Evidence of Remote Forcing in the Equatorial Atlantic Ocean

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ABSTRACT

An analysis of sea-surface temperature (SST) and surface winds in selected areas of the tropical Atlantic indicates that the nonseasonal variability of SST in the eastern equatorial Atlantic (Gulf of Guinea) is highly correlated with the nonseasonal variability of the zonal wind stress in the western equatorial Atlantic. A negative (positive) anomaly of the zonal wind stress near the north Brazilian coast is followed by a positive (negative) SST anomaly in the Gulf of Guinea about one month later. Furthermore, the correlation between the local wind stress anomaly and SST anomaly in the Gulf of Guinea is considerably smaller. These preliminary results indicate that remote forcing in the western equatorial Atlantic Ocean is an important factor affecting the eastern equatorial Atlantic sea-surface temperature. Recent equatorial theories are consistent with these observations.

1. Introduction

The variations of temperature in the eastern equatorial Atlantic (Gulf of Guinea) have a significant effect on the abundance of fish (Bakun, 1978; A. Fonteneau, pers. comm., 1981) and on the rainfall distribution along the western African coast and the sub-Saharan countries (Lamb, 1978). For this reason, understanding the dynamics of this region is of great interest to both meteorologists and oceanographers. Several ideas have recently been proposed to explain these fluctuations of temperature. Philander (1978, 1981) argues that the local wind forcing may account for this variability. Moore *et al.* (1978) suggested that remote forcing by winds in the western Atlantic may provide an alternate explanation. In this study we present evidence, based on historical SST and wind data, that supports the latter idea.

2. Data and processing

Merchant ship data are routinely collected at the U.S. National Climate Center at Ashville, NC. There they are put in the form of individual monthly averages of various meteorological parameters for 1° latitude × 1° longitude squares in the tropical Atlantic. Full details of this data set are given in Hastenrath and Lamb (1977). These authors have further processed this data to produce 60-year time series (1911–1972) of the field of wind and SST averaged over 5° squares, and have kindly offered us this derived data set.

Fig. 1a shows selected areas of relevance to our study: a Brazilian region (BR) 5°N–5°S, 35–25°W, a north Guinea region (GN) 5°N–0°, 10°W–10°E, a west Guinea region (GW) 5°N–5°S, 10–5°W, a northeastern region (NE) 30–25°N, 20–15°W, and a southeastern region (SE) 30–25°S, 10–15°E. These regions have been chosen to test, in the most simple way, the effect of remote or local wind forcing on sea-surface temperature in areas where the distribution of the data along shipping lanes are the most favorable. Fig. 1b indicates the temporal distribution of wind data in a part of the BR area. According to this graph there are two periods of extensive data coverage: from 1923 to 1938 (between the two world wars), and from 1964 to 1969. For this reason, our analysis is limited to these two time periods.

Because the annual cycle is strong in all the data sets, we expect to have a high correlation between all the variables. If we remove the monthly mean, any significant correlations we find between the resulting nonseasonal data sets will be a more sensitive indicator of the dynamical processes involved in the response. However, removing the monthly mean does not imply that we have completely suppressed the annual cycle. The anomalous years are due, in part to an increase or decrease in amplitude, or to a shift of phase, of the normal seasonal cycle. This property is clearly illustrated in Fig. 2. An alternate method of suppressing the annual cycle is to low-pass filter the time series. However, this method will also elim-

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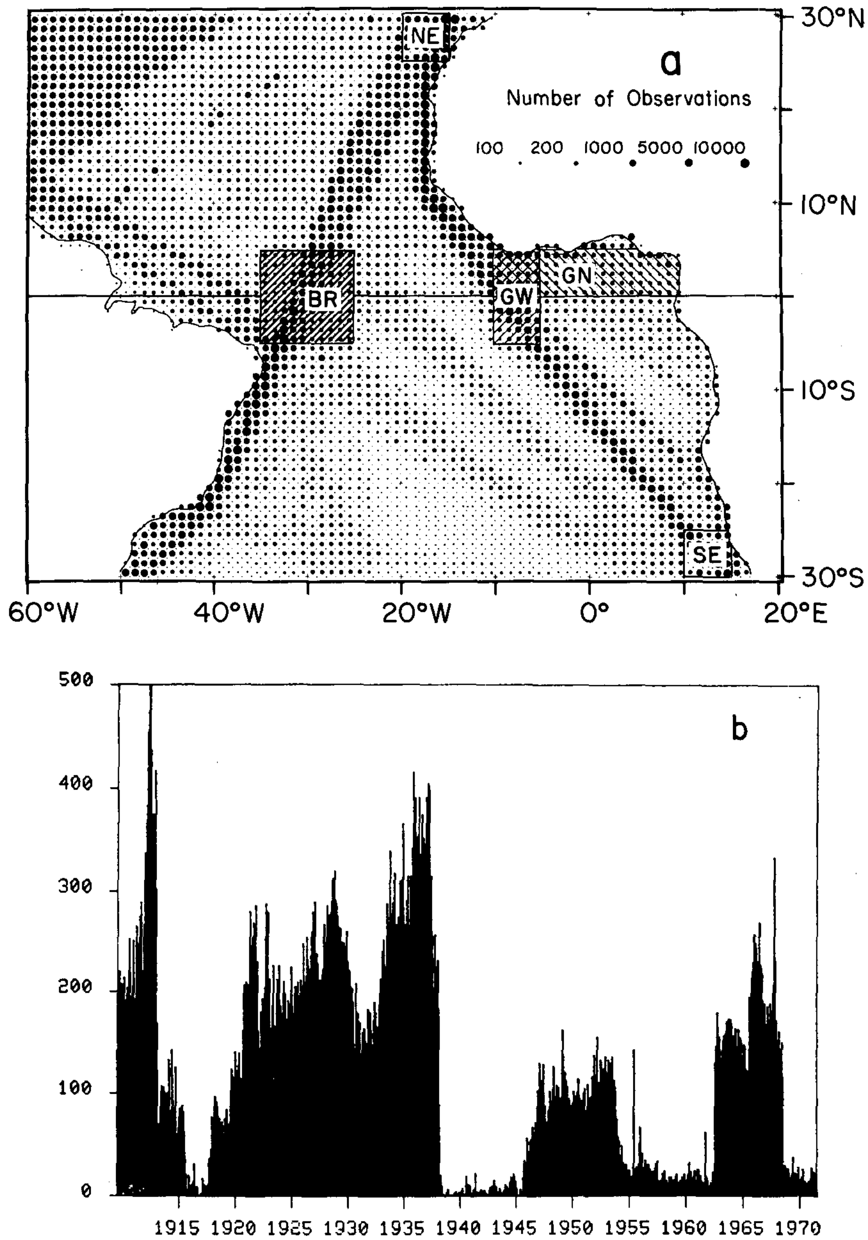

FIG. 1. (a) Position of the studied areas (composite picture from Hastenrath and Lamb, 1977). (b) Number of observations by month in a part of the BR area (0–5°S, 30–35°W).

inate the short time-scale anomalous events evident in Fig. 2. Therefore, the commonly used method of nonseasonal decomposition is preferable in the present case, and so is used here.

The nonseasonal variations of the zonal (τ^x) and meridional (τ^y) components of the wind stress and SST were computed for the 5° square areas by subtracting out the monthly mean values. The monthly means were found by averaging over the whole 1911–1972 period, excepting those months where less than 10 observations are available. A drag coefficient of 2.0×10^{-3} was used for the calculation of the wind stress. To find the mean in each of the selected regions (BR, GN, GW, NE, SE), individual 5° square mean values were weighted according to the mean number of observations in these areas.

3. Results

Fig. 3 shows the interannual anomalies of SST in the NE, SE, GN and GW regions. The Gulf of Guinea areas (GN, GW) appear to share a common

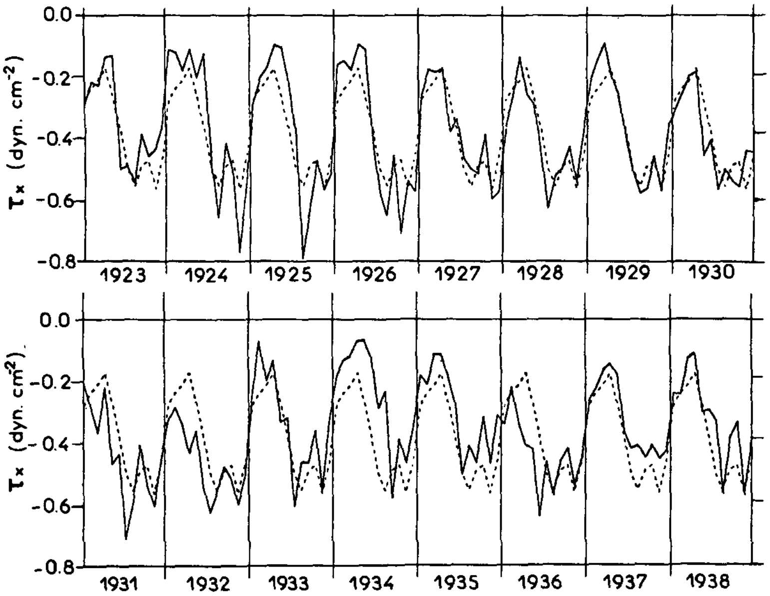

FIG. 2. Monthly mean of τ^x in BR area. Solid line: individual years from 1923–1938. Dotted line: mean year on the period 1911–72.

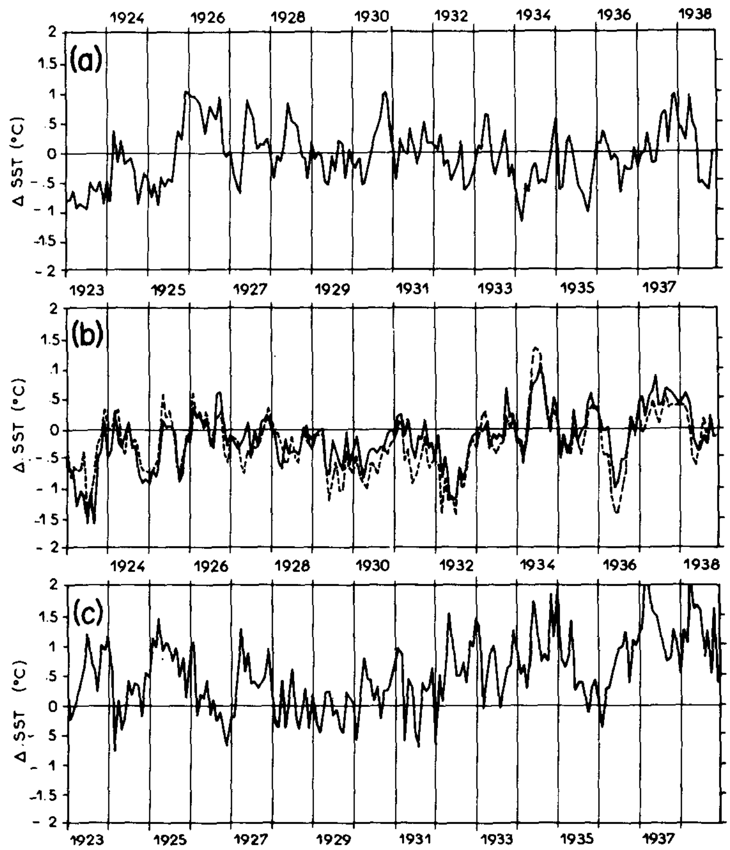

FIG. 3. Monthly anomalies of SST from 1923 to 1938. (a) Northeastern area (NE). (b) Guinea Gulf (dashed line GW, solid line GN). (c) Southeastern area (SE).

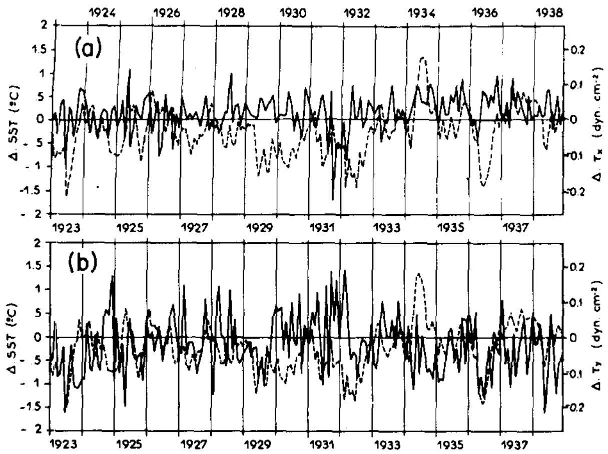

FIG. 4. Monthly anomalies from 1923 to 1938 in the western Guinea Gulf (GW area). (a) Solid line τ^x, dotted line SST. (b) Solid line τ^y, dotted line SST.

interannual variability of SST (Merle *et al.*, 1980). On the contrary, the anomalies of SST in the north and southeast tropical regions (NE, SE) are very different (Fig. 3), and are both different from those of the Gulf of Guinea. Thus, we can consider the SST anomalies to have a spatial extent affecting the whole eastern equatorial basin and to be characteristic of this region.

According to the qualitative model of Cromwell (1953), variations in the northward and westward wind stress near the equator should induce local temperature fluctuations there. Furthermore, according to Philander (1978) the northward equatorial wind can affect SST along the northern coast of the Guinea Gulf. So, following these authors, we might expect to have a good correlation between the τ^x and τ^y anomalies in the western Guinea region (GW) and the corresponding SST anomalies. In Fig. 4a a positive (negative) peak in the τ^x anomaly corresponds to a decrease (increase) in the westward wind stress, and for local forcing we should expect an increase (decrease) in SST anomaly. In Fig. 4b a positive (negative) peak in the τ^y anomaly corresponds to an increase (decrease) in the northward wind stress and for local forcing we should expect a decrease (increase) in the SST anomalies. Therefore the two curves in Fig. 4a would be in-phase and the two curves in Fig. 4b would be out-of-phase. The poor visual agreement between these curves is confirmed by a lagged regression correlation analysis (Fig. 5). For zero lag the correlation coefficient is on the order of the noise level. However, notice a surprising maximum of correlation coefficient at roughly three months lag. We will comment on this feature in the last section.

According to the remote forcing idea of Moore *et al.* (1978), variations in the westward wind stress in the western equatorial Atlantic should induce temperature fluctuations in the eastern equatorial Atlantic. So, we expect to have a good correlation between the τ^x anomaly in the Brazilian region (BR) and the SST anomaly in the Guinea Gulf (GW or GN). In Fig. 6 a positive (negative) peak in the τ^x anomaly of BR corresponds to a decrease (increase) in the westward wind stress anomaly, and for remote forcing we should expect an increase (decrease) in

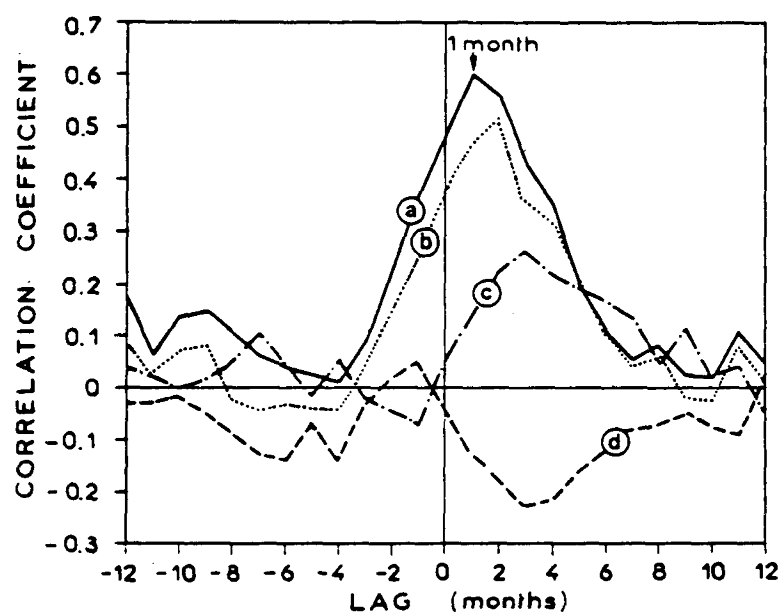

FIG. 5. Correlation with time lag. Curve a: τ^x–BR with SST–GW. Curve b: τ^x–BR with SST–GN. Curve c: τ^x–GW with SST–GW. Curve d: τ^y–GW with SST–GW.

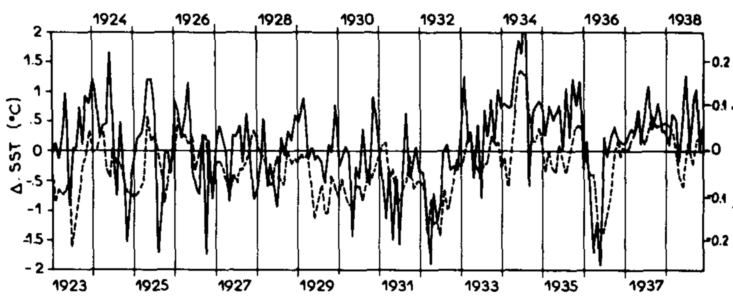

FIG. 6. Comparison of monthly anomalies from 1923 to 1938 of τ^x in BR area (solid line) and SST in GW area (dotted line).

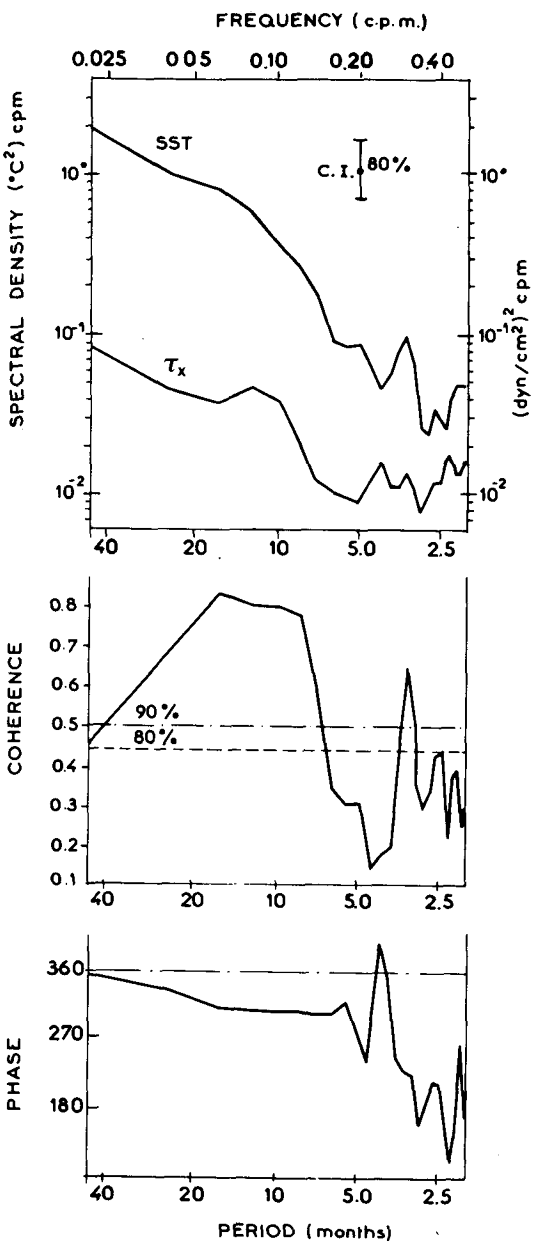

FIG. 7. Cross spectrum of τ^x in BR area and SST in GW area.

SST anomaly of GW. Therefore these two curves would be in-phase. The lagged correlation analysis confirms the good visual agreement between these two curves (Fig. 5). A highly significant maximum of correlation coefficient is obtained with a one month lag, the wind stress anomaly preceding the SST anomaly.

These last results are supported by a cross-spectral analysis of the same data set (τ^x – BR and SST-GW, 1923–38). Despite the absence of marked peaks in the power spectral densities, the coherence is characterized by maxima around a 3-month period and from 7 to 24-month periods (Fig. 7). For this high coherence we have plotted the phase difference in months versus the frequency and added the corresponding error bar for the 80% confidence level (Fig. 8). It is clear from this figure that a phase lag of one to two months occurs over a broad frequency range, and so justifies *a posteriori* our lagged correlation analysis.

4. Discussion and conclusion

This analysis shows that the SST anomaly in the eastern equatorial Atlantic is dominated by remote

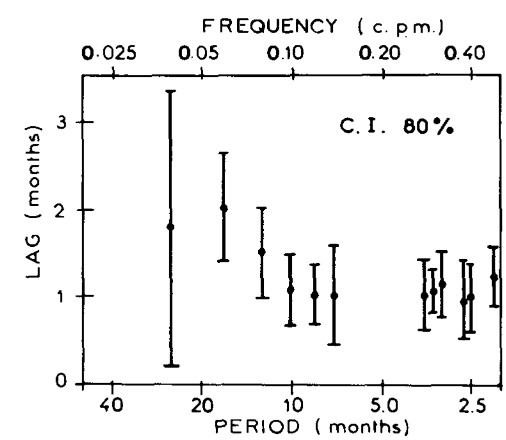

FIG. 8. Time lag versus frequency of the cross-spectrum of τ^x in BR area and SST in GW area (Fig. 7). Only frequencies where coherence is high are considered. The confidence interval is for probability of 0.8.

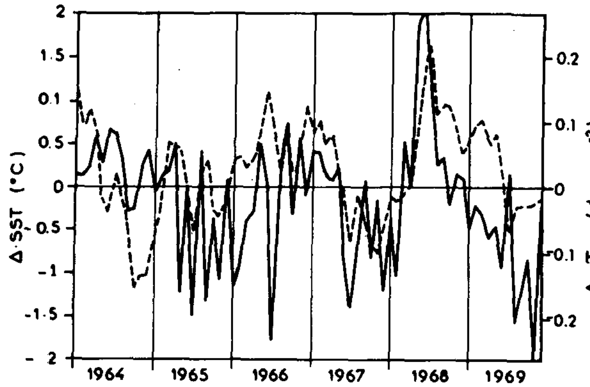

FIG. 9. Comparison of monthly anomalies from 1964 to 1969 of τ^x in BR area (solid line) and SST in GW area (dotted line).

wind forcing in the western equatorial Atlantic. The nonseasonal variability of SST in the Guinea Gulf is poorly correlated with the nonseasonal local wind forcing and highly correlated with the nonseasonal variability of the wind forcing off Brazil. Furthermore, a time lag of roughly one month exists between this remote forcing and its response in the Guinea Gulf.

Moore *et al.* (1978) proposed a simple remote-forcing mechanism to account for the Guinea Gulf upwelling. According to this hypothesis an increase of the easterly wind in the western equatorial Atlantic excites an internal upwelling equatorial Kelvin wave that propagates into the eastern equatorial Atlantic. When this disturbance reaches the eastern boundary it splits into Kelvin waves propagating poleward along the coasts in both hemispheres and a large number of westward propagating equatorial Rossby waves. Thus, the lag between the response in the Gulf of Guinea and the winds off Brazil is simply related to the speed of Kelvin waves. This theoretical idea has been illustrated in the numerical, one-and-a-half layer model studies of O'Brien *et al.* (1978) and of Adamec and O'Brien (1978), in which the wind in the western Atlantic was switched on impulsively. In contrast, Cane and Sarachik (1981) studied the response of a single baroclinic mode to periodic forcing, and found that the equatorial Kelvin wave, and associated time lag, was completely masked by reflected Rossby waves. However, in three-dimensional models remotely forced by periodic winds the equatorial Kelvin wave is again clearly evident (Philander and Pacanowski, 1981; McCreary *et al.*, 1982). In any case, the rapid increase of the easterly wind in spring 1979 at St. Paul Rocks (0°55′N, 29°21′W) (Katz *et al.*, 1981), which is not unusual (Fig. 2), seems to indicate that impulsive forcing is most relevant to the present study.

The mean lag observed here implies a wave propagation speed of ~ 1 m s^{-1}, which is slower than the phase speed of the first baroclinic mode Kelvin wave. It compares better to the speed of the second mode or to a combination of several low-order modes. An interesting aspect of Fig. 8 is that there is an apparent increase of this lag with the period of the forcing. In a recent model, which includes many baroclinic modes, J. P. McCreary (pers. comm., 1981) has shown that lower-frequency forcing excites higher-order modes in the Kelvin wave and so decreases the apparent horizontal phase speed, in agreement with our observations.

Similar results hold for the mean seasonal signal as well. Picaut (1982) shows poleward propagation of the mean seasonal coastal upwelling at speeds slower than the first baroclinic mode. In addition, south of Abidjan there is a vertical propagation of the upwelling signal. This upward phase shift is clear evidence that the coastal signal is a combination of modes, and is also a feature of the model of McCreary *et al.* (1982).

A correlation analysis, not shown here, reveals that the anomalies of the wind stress in the western equatorial Atlantic are highly correlated with the wind stress anomalies in the Guinea Gulf with a lag of two months. This property could explain the intriguing 3-month lag obtained for the maximum of correlation between the SST anomalies in the Gulf of Guinea and the local wind stress anomalies (Fig. 5). Consider the following scenario. A wind anomaly appears first in the Gulf of Guinea, but apparently does not always significantly affect the ocean there. Two months later a similar wind anomaly appears in the western equatorial Atlantic. Only after a delay of one more month is the ocean in the Guinea Gulf affected by the remote forcing.

Finally, it should be pointed out that there is a similarity between the temperature variability of the eastern equatorial Atlantic and Pacific Oceans. The "El Niño" phenomenon in the eastern equatorial Pacific is probably another dramatic example of the response of the equatorial ocean to remote wind forcing (Wyrtki, 1975; McCreary, 1976; Hurlburt *et al.*, 1976; Busalacchi and O'Brien, 1981). A striking example of such warming in the Gulf of Guinea was particularly evident during the summer of 1968 (Hisard, 1980; Merle, 1980) and can be explained by abnormal wind stress forcing off the northern Brazil coast (Fig. 9).

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