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Assimilation of Temperature Profiles in a General Circulation Model of the Tropical Atlantic

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ABSTRACT

Thirty-six hundred temperature profiles collected during 1984 were assimilated into a multilayer primitive equation model of the tropical Atlantic Ocean. The method consists in a monthly correction of the simulated temperature field. Each month, an observed field is computed from the temperature profiles with a successive correction analysis starting from the simulated field. The difference between the observed field and simulation is computed, the model is restarted from the previous month, progressively introducing the difference as a Newton forcing in the heat equation. The sensitivity to the initial state is greatly reduced near the equator after six months, but persists for a longer time at higher latitudes. The assimilated temperature structure is closer to the observations than was the unassimilated simulation. The thermocline has strengthened, and low-frequency variability near the equator is close to the observed one, resulting in a more realistic zonal slope of the thermocline. The current structure, although it still differs noticeably from the observations, is more realistic, with stronger near-surface countercurrents and a faster equatorial undercurrent.

1. Introduction

A major objective of the TOGA (Tropical Ocean Global Atmosphere) programme is to test whether a coupled ocean–atmosphere model can predict the evolution of the climate system at low latitudes. An important intermediate goal should be tested before the coupled models become operational, namely whether tropical ocean models can realistically hindcast the variability in the three oceans. Leetma (1987) has performed a near real time simulation of the upper Pacific with the GFDL model (Philander et al. 1987). More recently, our team has attempted a similar experiment in the tropical Atlantic (Merle and Morlière 1988) using an ocean general circulation model (OGCM) developed at the Laboratoire d'Océanologie DYnamique et de Climatologie (LODYC) (Chartier 1986; Delecluse, Andrich and Foujols, personal communication. The first runs without assimilating data were presented in Morlière et al. (1988). A comparison with temperature observations suggested that the simulation did have some flaws. For instance, the temperature profile was too diffuse, particularly in the eastern Atlantic, and the thermocline was poorly represented.

One way to improve the simulations is to force the model with data by assimilating observations. While meteorologists have developed data assimilation schemes used operationally to forecast the atmosphere, assimilation experiments in OGCMs are still uncommon. Moore et al. (1987) have tested, in an "identical twin" approach, the sampling strategy of a TOGA-like network in the Indian ocean, by assimilating simulated "XBT observations" into their OGCM. Their results were inconclusive, showing that the assimilation of temperature profiles results in little improvement with an OGCM, but that definite improvements are achieved with a linear one-layer reduced gravity model. Hayes et al. (1989) have compared in the tropical Pacific ocean, the simulations with assimilation to an independent dataset. They suggest that the assimilation results in a significant improvement for sea surface temperature, but that thermocline depth and velocity field were simulated with mixed results.

In this note, we describe how we assimilate temperature profiles from XBTs in the tropical Atlantic OGCM, and discuss a one-year simulation (1984), with and without assimilation, and its sensitivity to initial conditions. The main results of the assimilation are that (i) the temperature profiles has considerably sharpened, (ii) the equatorial thermocline zonal slope is better simulated, (iii) the equatorial circulation is stronger.

2. Conditions of experiment

a. The model

A detailed description of the OGCM is presented in Chartier (1986) and Delecluse et al. (1988). Briefly, the model is a multilayer primitive equation OGCM

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with the Boussinesq, hydrostatic and rigid-lid approximations. The primitive equations are discretized by finite differencing methods on a C-grid. The parameterization of vertical diffusivity is the one used by Pacanowski and Philander (1981). The equation of state has been extremely simplified to a density chosen as a linear function of temperature. Density inversions are eliminated at each time step by vertical adjustment to stability of the unstable piece of profile. This version of the code extends from Africa to America and from 20°S to 20°N, which are closed boundaries. The grid step is variable with latitudinal resolution increasing from ⅓ of a degree at the equator to 1.5 degree at the meridional boundaries, and a longitudinal resolution that varies from 0.5 degree near the coasts to 1 degree at the center of the basin. There are 16 vertical levels from 0 to 3750 meters, and neither bottom topography nor islands are included. At 20°S and 20°N, a no slip boundary condition and a damping zone are applied. Finally, the time step has been chosen as 40 minutes.

The external forcing consist of heat fluxes at the air–sea interface, and a wind stress. The shortwave and longwave components of the radiation budgets are chosen to be constants (203 W m^{-2} and 56 W m^{-2}, respectively). Sensible and latent heat fluxes are computed from the usual bulk formula ($C_D = 1.4 \times 10^{-3}$). The air temperature is from the monthly climatology given by Esbensen and Kushnir (1981), and the sea surface temperature is the one from the model simulation.

The wind field is based on the analyses of ship observations by Servain et al. (1987). They provide, on a 2 × 2 latitude–longitude grid, a monthly mean of pseudostress. To construct a wind field, a drag coefficient is computed following Large and Pond (1981). A stability dependence is introduced which results in an increase of the stress in light wind conditions. A description of the wind stress is presented in Morlière et al. (1988), where it is shown that the ocean is better simulated with this product than by using a wind field for 1984 extracted from the operational atmospheric model at ECMWF (European Centre for Medium range Weather Forecasts).

The initial conditions for the model are (i) no currents, (ii) the spatially uniform thermal stratification used by Philander and Pacanowski (1980). An equilibrium seasonal cycle is reached after 4 years of model integration forced by climatological winds of Hellerman and Rosenstein (1983). The model was then run for two more years (1982–83) using a wind field derived from the initialized fields of the ECMWF operational atmospheric model (V. Cardone and Y. Tourre, personal communication). The simulations presented hereafter, start on 1 January 1984 from the model state on 31 December 1983.

b. Temperature observations

Temperature profiles from 1982 to 1984 have been collected into a single file and validated for the FOCAL/SEQUAL experiment (Reverdin et al. 1988). We focus the study on 1984 and 3610 profiles are extracted for this year including XBT data from the merchant ship network (now TOGA network), and XBT and CTD data from research vessels involved in the FOCAL/SEQUAL and others experiments (Fig. 1). Most profiles reach 400 m, for each profile we kept the temperature down to 325 m at the 13 upper levels of the model.

3. Method for data assimilation

The principle of our method is to compute for each month the difference between the simulated and observed temperature fields and to rerun the simulation for the month under study with a corrective term added in the model temperature equation. The corrective term is an additional "experimental" heat flux which

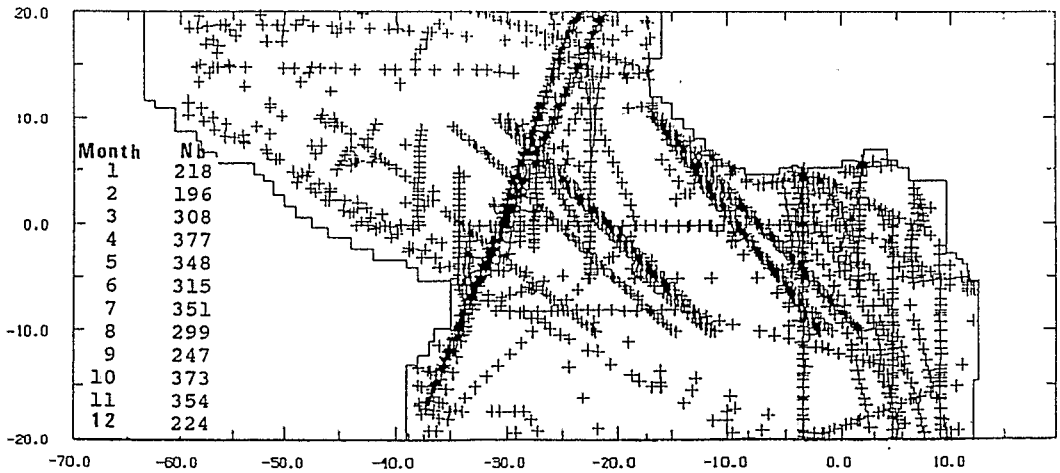

FIG. 1. Number of temperature profiles in 1984.

progressively introduces the difference between observed and simulated temperature data. The final state of the ocean from this "corrected" simulation is used as an initial state for the next month of simulation. More information on the different steps is now presented.

The first step is to estimate the model monthly field. Every 7.5 days a simulated temperature field of the ocean is recorded, then a midmonth simulated mean temperature field, T_m, is computed by a Gaussian-shaped weighted mean of five 7.5-day temperature fields. The temporal weighting function is the same as in the objective analysis described in the section hereafter.

From the observed data we produce an analyzed temperature field, T_0. This accomplished using a method of successive corrections described in Cadet and Reverdin (1981) and based on Berthorsson and Döös (1955), Cressman (1959) and Tripoli and Krishnamurti (1975). The initial guess field is the mean simulated temperature field T_m produced by the OCGM. Data are weighted in time with the Gaussian weighting function centered at midmonth. The final radii of influence (the finest resolved scales) are $R_x = 800$ km on the longitude axis and $R_y = 300$ km on the latitude axis.

Then a new simulation is carried from the previous month, adding, at each grid point, a corrective term in the heat equation during n time steps between day 7.5 and day 22.5. The corrective term is written:

$$K^*(T_m - T_0)/(nt)$$

where t is the time step. Because of some large differences between simulated observed fields (Morlière et al. 1988) the first assimilation experiment produced an unrealistic patchy field. Thus, we introduced a factor K reducing the magnitude of the corrective term. The value used is 0.4; it can be tuned later on when the model is closer to reality. The final state of the ocean from this "corrected" simulation is used as an initial state for the next month.

4. Results

To investigate the effect of the assimilation scheme, and to assess how dependent it is on its initial conditions, two experiments with assimilation were conducted and are compared to a control run, the run without assimilation. The first one starts from the simulated ocean on 31 December 1983 described earlier. The second one is initialized with the model fields after one year of month by month temperature assimilation. The thermal structure of the model is then much closer to reality than the non-assimilated fields. The state of the model on 31 December 1983, used as initial condition of the first experiment was too far from reality due to the fact the model progressively degraded the thermal structure especially at the level of the thermocline which becomes too diffuse (panel 1 of Fig. 2). Thus the first experiment can be considered as a spinup phase for the second experiment.

We present the effect of assimilation of temperature profiles by discussing successively, the vertical stratification, the zonal equatorial thermocline slope and the zonal current field. Of course, because we assimilate temperature data, an improvement of the simulated thermal field is expected, but the currents which are not assimilated, constitute an independent check.

a. The thermal field

Figure 2 presents, every two months, an average temperature profile from all available data in a given month, and compare it with the control run and the assimilation experiment 1, at the locations of the observations. It should be noticed that the mean simulated profiles with assimilation (dashed line) represents the mean of all the profiles corresponding to a simulation with assimilation during previous months and no assimilation during the month under study. So, the data used in the comparison have not been included in the model average profile with assimilation. Figure 2 indicates a progressive improvement in the simulated field with assimilation, particularly at the thermocline depth (largest stratification).

At the equator, the temperature correction results in a rise of the thermocline represented by the 20°C isotherm depth (Fig. 3). This rise is more important in the eastern part of the Atlantic and in summer. This upward displacement of the simulated thermocline is associated with a general cooling between 50 and 200 m linked with the improvement of the vertical temperature profile shown in Fig. 2. At 4°W, where the observations are plentiful, and result in a monthly time series of the 20°C isotherm depth, which are comparable to within a 3 m rms with moored thermistance chain data, this rise brings the simulation close to the observational time series (Fig. 4).

b. The equatorial thermal zonal slope

The equatorial zonal slope of the thermocline or of the surface dynamic height is often chosen as the most important parameter to describe the variability in the equatorial Atlantic (Merle 1980; Katz 1981). Here for 1984, we compute the slope of the 20°C isotherm depth from temperature profile for our two experiments (Fig. 5). This is comparable to the slope of the surface dynamic height presented by Hisard and Hénin (1987) with the smaller dataset of the FOCAL cruises. to the investigation of inverted echo sounders travel time in Katz (1987) or to the investigation of the equatorial mooring temperatures by Weisberg and Weingartner (1987).

The simulated 20°C isotherm depth is always deeper than the observed one (Fig. 5) with the largest differ-

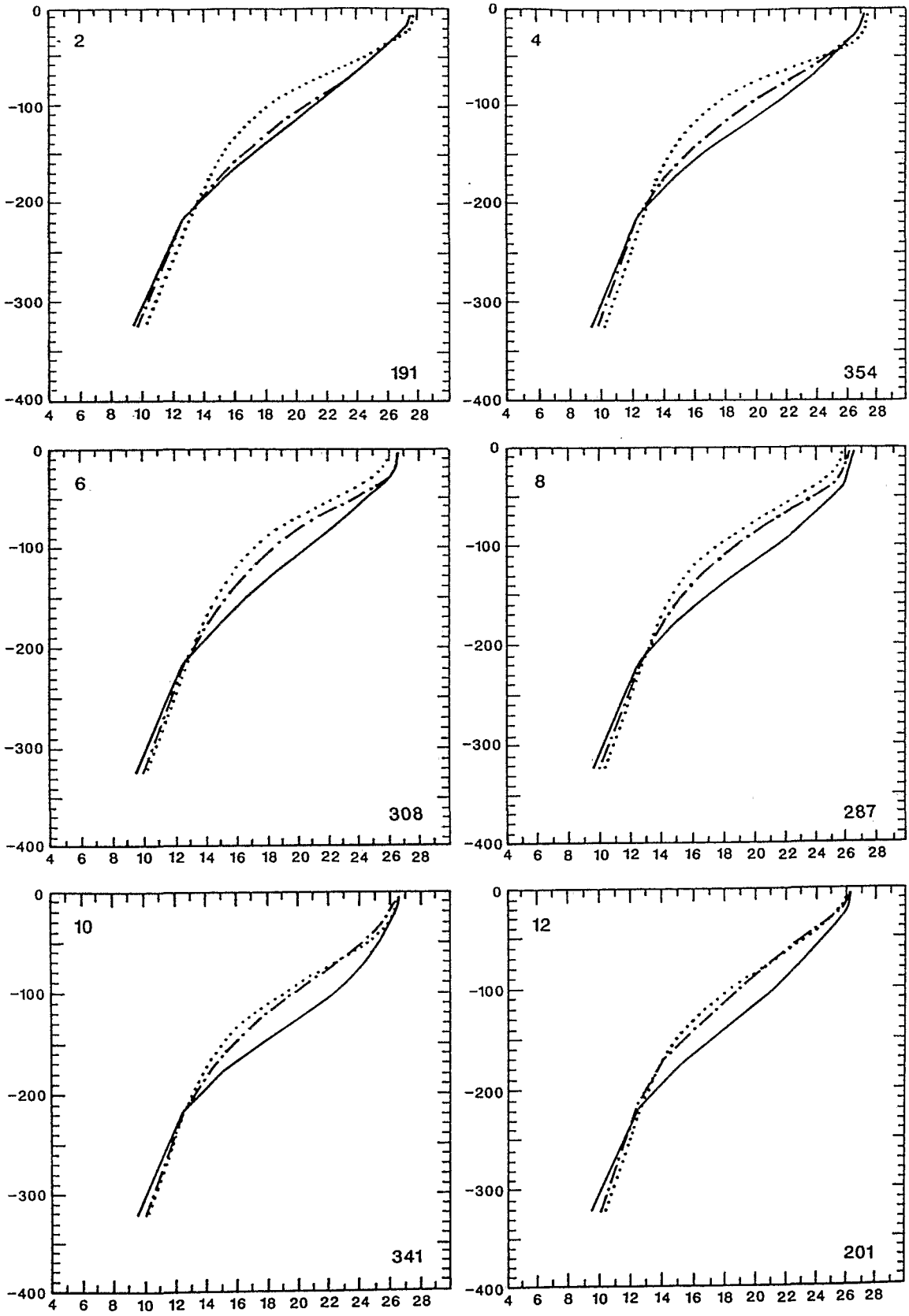

FIG. 2. Monthly mean temperature profiles from the observations (dotted line), the control run with no assimilation (solid line), the experiment-1 run with assimilation (mixed line). This last profile correspond to a simulation with assimilation of data in the previous months in 1984, but no assimilation for the month shown. The month is indicated in the left upper corner, and the number of observed profiles is indicated in the right lower corner.

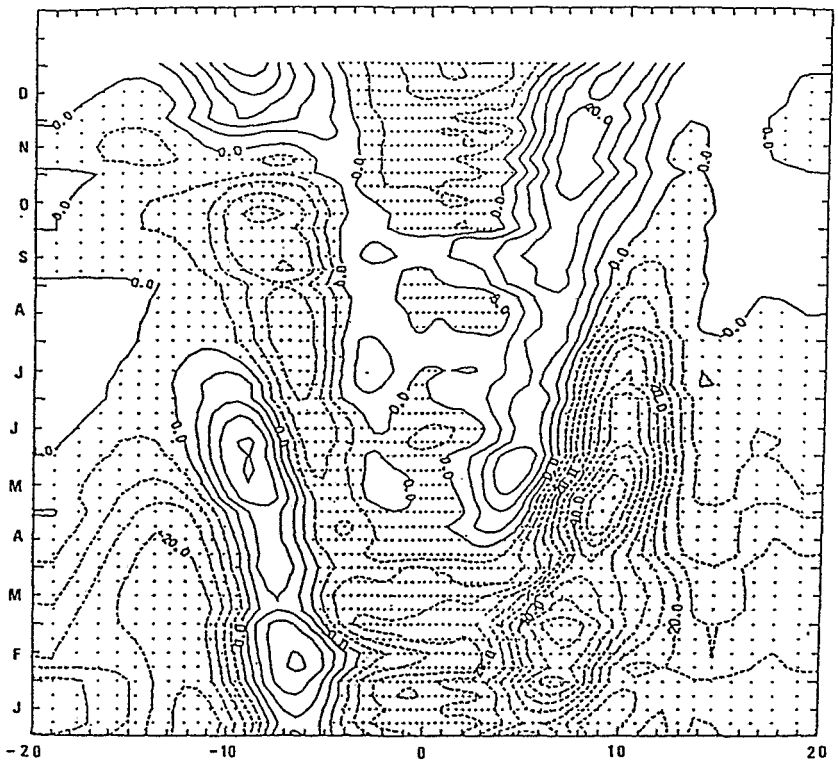

FIG. 3. Time–longitude diagram in 1984 of the difference in the 20°C isotherm depth between assimilated experiment 1 and control run with no assimilation (contouring every 5 m).

ences in the eastern Atlantic. The depth of the thermocline along the equator from noncorrected simulations is greatly dependent on the initial state and the differences from the observed one are large, although some of the time variability is present in the control run (Morlière et al. 1988).

After six months of assimilation, the equatorial slopes for the two simulations with different initial states are very close. However, 5° to 10° away from the equator, differences between the two experiments remain large for a long time (an example of these differences is presented on Fig. 6). These differences typically have a meridional scale less than 5°, for which the analysis T_0 was not strongly constrained by the observations, and therefore for which the "assimilated" run is only weakly influenced by observations.

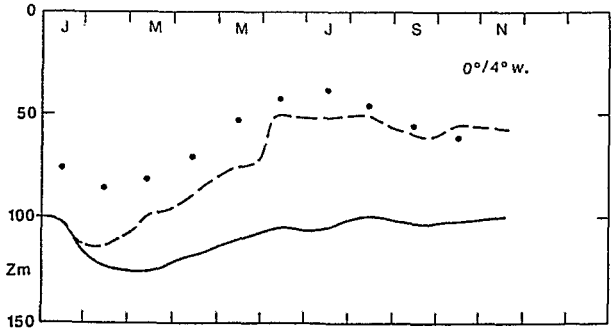

FIG. 4. The monthly time series of the 20°C isotherm depth at 4°W on the equator for the observations (dots), the control run without assimilation (solid line), and the experiment 1 assimilation (dashed line).

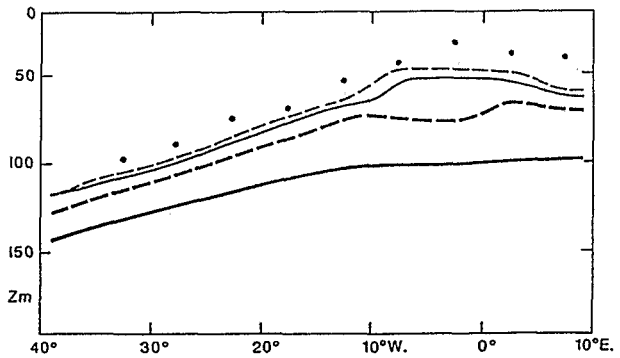

FIG. 5. The 20°C isotherm depth along the equator in July 1984 from the observations (dots), the control run without assimilation (heavy solid line), the experiment 1 with assimilation (solid thin line), and the experiment 2 with assimilation (dashed thin line). Another control run is also presented (dashed thick line) with the same initial condition as experiment 2.

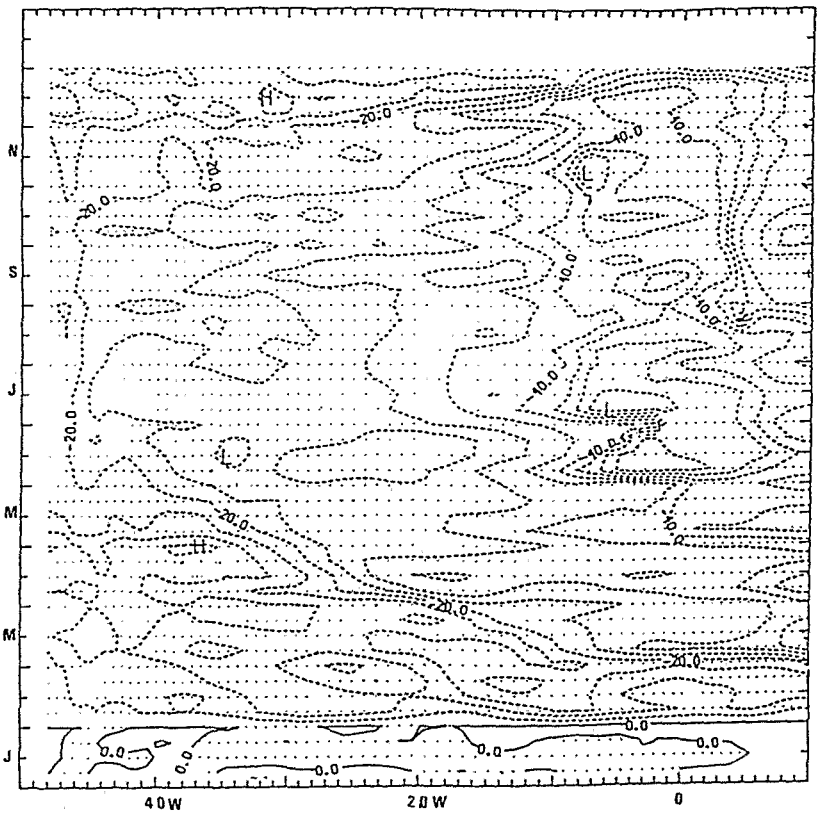

FIG. 6. Time–latitude diagram at 28°W of the difference of the 20°C isotherm depth between experiments 1 and 2 (contouring every 5 meters).

c. *The zonal currents*

The assimilation of temperature observations results in intensification of the equatorial currents, with magnitude closer to the observations from moorings, profiler sections or drifting buoys. Between 10°N and 10°S, the currents are in the average strengthened by the assimilation of the temperature profiles. At 20°W (Fig. 7), the core of maximum eastward speed in the Equatorial Undercurrent (EUC) is between 40 and 50 cm s^{-1} in June 1984 for the simulation without assimilation, and over 60 cm s^{-1} for the simulation with assimilation. This is in better agreement with the magnitude of the currents observed with current profiler during FOCAL cruises where the core velocity of the EUC reaches 80 cm s^{-1} in July 1984 (Hisard and Henin 1987). The eastward North and South Equatorial Countercurrents are 20 cm s^{-1} faster in the simulation with assimilation, while the westward South Equatorial Current has similar surface velocities, but a larger transport. A most unfortunate "bug" was hidden in the assimilation scheme: the temperature correction at 325 m was applied to the whole deep ocean, resulting in currents at these depths that are not realistic. It is very encouraging to find from the monthly fields (Fig. 7) that this did not have much effect in the upper thermocline.

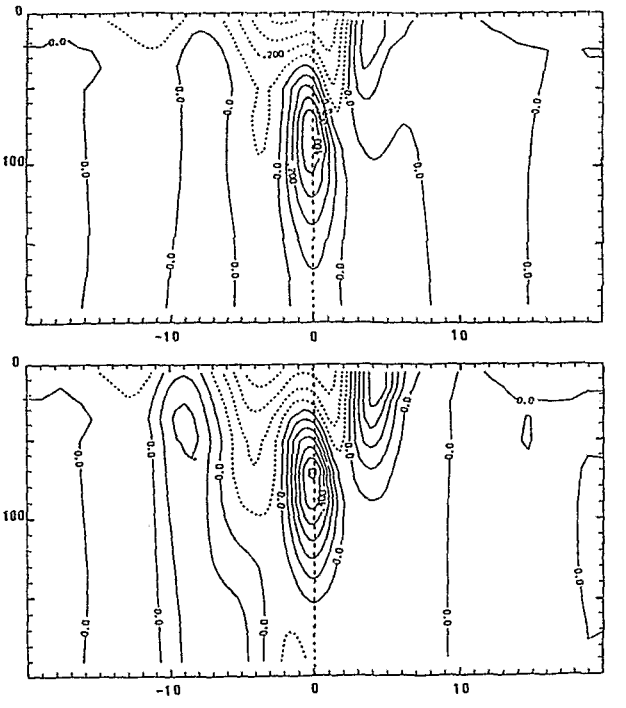

FIG. 7. Simulated meridional sections of the zonal current along 20°W in June 1984 for the control run without assimilation (upper panel) and for the experiment 1 with assimilation (lower panel).

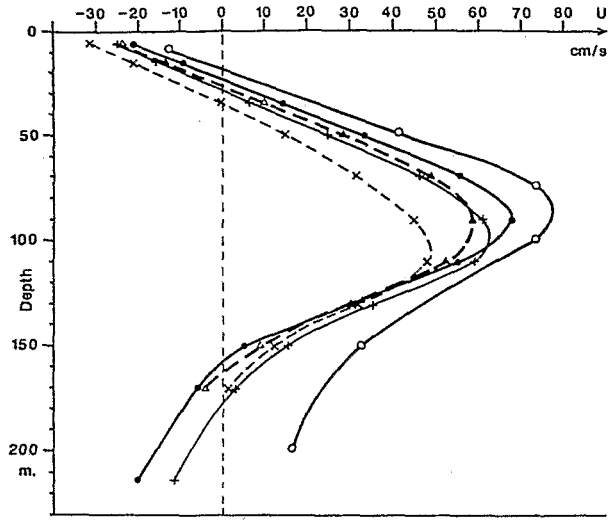

FIG. 8. The 1984 average vertical profile of the zonal current at 28°W, on the equator. The thick solid line with empty circle correspsond to currentmeter measurements on a surface mooring (Weisberg et al. 1988). The other lines have the meaning: The control run (thin dashed line with X), experiment 1 with assimilation (thick dashed line with triangles), and experiment 2 with assimilation (thick solid line with dots). Another control run without assimilation is also presented (thin solid line with +) with the same initial condition as experiment 2.

The annual mean profiles of the eastward component of the current observed on the equator at 28°W from a mooring (Weisberg et al. 1987) is compared with similar simulated data from our two experiments during 1984 (Fig. 8). The mean eastward component of the simulated current is ever slower than the observed one, but between the surface and 130 meters it has increased in magnitude with respect to the profile from the control run without assimilation. Beneath the upper thermocline, at 130 meters and lower, there is no improvement, and the simulated currents disagree with the measurements, whether or not data have been assimilated.

5. Conclusion

With a simple assimilation scheme, we have shown that it is possible to significantly improve the results of a three-dimensional OGCM. The future objective is to correct the simulated temperature field using available real-time temperature observations. The principle of our correction is empirical. After a monthly run, differences between observed and simulated fields are computed. The model is then run again for the same period adding a corrective term in the heat equation at all grid points in order to reduce progressively the gap between the two fields. This method improves the structure of the thermal field as well as the ocean equatorial current field. Of course, obvious improvements should be made to optimally include the information from temperature profiles. A better equation of state should be used, and salinity could be introduced.

The model efficiency is improved by using the method of assimilation–correction but still remains highly dependent on the simulated ocean's initial state. Therefore, it appears that, given the present stage of oceanic models, we cannot get realistic results from a model without numerous and continuous observations which can be used to correct the erroneous initial conditions, inadequate flux fields at the air–sea interface and the model imperfect physics.

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