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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

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

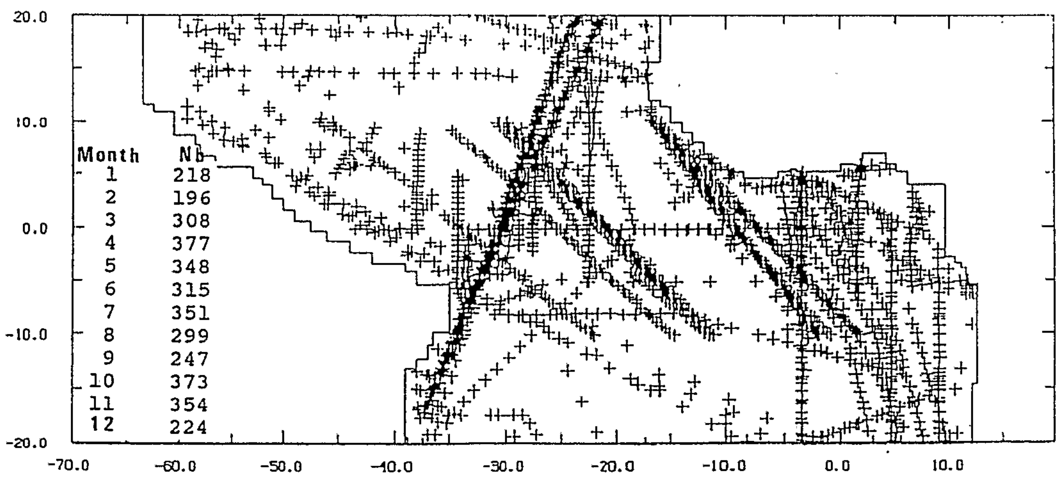


FIG. 1. Number of temperature profiles in 1984.

progressively introduces the difference between observed and simulated temperature data. The final state of the ocean from this "corrected" simulation is used

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

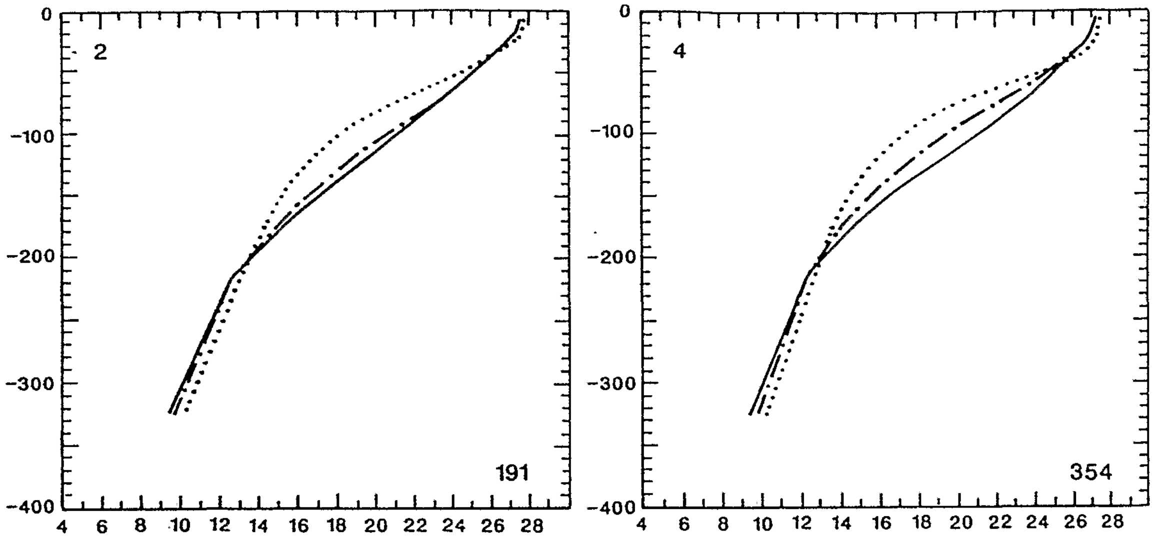
mocline which becomes too diffuse (panel 1 of Fig. 2). Thus the first experiment can be considered as a spinup phase for the second experiment.

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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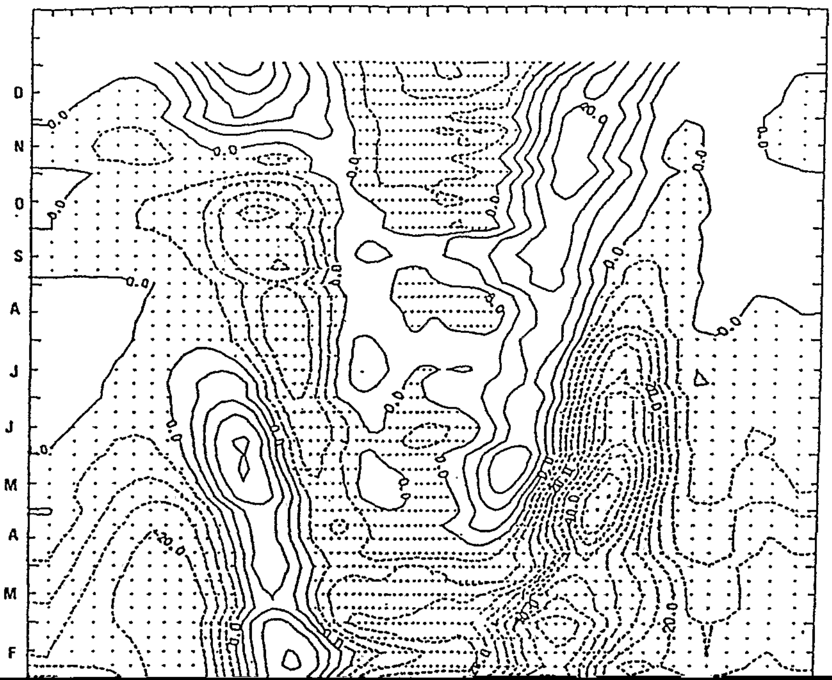
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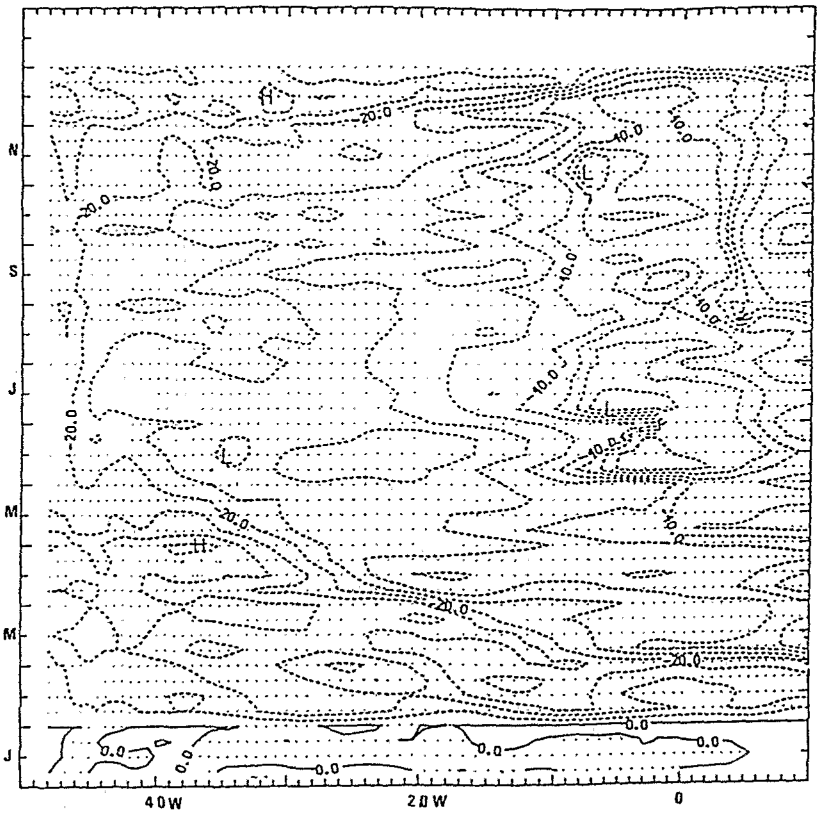


FIG. 6. Time–latitude diagram at 28°W of the difference of the 20°C isotherm depth between experiments 1 and 2 (contouring every 5 meters).

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.

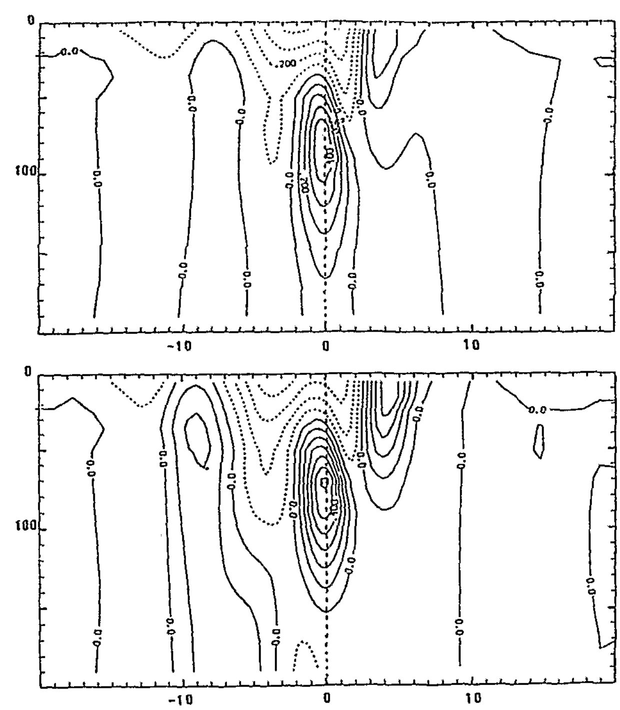
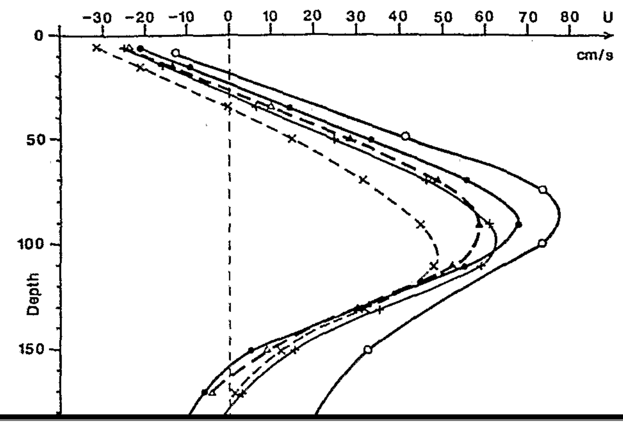


FIG. 7. Simulated meridional sections of the zonal current along 20°W in June 1984 for the control run without assimilation (upper panel) and for the experiment 1 with assimilation (lower panel).



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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