *4*, 347–376, 1969.

Bunker, A. F., Surface energy fluxes of the South Atlantic Ocean, *Mon. Weather Rev.*, *116*, 809–823, 1988.

Cox, M. D., A baroclinic numerical model of the world ocean: Preliminary results, in *Numerical Models of the Ocean Circulation*, pp. 107–120, National Academy of Sciences, Washington, D. C., 1975.

Cox, M. D., A primitive equation, 3-dimensional model of the ocean, *Tech. Rep. 1*, Geophys. Fluid Dyn. Lab. Ocean Group, Princeton University, Princeton, N. J., 1984.

Gill, A. E., Circulation and bottom water production in the Weddell Sea, *Deep Sea Res.*, *20*, 111–140, 1973.

Gordon, A. L., Interocean exchange of thermocline waters, *J. Geophys. Res.*, *91*, 5037–5046, 1986.

Gordon, A. L., and C. Greengrove, Geostrophic circulation of the Brazil-Falkland confluence, *Deep Sea Res.*, *Part A*, *33*, 573–585, 1986.

Hellerman, S., and M. Rosenstein, Normal monthly wind stress over the world ocean with error estimates, *J. Phys. Oceanogr.*, *13*, 1093–1104, 1983.

Levitus, S., Climatological atlas of the world ocean, Natl. Oceanic and Atmos. Admin., professional paper, U.S. Govt. Print. Off., Washington, D. C., 1982.

Lutjeharms, J. R. E., and R. C. van Ballegooyen, Anomalous upstream retroflection in the Agulhas Current, *Science*, *240*, 1770–1772, 1988.

Matano, R. P., A numerical study of the circulation in the South Atlantic Ocean, Ph.D. thesis, 165 pp., Atmos. and Oceanic Sci. Program, Princeton Univ., Princeton, N. J., 1991.

Matano, R. P., On the separation of the Brazil Current from the coast, *J. Phys. Oceanogr.*, in press, 1992.

McCartney, M. S., Subantarctic Mode water, in A Voyage of Discovery, edited by M. Angel, *Deep Sea Res.*, suppl., 103–109, 1977.

Mellor, G. L., C. R. Mechoso, and E. Keto, A diagnostic calculation of the general circulation of the Atlantic Ocean, *Deep Sea Res.*, *Part A*, *29*, 1171–1192, 1982.

Nowlin, W. D., and J. M. Klinck, The physics of the Antarctic Circumpolar Current, *Rev. Geophys.*, *24*, 469–491, 1986.

Peterson, R. G., and L. Stramma, Upper-level circulation in the South Atlantic Ocean, *Prog. Oceanogr.*, *26*, 1–73, 1991.

Reid, J. L., On the total geostrophic circulation of the South Atlantic Ocean: Flow patterns, tracers and transport, *Prog. Oceanogr.*, *23*, 149–244, 1989.

Rintoul, S. R., South Atlantic interbasin exchange, *J. Geophys. Res.*, *96*, 2675–2692, 1991.

Sarukhanyan, E. I., *Structure and Variability of the Antarctic Circumpolar Current*, Amerind, New Delhi, 1985.

Semtner, A. J., and R. M. Chervin, A simulation of the global ocean circulation with resolved eddies, *J. Geophys. Res.*, *93*, 15,502–15,522, 1988.

Sommerfeld, A., *Partial Differential Equations*, 333 pp., Academic, San Diego, Calif., 1949.

Stommel, H., and F. Schott, The beta spiral and the determination of the absolute velocity field from hydrographic measurements, *Deep Sea Res.*, *24*, 325–329, 1977.

Warren, B. A., The deep circulation of the world ocean, in *Evolution of Physical Oceanography*, pp. 6–41, MIT Press, Cambridge, Mass., 1981.

Whitworth, T., and R. G. Peterson, The volume transport of the Antarctic Circumpolar Current from bottom pressure measurements, *J. Phys. Oceanogr.*, *15*, 810–816, 1985.

Wunsch, C., The North Atlantic general circulation west of 50°W determined by inverse methods, *Rev. Geophys.*, *16*, 583–620, 1978.

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