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On the connection between the 1984 Atlantic warm event and the 1982–1983 ENSO

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ABSTRACT

The warm event which spread in the tropical Atlantic during Spring–Summer 1984 is assumed to be partially initiated by atmospheric disturbances, themselves related to the major 1982–1983 El-Niño which occured 1 year earlier in the Pacific. This paper tests such an hypothesis. For that purpose, an atmospheric general circulation model (AGCM) is forced by different conditions of climatic and observed sea surface temperature and an Atlantic ocean general circulation model (OGCM) is subsequently forced by the outputs of the AGCM. It is firstly shown that both the AGCM and the OGCM correctly behave when globally observed SST are used: the strengthening of the trades over the tropical Atlantic during 1983 and their subsequent weakening at the beginning of 1984 are well captured by the AGCM, and so is the Spring 1984 deepening of the thermocline in the eastern equatorial Atlantic, simulated by the OGCM. As assumed, the SST anomalies located in the El-Niño Pacific area are partly responsible for wind signal anomaly in the tropical Atlantic. Though this remotely forced atmospheric signal has a small amplitude, it can generate, in the OGCM run, an anomalous sub-surface signal leading to a flattening of the thermocline in the equatorial Atlantic. This forced oceanic experiment cannot explain the amplitude and phase of the observed sub-surface oceanic anomaly: part of the Atlantic ocean response, due to local interaction between ocean and atmosphere, requires a coupled approach. Nevertheless this experiment showed that anomalous conditions in the Pacific during 82–83 created favorable conditions for anomaly development in the Atlantic.

1. Introduction

The El-Niño event which developed in the equatorial Pacific during 1982–1983 was certainly one of the largest anomalous climatic episodes for many decades: positive sea surface temperature (SST) anomalies exceeded 3°C during many months over large areas. This event and its impact on the world climate are well documented, thanks to diagnostic studies (e.g., CAC Climate Diagnostics Bulletin, 1982–1983 issues; Quiroz, 1983), and modelling studies (e.g., Fenessy et al., 1985).

A warm SST episode also occurred in 1984 in the entire equatorial Atlantic. Positive SST anomalies exceeded 2°C during the Summer 1984 along the African coast where the seasonal upwelling practically vanished. In fact, July 1984 was very different from July 1983, when the upwelling was encouraged by strong trades (Fig. 1). This 1984 Atlantic warm event is considered to be one of the largest anomalous episodes that occured in the last 30 years, presented in Fig. 2 for the eastern tropical Atlantic. The averaged SST in that region was warmer than usual during the whole year 1984. This warm event is also well documented (e.g., Philander, 1986; Servain and Seva, 1987). This SST anomaly reflects strong subsurface anomalies in the ocean that were detected in the

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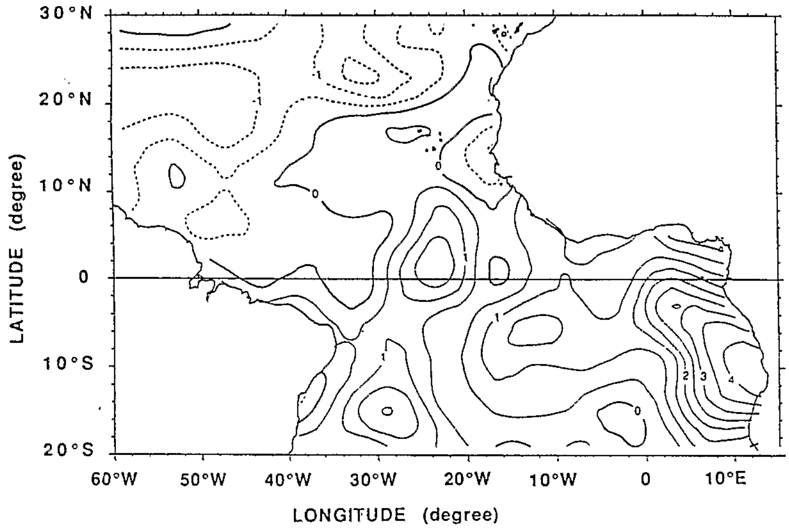


Fig. 1. Observed SST difference between July 1984 and July 1983 (in °C). Note the large positive anomalies in the Gulf of Guinea.

observations made during the 1982–1984 FOCAL/SEQUAL program (see the FOCAL/SEQUAL 1987 special issue of J. Geophys. Res. and Katz et al., 1986). It was showed that the thermocline and sea surface slopes were anomalously flat in early 1984. Oceanic simulations (e.g., du Penhoat and Gouriou, 1987; Reverdin et al., 1991) allowed us to understand relatively well the sequences of the Atlantic basin-wide 1984 episode.

The 1982–1983 Pacific warm event and the subsequent 1984 Atlantic warm event seem to be connected: discussions about possible links can be found in a few diagnostic studies (e.g., Horel et al., 1986) and atmospheric modelling studies (e.g., Mechoso et al., 1990). The physical support of such a link could be explained by an anomalous Walker-type circulation in the atmosphere between central/eastern equatorial Pacific and equatorial Atlantic:

• From mid-1982 to mid-1983 (this period corresponds to the mature phase of the 1982–1983 El Niño), the trades were very weak in the central part of the equatorial Pacific. Above the equatorial

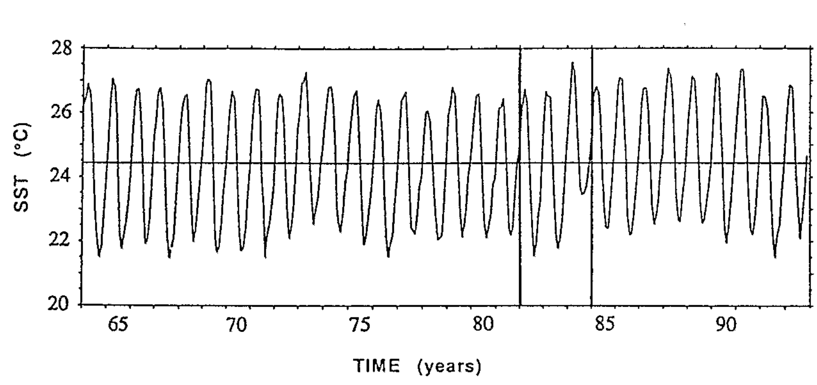


Fig. 2. Time series of the observed monthly SST (in °C) averaged over the eastern tropical Atlantic (from 10°W and 20°S to the coast) from 1964 to 1993. The 1982–1984 period is indicated by vertical lines. Note the warm event in 1984, especially during the cold summer season.

Atlantic during the same time, the westerly wind at 200 hPa increased while the easterly wind at 850 hPa were stronger than normal. There was a decrease of convective activity with a drought over Northeast Brazil in February–May 1983.

• A reversed situation occurred with rapid relaxation of zonal wind at both 200 hPa and 850 hPa from mid-1983 until the beginning of the year 1984, and a peak of convective activity over the equatorial Atlantic from February to May 1984. This decrease in the Atlantic atmospheric circulation induced an important weakening of the wind-stress over the western equatorial basin. This dramatic climatic event led to the inverse tilting of the equatorial ocean slope observed during the FOCAL oceanographic experiments in early 1984 and to the warm SST episode which reached its highest value during June–July and August 1984 in the eastern tropical Atlantic basin.

Additional work is needed to understand whether the Atlantic signal develops as a local coupled response or whether it is remotely initiated by the Pacific ocean anomalies, with a time-delay due to the development of the coupled response. The present study is the first step to investigate this question through an original combination of outputs from both a global atmospheric model and an ocean model of the tropical Atlantic. The episode 1982–83–84 is selected for its large amplitude: if the connection does not work for this one event, it will not work for events with smaller amplitude. Different atmospheric forcings for the ocean are firstly calculated by the atmospheric general circulation model. The ocean response to these different forcings is then analysed. A series of numerical experiments is performed in order to answer the following questions.

(a) Does the atmospheric model, forced by globally observed SST, correctly simulate the wind-stress anomalies over the tropical Atlantic ocean in 1982–83–84?

(b) Does the oceanic model, forced by observed wind-stress, accurately simulate the Atlantic warm event in 1984?

(c) Does the oceanic model, forced by the atmospheric model outputs, succeed to simulate the 1984 Atlantic warm event?

(d) Can the 1982–83 El Niño SST remotely initiate the Atlantic warm event?

2. Description of the experiment

2.1. *Atmospheric runs*

The AGCM outputs have been provided by P. Lenzen and K. Arpe from the Max-Planck-Institut für Meteorologie in Hamburg. The experiments have been performed with the ECHAM3 model (version T42/19 levels), described by Roeckner et al. (1992).

Two experiments of the AGCM were performed with different boundary conditions for SST. In each case the SST fields used to force the AGCM derived from the SST data set prepared by the Climate Analysis Center (CAC, NOAA/National Weather Service, National Meteorological Center, Washington, DC 20233), for the 1979–1988 Atmospheric Model Intercomparison Project (known as the "AMIP-SST" data set).

(i) 1st experiment: Atmospheric Worldwide Run (AWR).

The AGCM was forced by the worldwide observed AMIP-SST data set, month by month, from the period January 1979 until 1992. The period January 1982 to December 1984 was then extracted. This experiment was repeated 5 times for the whole series with different initial conditions and some 6 month periods (Spring 83 and Spring 84 for instance) were repeated.

(ii) 2nd experiment: Atmospheric Pacific Run (APR).

The AGCM was forced by monthly observed AMIP-SST data set for the tropical Pacific and monthly climatological SST elsewhere. The run had the same initial atmospheric data like the one above for January 1979 but anomalies were restricted to the tropical Pacific from January 1982 onward.

The climatology for the atmospheric runs is defined as the ten year mean of AWR.

2.2. *Oceanic runs*

The OGCM has been developed in the Laboratoire d'Oceanographie Dynamique et de Climatologie (LODYC) and is described in Reverdin et al. (1991) and in Blanke and Delecluse (1993). It extends over the tropical Atlantic ocean but the analysis is limited to the equatorial region where the model has a high resolution in latitude (0.3° at the equator) and is entirely prognostic. It is forced by the outputs from the AGCM in two different experiments. For the present study we chose the most recent version which includes a parameterization of vertical mixing based on the local formulation of turbulent kinetic energy (Blanke and Delecluse, 1993). The OGCM is forced by the momentum flux, the heat flux and the fresh water flux from the AGCM. To force the OGCM, atmospheric outputs were selected inside the geographical limits of the OGCM (56°N–34°S, 100°W–16°E) and averaged for 10 day period. The ocean model is initialized from climatological fields for temperature and salinity with the forcing conditions from the atmospheric climatology (from AWR) for 3 years. After 2 years, tropical regions are in quasi-equilibrium. A 3rd year is added and used as a control run for the ocean.

Two experiments were performed with the OGCM:

(i) The 1st run (OWR) starts from the last day of the control run. It is integrated with the 10 day mean forcing from AWR for 3 years.

(ii) The 2nd run (OPR) starts from the last day of the control run. It is integrated with the 10 day mean forcing from APR for 3 years.

3. Results

3.1. The atmospheric field over the tropical Atlantic

An ensemble of simulations has been done for AWR with different initial conditions. Over the equatorial Atlantic ocean, the wind-stress patterns are very similar from one run to the other. Fig. 3 presents 6 different realizations of wind-stress for Spring 83 (a) and Spring (84) (b). The contrast between both years is much higher than the differences between the simulations. The trades present a maximum in the western equatorial Atlantic in Spring 1983 and a clear weakening in Spring 1984. As a single simulation is representative of the ensemble mean, the following discussion will refer to one atmospheric run.

In order to estimate the quality of the wind-stress simulated by the atmospheric model, a comparison was done with wind-stress derived from wind observations collected by ships of opportunity. Such an analysis was performed using the wind-stress patterns prepared by Servain et al. (1987). The general pattern of the wind-stress simulated with observed SST (AWR) is in good agreement with observations. A test zone was chosen in the western equatorial Atlantic, between 2°N–2°S, from the Brazilian coast to 20°W, in an area where strong interannual variability in the wind-stress has been detected. The theory of atmospheric remote forcing developed by McCreary et al. (1984) relates the variability observed in the eastern Atlantic ocean to the variability of the zonal forcing in the western Atlantic ocean through the propagation of equatorial Kelvin waves. It is thus important to verify the wind-stress estimated from the AGCM in the western Atlantic. Fig. 4a presents the evolution (monthly filtered) for the observed and simulated (AWR) zonal wind-stress from 1982 to 1984 in this area. The simulated wind-stress is relatively close to the observed patterns. The seasonal cycle (low zonal wind-stress until the end of spring, increase from the end of spring until September–October, and then rapid decrease from December) is in good agreement, in amplitude and phase, with the observations. This indicates a clear improvement in the AGCM physics as a previous comparison between observations and analyses from ECMWF indicated weak trades along the equator (Morlière et al., 1989). In order to detect the interannual variability, the anomalies from the seasonal cycle are computed for both the model run and the observations (the seasonal cycle is defined as a mean over 27 years for the observations). Note that the interannual signal is clearly weaker in amplitude than the seasonal signal in the tropical Atlantic (and it is thus difficult to detect it) but it presents a consistent low-frequency pattern. The anomalies from the seasonal cycle are presented in Fig. 4b. An interannual signal is reproduced by AWR, in agreement with observed signal: the zonal wind-stress is stronger than usual in mid-1983 and decreases from mid-83 to mid-84 in the model, with a slightly stronger amplitude than calculated from the observations (the mean difference is -8.0×10^{-3} N/m^2 and the RMS is 15.7×10^{-3} N/m^2).

The spatial anomaly pattern of the simulation can be contrasted between the month of May 1983 and May 1984 (Fig. 5). Between 10°N and 10°S, most of the wind-stress anomalies are concentrated in the western tropical Atlantic and indicate a strengthening along the Brazilian coast during

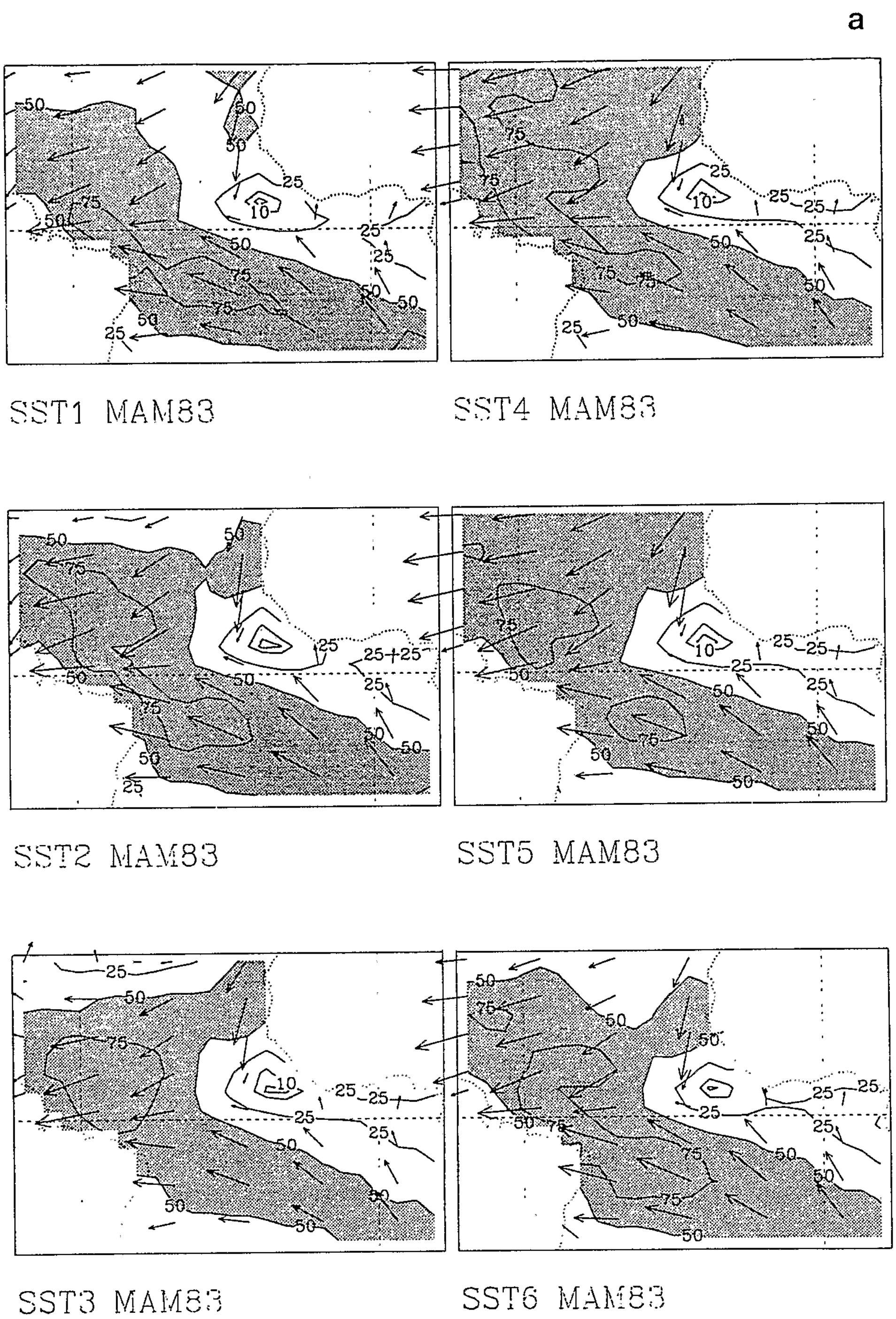


Fig. 3. Simulated wind-stress patterns for Spring (March–April–May) in 1983 (a) and 1984 (b) from 6 experiments with different initial conditions (AWR). The contour interval is 25×10^{-3} N/m^2 and shaded areas correspond to wind-stress amplitude over 50×10^{-3} N/m^2.

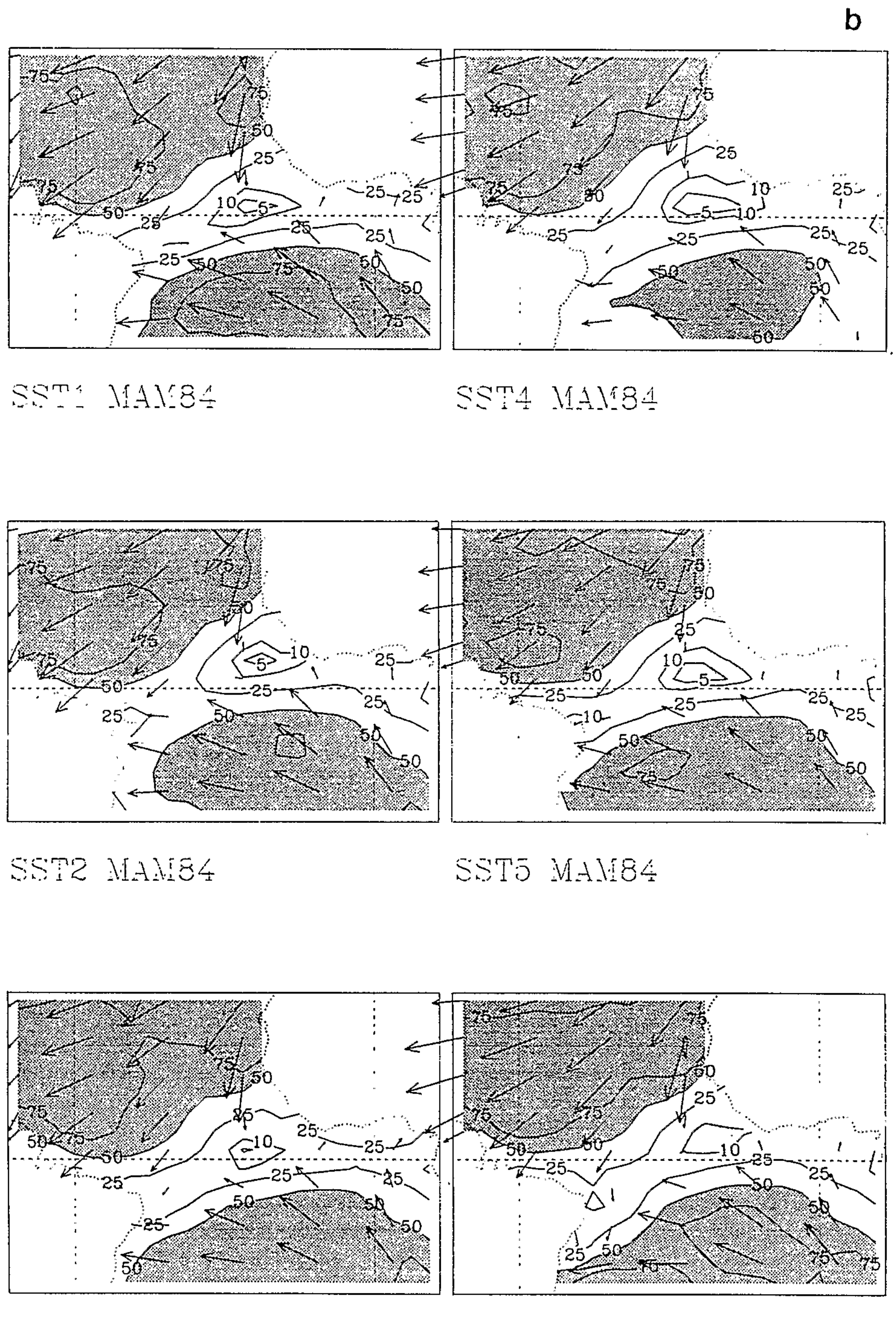

b
SST1 MAM84
SST4 MAM84
SST2 MAM84
SST5 MAM84
SST3 MAM84
SST6 MAM84

Fig. 3 (cont'd.)

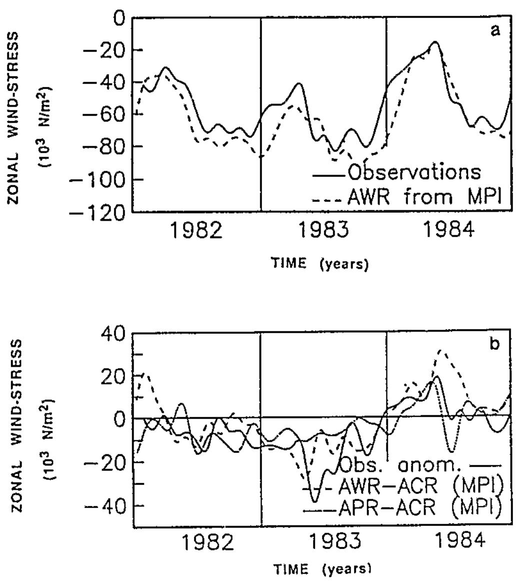


Fig. 4. (a) Time series of the observed (full line) and simulated by AWR (broken line) monthly zonal wind-stress (in 10^3 N/m^2) averaged over the western equatorial band (2°S–2°N, 20°W to the Brazilian coast) during 1982–1984. (b) Zonal wind-stress anomaly from monthly climatology in the same region. The continuous line corresponds to observations, the thick broken line to AWR simulation and the thin broken line to APR simulation. Note that the weakening of the zonal wind-stress from mid-83 to mid-84 is apparent on the three curves.

the Spring 1983. On the contrary, the Spring 1984 is marked by a strong weakening of the trade wind, located slightly off-shore.

In the APR run, the imposed SST related to the tropical Atlantic ocean follows the seasonal cycle. However, the wind-stress simulated in the APR experiment exhibits an interannual variability (Fig. 6), with the decrease in amplitude from mid-1983 and mid-1984, with a weaker amplitude than for the AWR experiment. The spatial pattern of the anomalies between May 1983 and May 1984 indicates an eastward displacement of the anomaly centre compared to the AWR simulation. However this displacement is not sufficient to explain the reduced amplitude of the strengthening of the wind-stress in 1983 in the selected test zone which is large enough to contain both anomaly centres.

Note that a very rapid increase in wind-stress appears in June 1984 with the APR run (Fig. 4): in July 1984, the simulated wind-stress is above normal in this run. One can anticipate that any warm SST anomaly simulated in OPR during the Spring 1984 will not persist until July as the return of the trades in APR will force the upwelling back.

The results from the comparison between the simulated and observed wind-stresses indicate that the atmospheric model (AWR) is in good agreement both in phase and amplitude with the ship-wind over the tropical Atlantic and it could be used to test the ocean variability. The interannual variability is not only locally constrained by the Atlantic SST and part of the Atlantic response is dictated by the Pacific variability. The results provided by APR seem also to go in the right direction, with a reduced amplitude and slightly different patterns. The impact of these wind-stress fields will now be tested in an ocean model.

As direct measurements of heat flux are still unattainable at oceanic basin scales, the heat flux patterns derived from the atmospheric runs can only be compared to available climatic products. A quick comparison with monthly means computed by Oberhuber (1988) proved that the heat flux values simulated by the AGCM runs are within the range of uncertainties. The fresh water budget is supposed to play a minor rôle in these first experiments compared to the wind-stress forcing.

3.2. The oceanic variability of the tropical Atlantic

When forced with observed atmospheric forcing, a model of the tropical Pacific ocean can simulate the interannual behaviour of this ocean (see, for instance, Philander and Seigel, 1985). The amplitude of the interannual signal is strong in the tropical Pacific and a clear coupled response amplifies over large areas in this basin. The amplitude of the interannual variability is smaller in the Atlantic than in the Pacific and is superimposed on a strong seasonal signal (Servain et al., 1985). Moreover the size of the tropical Atlantic and the continental influence prevent a strong interannual coupled signal to develop. Nevertheless, the interannual signal exists in the tropical Atlantic and it was particularly pronounced in 1984. This period is thus the first to test.

In order to validate the ocean model, two data sets are being considered. The first one is the analysed SST from ship observations, computed by Servain et al. (1987). It is a $2°\times 2°$ field of

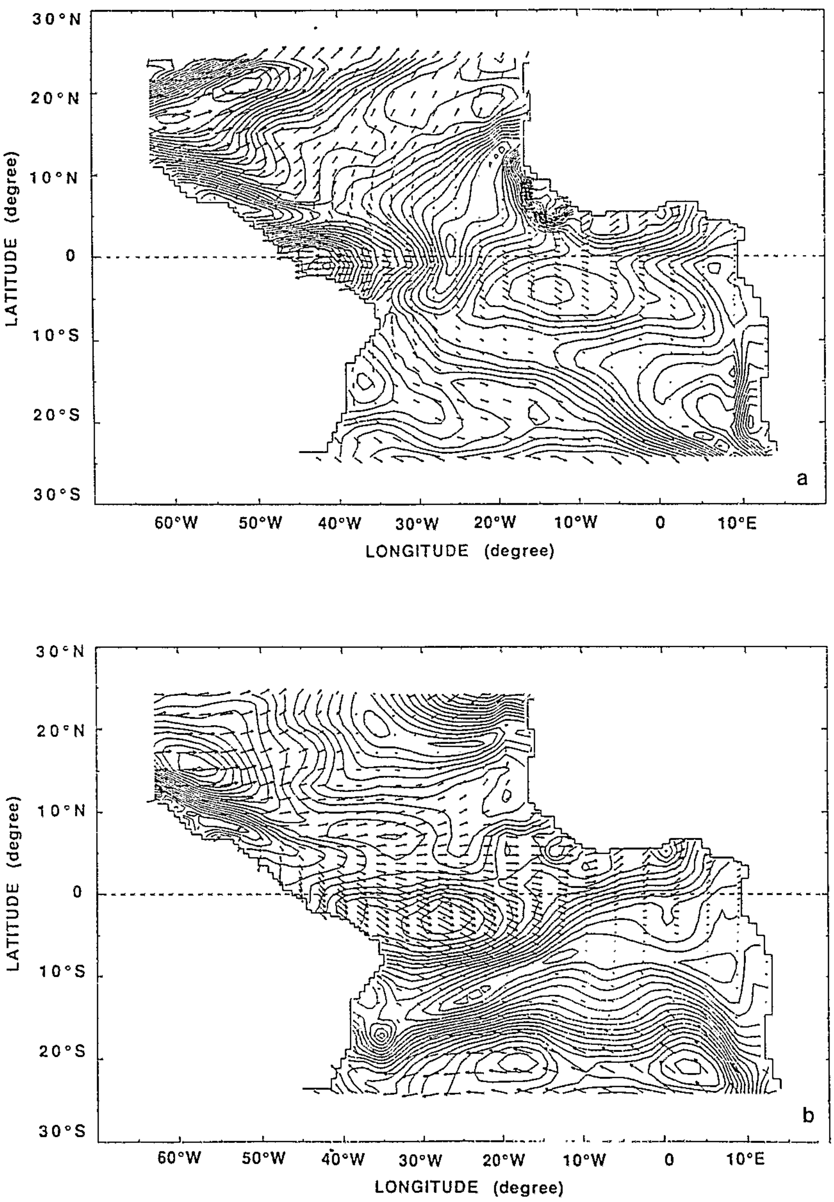


Fig. 5. Simulated wind-stress departure of AWR from climatology in May 1983 (a) and 1984 (b). The contour interval is 10^{-3} N/m^2.

monthly mean SST over the tropical Atlantic, from 20°S to 30°N.

A 2nd field is the analysis of the depth of the 20°C isotherm which traces the vertical displacement of the thermocline, perhaps except in the far eastern part of Gulf of Guinea. This data set has been obtained from the FOCAL/SEQUAL data set (Reverdin et al., 1991). It contains monthly analyses of the depth of the 20°C isotherm, on a $1° \times 5°$ grid, for the tropical area, from 1982 to 1984. The period 1983–1984 appears as a strengthening of the equatorial slope in the

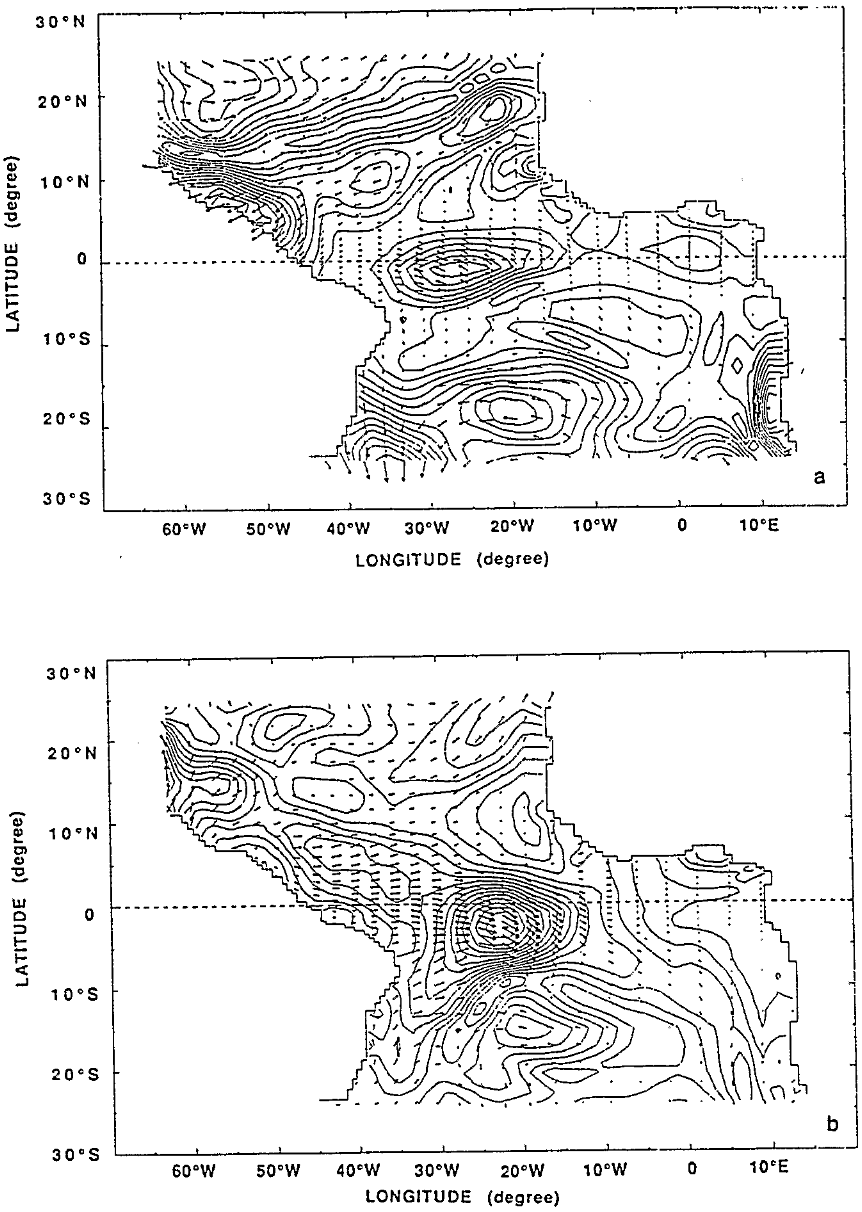


Fig. 6. Simulated wind-stress departure of APR from climatology in May 1983 (a) and 1984 (b). The contour interval is 10^{-3} N/m^2.

western Atlantic during the Spring–Summer 1983, followed by a flattening of the equatorial slope in the eastern Atlantic in the Spring 1984. The discussion for observed and simulated patterns will compare April 1983 and 1984. The observed signal has a consistent and distinctive pattern in space (Fig. 7), but the difference between the two maps is of the order of 20 m. It will be quite difficult to detect it with a numerical model, as the vertical spacing of the model is of the same order.

The ocean model has already been forced with the observed 6-day wind-stress computed by G. Reverdin for the FOCAL/SEQUAL period as reported in Blanke and Delecluse (1993). The

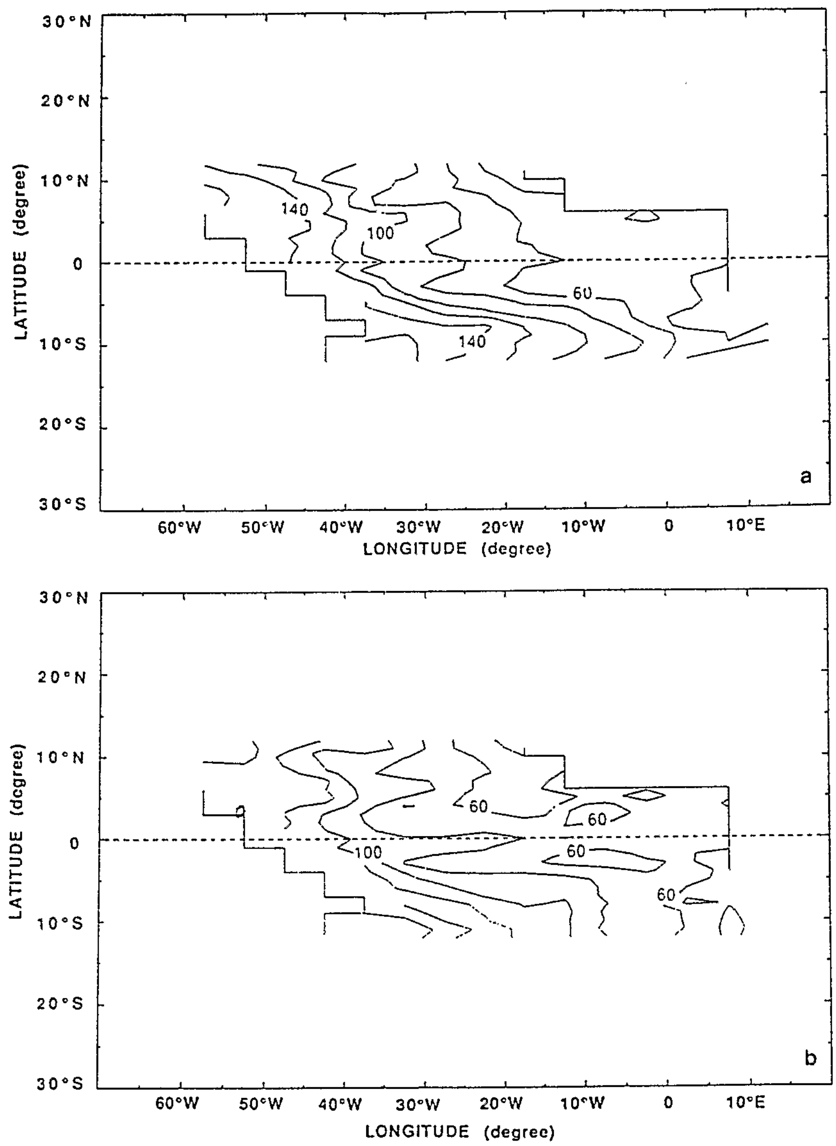


Fig. 7. Depth of the 20°C isotherm for April 1983 (a) and April 1984 (b) from observations. The contour interval is 20 m.

depth of the 20°C isotherm is presented in Fig. 8. The agreement between simulation (Fig. 8) and observation (Fig. 7) is quite good for the position of the 60 m and 80 m contours. The equatorial slope is underestimated by the model in the western part but the eastern part seems correct. The variability simulated in the Gulf of Guinea was also compared to the mooring located at 4°W along the Equator (Houghton and Colin, 1986) and it was clear that a depression of the isotherms started in the eastern equatorial Atlantic in November–December 1983, wiping out the small winter upwelling.

Before discussing the Pacific experiment (OPR), it is important to show that OWR is able to simulate the observed signal in 1983–84. When the

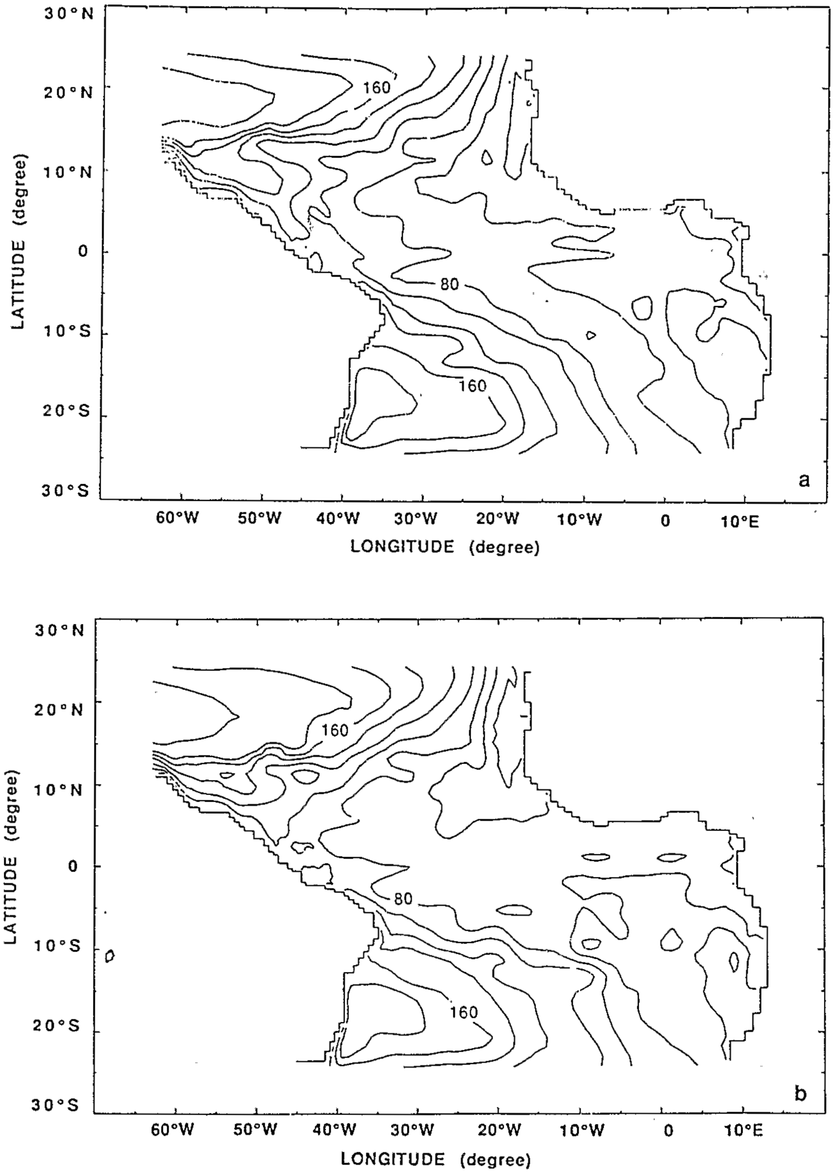
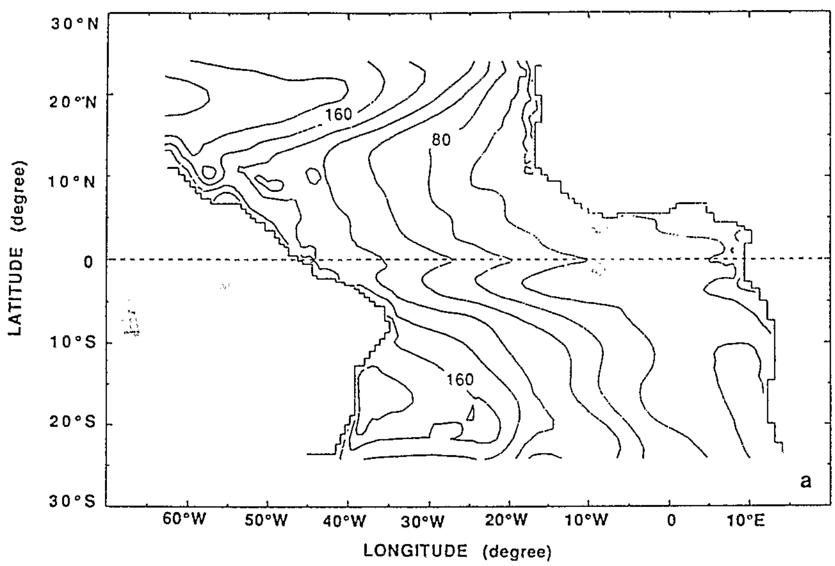


Fig. 8. Depth of the 20°C isotherm for April 1983 (a) and April 1984 (b) from oceanic simulations with observed wind-stress. The contour interval is 20 m.

oceanic model is forced by the AWR, the contour map of the 20°C isotherm is very close to the one obtained when the ocean is forced with observed data (Fig. 9). The same flattening appears in the Gulf of Guinea during the Spring 1984 and it seems that the interannual amplitude and phase of the equatorial anomaly at sub-surface levels is captured. At 4° W, along the equator, OWR ocean stays very warm (Fig. 10): there is no indication of the seasonal winter upwelling. The 20°C isotherm reaches its deepest level from February until June 1984. The summer upwelling starts again in July with reduced amplitude. This behaviour is in very good agreement with the observations which



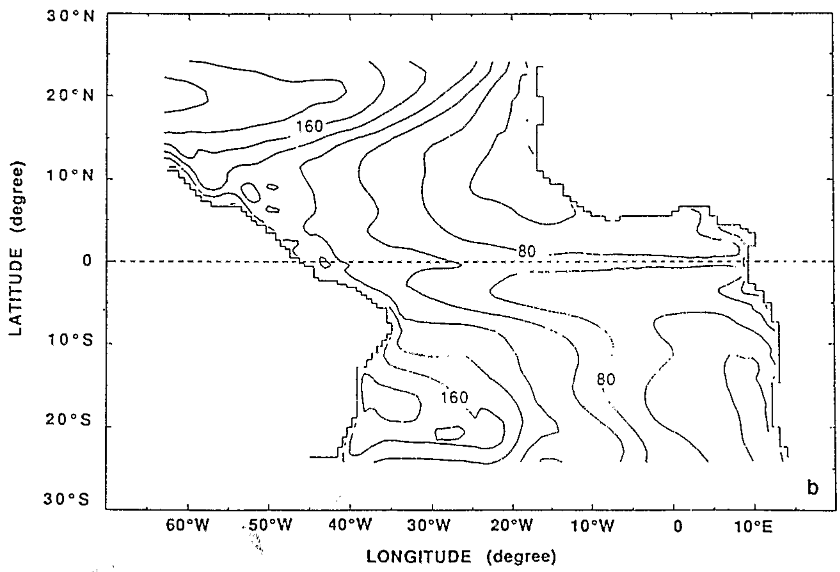


Fig. 9. Depth of the 20°C isotherm for April 1983 (a) and April 1984 (b) from OWR simulations. The contour interval is 20 m.

showed the very anomalous pattern of the thermocline during the Winter 1983–84. It proved that the quality of the AGCM and the OGCM is good enough to capture the interannual variability in the equatorial Atlantic.

With the APR forcing, the seasonal signal is dominant. The 20°C isotherm for the months of April 83 and 84 (not shown) does not indicate the flattening in the Gulf of Guinea. This result was disappointing and not totally convincing because we expected an oceanic anomaly in response to the wind-stress anomaly. The time-series of temperature at 4° W along the equator (Fig. 10) are in strong contrast between OWR and OPR: OWR

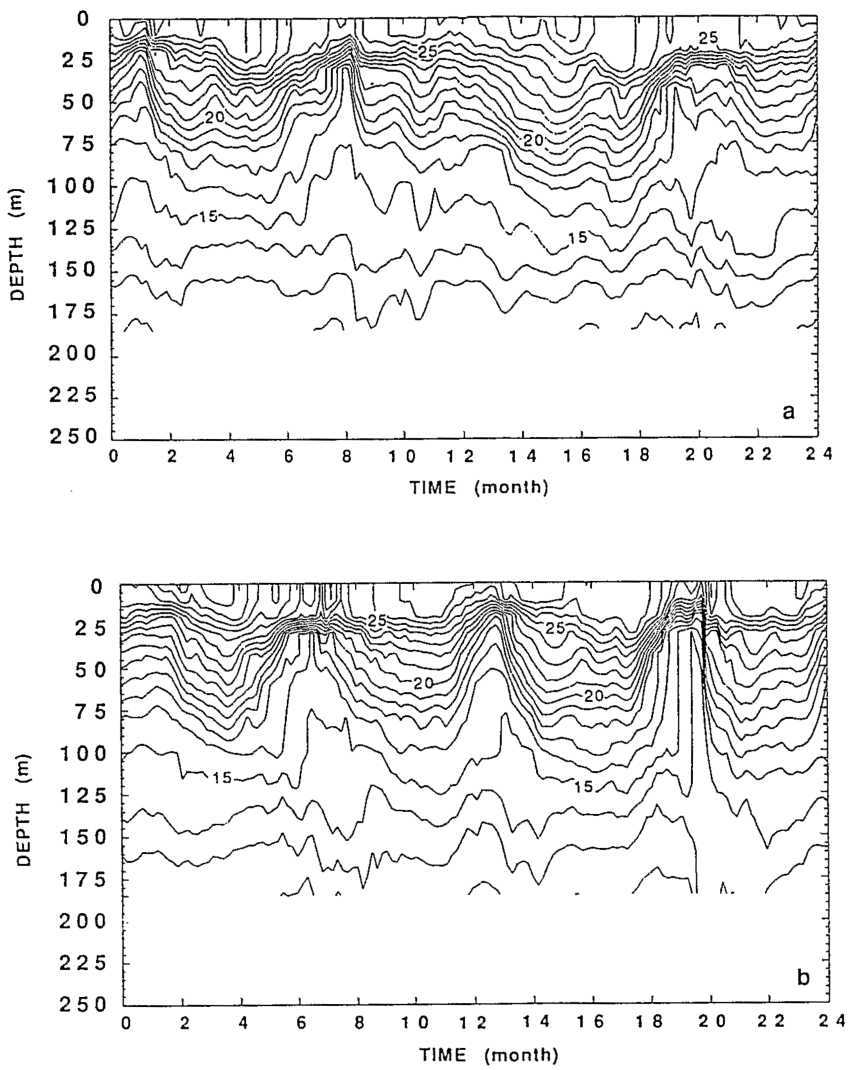


Fig. 10. Evolution in function of time (for 1983–1984) of the temperature profile at 0°, 4° W, for OWR simulation (a) and for OPR simulation (b). The contour interval is 1 °C.

shows an important deepening of the thermocline from the beginning of 1984 past the summer upwelling period which is much weaker than the previous year; OPR shows the preponderance of the seasonal cycle over the anomalies.

Comparisons between OWR and OPR simulations may suffer from a different space-time structure of the anomaly of the 20°C isotherm depth. In order to overcome this difficulty, a time-series of the difference of the depth of the 20°C isotherm between 1984 and 1983 was calculated, along the equator for both runs (Fig. 11). In OWR, a single anomaly pattern appears, increasing from January until May and moving (under some change) slowly eastwards. The amplitude is above 20 m from March to April, from 15° W to the coast where the maximum is above 60 m. From May, the amplitude of the anomaly decreases, still with an eastward propagation. The same time-series with OPR indicates a very different pattern. Two anomaly patterns cross successively the basin from west to east, the first one in February–March and

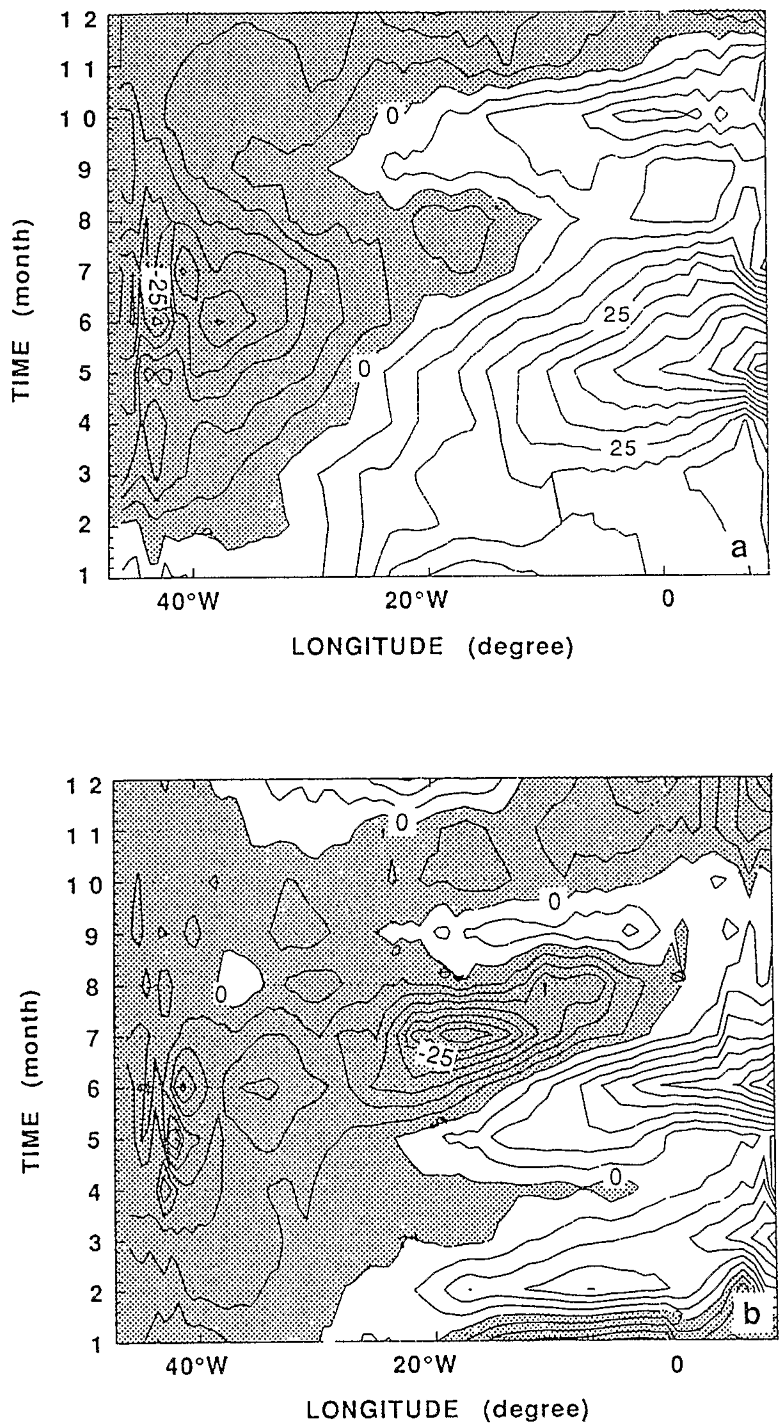


Fig. 11. Time-series along the equator of the difference of the depth of the 20°C between year 1984 and year 1983 for OWR (a) and OPR (b). Shaded area corresponds to a deeper level in 1983 than in 1984. The contour interval is 5 m.

the second one around June. Both these anomalies travel quickly eastwards with a speed slightly above 2 m/s, not far from the theoretical speed of oceanic Kelvin waves. The difference in the depth of the 20°C isotherm reaches 20 m for a brief period of time, in June 1984, east of the Greenwich meridian. These two downwelling events are remotely forced by the wind-stress anomalies but they look quite independent from each other and they do not show a low-frequency pattern in time: in April–May, OPR indicates a break in the anomaly pattern as OWR highlights a maximum.

These results suggest that externally initiated wind-stress anomalies are very important to initiate the interannual variability in the tropical Atlantic but this ocean is able to develop a coupled response of its own, which is implicitly present in AWR, forced by the worldwide SST patterns. This SST coupled response is of course missing in APR, where only the remotely forced atmospheric wind-stress is present. The coupled response of the Atlantic ocean is found surprisingly high: it can triple the wind initiated anomalies. To explore further the answer of the tropical Atlantic ocean, it is necessary to develop a coupled model in order to study the ability of the Atlantic ocean to develop a coupled response of its own. The coupled response of the equatorial Atlantic ocean has been recently explored by Zebiak (1993). His results confirmed the fact that part of the equatorial Atlantic variability can be explained by the equatorial coupled mode but it cannot explain the entire interannual variability of the Atlantic which has to be sustained by external factors. The present work confirms the importance of the tropical Pacific SST as an external factor.

The comparison between the SST simulated by the model and the observed SST from ship data is difficult because it superimposes the error contained in the ocean physics and the error from the heat budget of the atmospheric model. The observed fields themselves are not free from error sources. The mean position of the isotherms is well located (for instance, the warm water delimited by the 26°C isotherm follows nicely the ITCZ) but systematic biases appear in the Gulf of Guinea. The model, in all simulations, indicates a strong front, just north of the equator, from the middle of the Atlantic ocean till the coast in July. A pool of warm water always separates the equatorial upwelling from the coastal upwelling along Nigeria-Ivory Coast. This front does not appear in the observations where a connection exists around 0° longitude. The simulated upwelling is slightly warmer than the observed one and is longitudinally stretched. The coastal upwelling, along the east African coast, is systematically pushed southward in the simulation. These biases

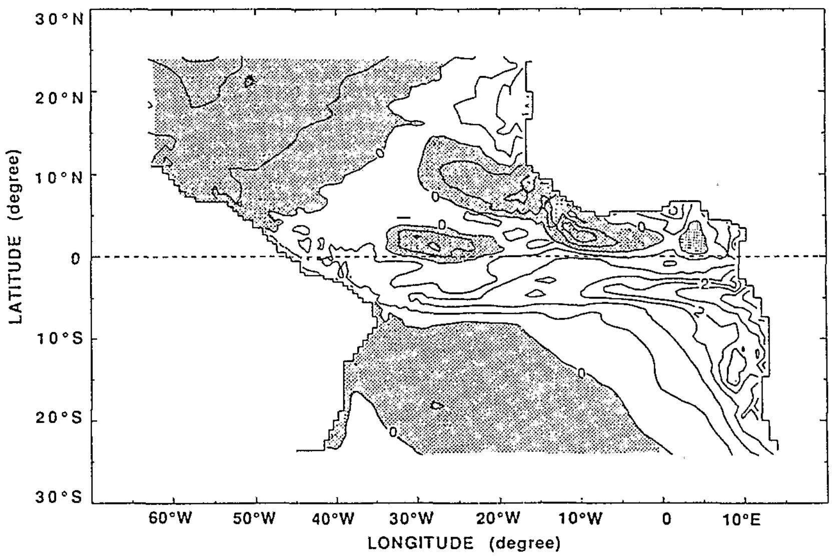


Fig. 12. SST anomaly (1984 minus 1983) for July in the OWR simulation. The contour interval is 0.5°C. The shaded area corresponds to negative anomalies

make the comparison difficult. Nevertheless a difference map between July 1983 and July 1984 indicates a very clear warm anomalous pattern in the southeast equatorial Atlantic with OWR (Fig. 12). The warm waters extend from the equator to 8° S in the west, and to 30° S in the east. The maximum of anomaly is located around 4°–5° S, with an amplitude reaching 2.5°C at the coast and 2°C around the Greenwich meridian. Note that the July map for OPR indicates the return of the summer upwelling and does not show the 1984 warming: the response of the ocean, located in sub-surface, was too weak to generate a surface response, in a forced mode.

4. Conclusion

This experiment was very instructive about the present ability of ocean and atmosphere GCM's to simulate interannual variability in the tropical Atlantic. Forced with globally observed SST, the MPI ECHAM3 atmospheric model simulates the evolution in amplitude and phase of the Atlantic trade winds in very good agreement with ship observations. The increase of trades during 1983 and the weakening at the beginning of 1984 is well captured. The oceanic model, forced by this simulated wind-stress, is also successful. The sub-surface levels of the ocean follow closely the wind-stress variations and develop both a remotely forced and a locally amplified response.

The Pacific SST anomalies generate a wind signal in the atmospheric model run as far as over the tropical Atlantic from 1983 till the beginning of 1984. This wind signal bears some resemblance with the observed wind-stress anomalies. However, it has a reduced amplitude and is displaced eastwards. It generates an anomalous sub-surface signal in the tropical Atlantic ocean, associated with a flattening of the thermocline, travelling eastward as a Kelvin wave front. These events cross quickly the basin but the amplitude remains small and within the equatorial waveguide. The tropical Atlantic ocean cannot develop its full response in this forced run.

A closer investigation of the role of the Pacific anomaly would therefore require a coupled model in order to let the air–sea interactions amplify the response. However, the variation of the depth of the 20°C isotherm (which represents the thermocline signal) seems to act as a catalyst for the interannual Atlantic signal, as the thermocline signal preceeds the establishment of coupled surface modes in El-Niño episodes. This result is in agreement with Zebiak's analysis (1993) of the Atlantic coupled system where the primary mechanism underlying the oscillator behaviour in the Atlantic ocean resembles the delayed oscillator mode of his Pacific coupled model. As a potential source for external variability, the SST of the tropical Pacific is obviously a serious candidate. The present work shows clearly that the Pacific anomalous behaviour during the 1982–83 episode was able to initiate anomalous patterns in the tropical Atlantic through disturbances in the atmospheric circulation whose details remain to be studied.

5. Acknowledgments

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