DIAGNOSTIC STUDIES OF THE THERMAL STRUCTURE OF THE TROPICAL ATLANTIC OCEAN

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1. INTRODUCTION

1.1 Tropical Atlantic ocean and climate studies

The tropical regions cover a large part of the globe and play a key role in the large scale interactions between the atmosphere and the oceans, on a global basis. Changes in SST conditions in the tropical areas have been shown to have an influence on the global atmospheric circulation (HOREL and WALLACE - 1981 ; PAN and OORT-1983). A number of General Circulation Models (GCM) experiments led to the same conclusions. The tropical oceans play also a major role in meridional heat transport (OORT and VONDER HAAR - 1976). Furthermore the vicinity of the equator and the vanishing of the Coriolis paramameter imply that a stratified ocean can respond strongly and rapidly to basinwide wind fluctuations and therefore the wind forcing is responsible for a large part of the changes of the thermal structure and of the Sea Surface Temperature (SST) on a seasonal and interannual time scale.

Several sequential studies of the steady state and of the low-frequency variability of the tropical Atlantic ocean have been carried out during the last decades (EQUALANT in 1962-63, GATE in 1974, FGGE in 1979 and more recently SECTION and SEQUAL/FOCAL). The two last experiments had the objectives of understanding the seasonal response of the global tropical Atlantic basin to the seasonally varying winds. This scientific interest for the Atlantic ocean has many justifications. Its direct contribution to the forcing of the global atmosphere is certainly weaker than that of the Pacific and Indian ocean, but it cannot be omitted and furthermore it represents and ideal case study of tropical oceans dynamic (laboratory). The three oceans are dynamically similar and of the three the Atlantic ocean is the most predictable, potentially be.cause of its limited zonal extension and the relative simplicity of the wind field. As a consequence, the response of the tropical Atlantic is almost in phase with the wind changes, which are usually very regular and dominated by the annual signal, except particular years, when abrupt wind bursts excite equatorial waves that are believed to have a relation with the observed warm and cold SST anomalies in the Gulf of Guinea (SERVAIN, PICAUT and MERLE - 1982 ; SERVAIN, PICAUT and BUSALACCHI - 1985).

This logistically compact tropical Atlantic ocean has a number of other virtues. It is more accessible and in several respects better known than the other tropical oceans. The past experiments mentionned above, and particularly SEQUAL/FOCAL, have provided what can be considered as the best tropical basinwide data set documenting the low frequency response of a global ocean to the wind forcing. A number of African and South-American countries are suffering severe climatic conditions (Chilian, North-Brazil and Sahel droughts) which can be partially considered as the result of particular thermal conditions of the tropical Atlantic ocean and possibly could be forecasted at short time scales (MOURRA and

SHUKLA - 1981). Finally the Atlantic ocean is a meridional ocean and its tropical part is one of the principal meridional heat exchange region of the world's oceans and so has an important influence on the global earth's climate at longuer time scales (Fig.1).

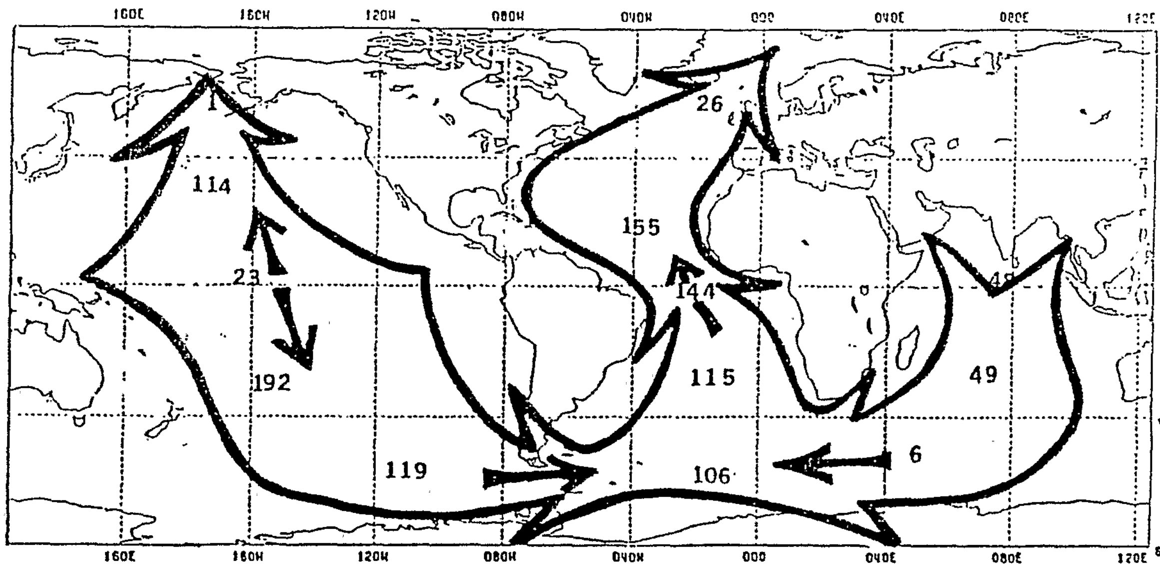


FIG.1 The ocean mean heat transport in 10^{13}W from STOMMEL(1979)

1.2 Two forcings and two responses of the tropical oceans.

A thin thermocline layer is almost permanent in the tropics. It separates a well mixed and warm superficial layer from the deeper cold waters in an almost perfect two-layers ocean. This thermal structure leads to separate the response of the tropical ocean into two parameters : (i) the depth of the thermocline, (ii) the temperature of the mixed layer (or of the Sea Surface). These parameters are, at the first order, independents. The depth of the thermocline is the dynamical response of the ocean to the wind forcing (either local, remote or global). The temperature of the mixed layer or SST is the result of the local thermodynamic forcing of the atmosphere. These simplified views apply especially in the western regions where the thermocline is deep. In the Eastern equatorial regions where the thermocline is shallow and reaches the surface it creates particular surface conditions (cold water) that in turn modify the conditions of the local thermodynamic forcing (the ocean absorbs heat from the atmosphere). And there the two forcings and the two corresponding responses of the ocean are clearly not independent.

Nevertheless understanding the response of the ocean in the tropics require, in a first approach, to study the oceanic dynamical response (depth of the thermocline) to the wind forcing. The second order phenomenon is the thermodynamic coupling (forcing and response) that is almost entirely confined in the oceanic mixed layer.

We propose a synthetic overview of the present knowledge on the large scale, low frequencies forcings and responses of the tropical Atlantic ocean. In chapter two the forcings over the tropical Atlantic are described. In chapter three, the dynamical response of the ocean. In chapter four, the heat budget and heat transport. In chapter five, the SST response. In chapter six, conclusions are derived.

2. THE FORCINGS

Several million surface observations from merchant and weather ships, compiled by the National Climatic Center and desiganted TDF-II, have been used by many authors to compute various meteorological elements of the surface oceans.

In the tropical Atlantic HASTENRATH and LAMB (1977) produced a climatological Atlas including monthly charts of resultant wind, divergence, wind steadiness, wind-stress curl, relative vorticity. HELLERMAN (1980) computed the annual mean and standard deviations of wind-stress, wind-stress curls and divergence of Ekman transport in the tropical Atlantic. BUNKER (1976) and HASTENRATH and LAMB (1978) also calculated the individual energy fluxes across the surface to obtain monthly and annual maps of these parameters. They also computed the resulting net energy flux across the surface.

2.1 Wind forcing

The wind stress means for March and August are presented Fig.2 from HELLERMAN (1980). The northeast and southeast tradewinds are seen to prevail over the western part of the Atlantic ocean. Along the African coast and in the Gulf of Guinea, the winds are predominantly meridional. The tradewinds converge on the

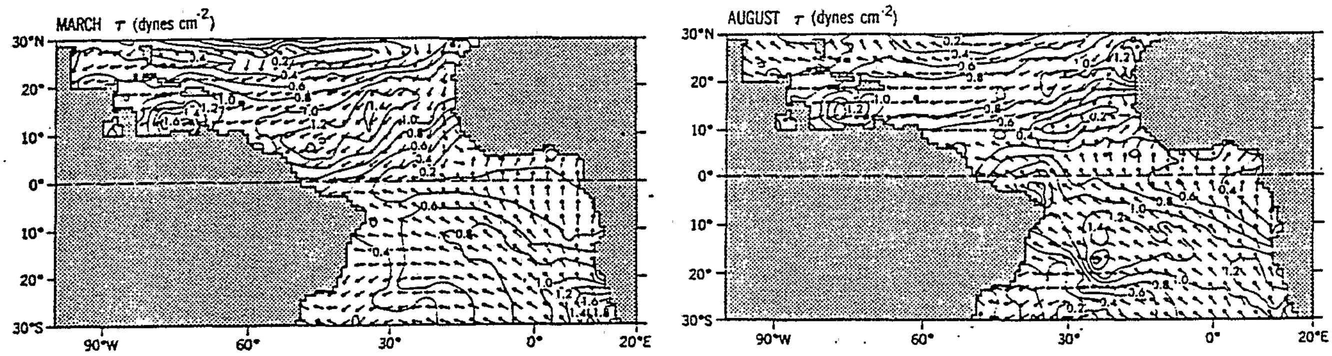


FIG. 2 Mean Wind-stress for the months of March and August (from HELLERMAN - 1979)

Intertropical Convergence Zone (ITCZ) which present a large seasonal migration. In March it is practically along the equator, in August it is in its northernmost position from 10°N in the west to 15°N in the East.

The seasonal and interannual variability of the zonal and meridionnal component of the wind-stress have been studied more recently by SERVAIN, PICAUT and BUSALACCHI (1985). They use a sixteen years SST and wind data set monthly averaged from January 1964 to december 1979 issued from the TDF-II data file. Regions of the maximum seasonal variability of the zonal and meridional wind stress (Fig.3 upper panels) are contained within an envelope defined by the seasonal excursion of the ITCZ. The maxima of the zonal wind stress fluctuations straddle the mean position of the ITCZ whereas the largest deviations of the meridional wind stress occur along this line. Interannual wind stress fluctuations do not necessarily coincide with the regions of the maximum seasonal variability. The largest year to year changes in the zonal wind stress are at the poleward extremes of the tropical area. The amplitude of the interannual variability of the meridional wind stress is nearly constant. Minima are however noted inside the gulf of Guinea and offshore the coasts of Guiana and Brazil (Fig.3 lower panels).

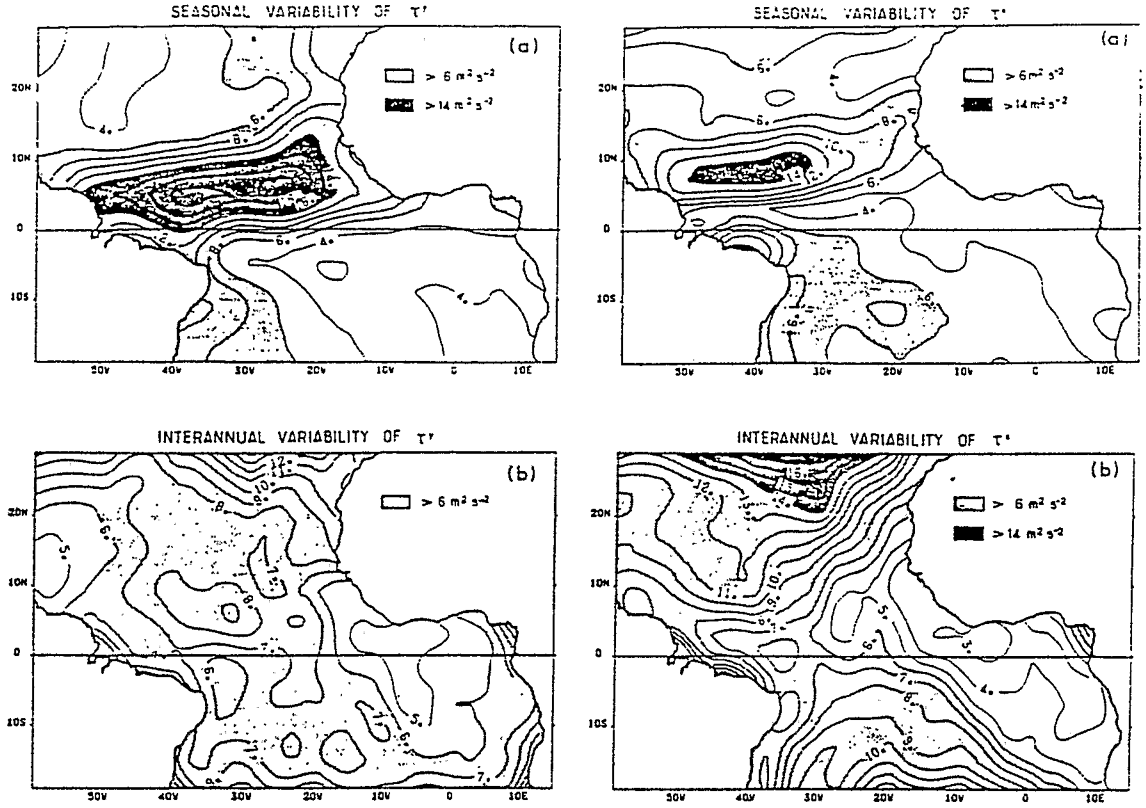


FIG. 3 Seasonal and interannual variability of the zonal and meridional component of wind-stress(from SERVAIN et al - 1985)

2.2 Heat flux forcing.

The energy exchange at the ocean surface has been computed first by Budyko (1963) from the equation :

$$A = Q(1 - \alpha) - I - L - S$$

where A is the monthly heat gain by the ocean, $Q(1 - \alpha)$ is the solar radiation absorbed by the ocean, α is the surface albedo, I is the net infrared radiation exhange at the surface, L is the latent heat flux due to evaporation and S is the sensible heat exchange with the atmosphere. BUNKER (1976) and HASTENRATH and LAMB (1978) used the same general formula with some difference in the computation of each component of the budget. Their results are not different in their general patterns but differs in their extremes value of about 10 %. For the tropic the most detailed picture are those of HASTENRATH and LAMB (1978) and HASTENRATH (1977). Fig.4 presents the annual mean of the net heat flux at the surface. The cold waters of the equatorial and coastal upwelling regions reduce the latent heat loss by weaker evaporation and therefore trap the heat from the atmosphere. Seasonal variations of the net surface heat flux broadly follow the annual cycle of insolation with largest gain in the respective summer and loss in the winter half-year (Fig.5). The maximum ocean heat gain appears near the equator (between 0 and 5°S) in northern summer and correspond to the most intense upwelling period. The highest values (120 W/m^2) are found in August near 10° W (HASTENRATH and LAMB 1978).

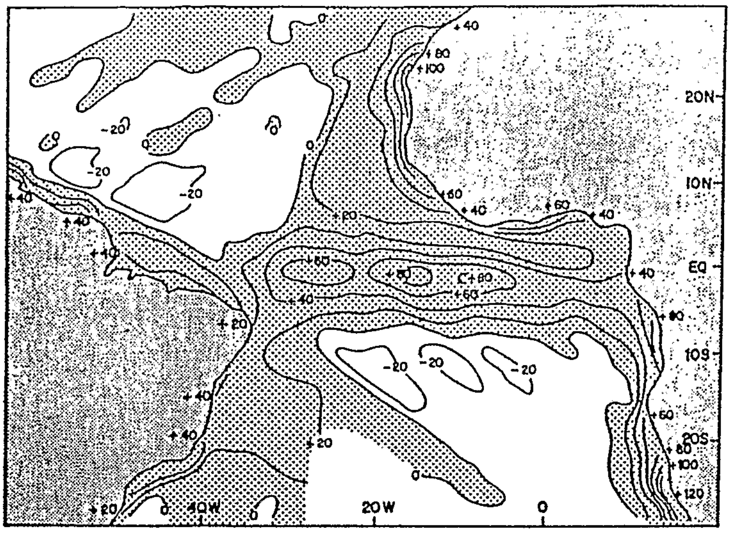


FIG. 4 Mean annual net heat flux at the air-sea interface. Shaded area indicates heat gain by the ocean.units W/m^2. (From HASTENRATH - 1977)

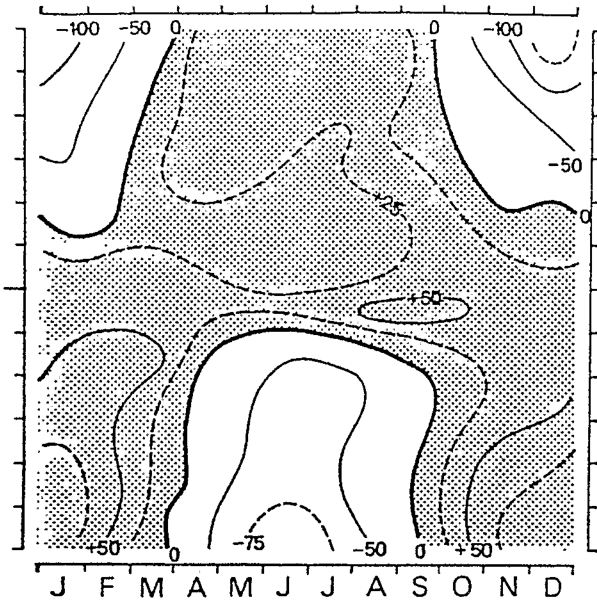


FIG. 5 Seasonal variability of the net heat flux zonaly averaged in the tropical Atlantic from 30°N to 30°S. Shaded area indicates heat gain by the ocean. Units W/m^2 (From HASTENRATH - 1977)

3. THE DYNAMICAL RESPONSE

3.1 Observations

A merged data set including MBT, XBT, and NANSEN - CTD observations, covering the 20°S - 30°N tropical Atlantic ocean until 1978, has been used to study the seasonal changes in the upper ocean thermal structure (MERLE 1983).

Fig.6 shows the topography of the 20°C isotherm in March and August. The difference is clear. In March the 20°C isotherm present a simple structure with a zonal East-West general slope and a slight upwelling along the equator. In August the pattern is different. A zonal downwelling appear at about 3-4°N boardering further North a zonal ridge under the northernmost position of the ITCZ (10°N - 15°N). The maximum seasonal changes are-found in three regions : (i)10° North (ii) equator and West (iii) Eastern equator (Fig.7 upper left panel). Annual signals of thermocline depth variations are not in phase in the three regions (Fig.7 upper right panel). The maximum depth of the thermocline is observed in fall in the West equatorial region and in Spring in the Eastern equatorial region. A double seasonal tilt of the thermocline is resulting of these phase changes : one along the equatorial plan, the other one along a zonal pivot line situated under the mean position of the ITCZ. This double axis of rotation of the thermocline on a seasonal time scale is the result of two distinct responses of the ocean to the wind forcing : (i) a remote equatorial response which mainly affects the Eastern equatorial region. (ii) A local North equatorial response associated to the seasonal migration of the ITCZ that induces a change in the sign and the magnitude of the curl of the wind stress.

The Equatorial response has been first tentatively explained by MOORE et al (1978) that proposed a simple remote forcing mecanism to account for the Guinea Gulf upwelling. According to this hypothesis an increase of the easterly wind in the western

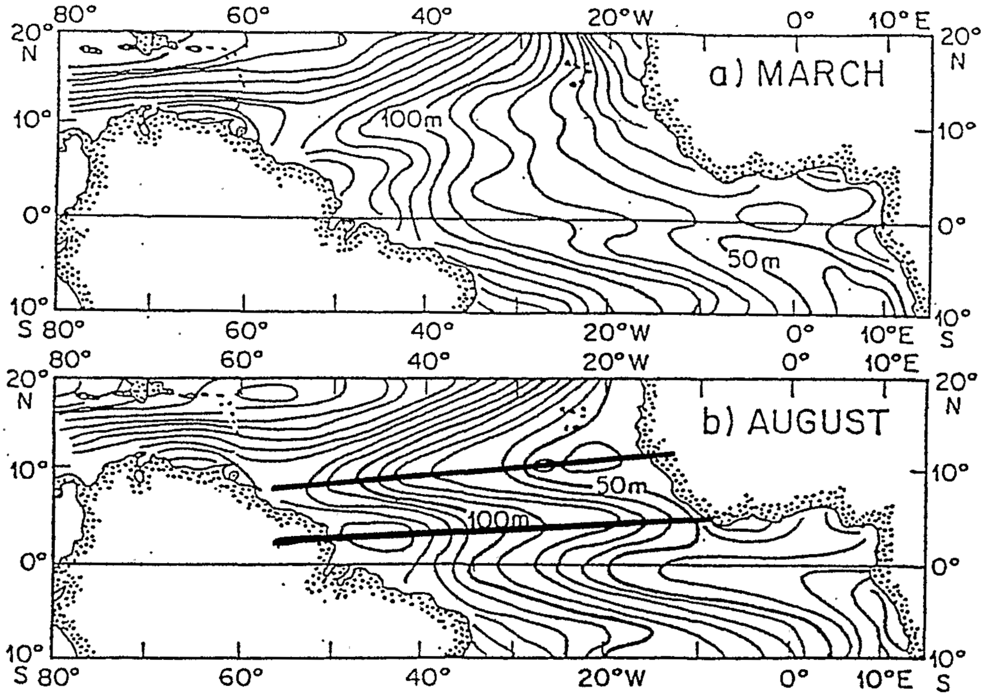


FIG. 6 Topography of the 20°C isotherm in March and August.

equatorial Atlantic excites an internal equatorial KELVIN wave that propagate Eastward and affect the thermal structure in the Guinea Gulf. Latter on, observations of HOUGHTON (1983) in the Guinea Gulf confirmed that a trapped equatorial phenomenon affect the thermocline near 4°W (Fig.8). Furthermore SERVAIN et al (1982) found a hight correlation beetween the wind stress anomalies off Brazilian coast and the SST anomalies in the Guinea Gulf one month later (see chapter 5).

The north equatorial response has been studied by GARZOLI and KATZ (1983). They found that the seasonal reversal of the North Equatorial countercurrent and the associated change of the depth of thermocline beetween 3°N and 10°N (Fig.9) is due to the changes of the wind-stress curl throught the combined mechanisms of the local EKMAN pumping and the divergence of the geostrophic currents.

3.2 Modelling

Important progresses have been achieved during the last decade on modelling the tropical oceans. Historically, theoretical approaches have evolved in several stages. First idealized winds blowing over shallow waters models were used both in the Pacific and in the Atlantic (Mc CREARY 1976 ; CANE and SARACHICK - 1981 ; BUSALACCHI and O'BRIEN - 1980 ; O'BRIEN, ADAMEC and MOORE - 1978) to understand the forced ocean response to abrupt changes of the wind forcing. The equatorial Kelvin waves and their effect on the thermal structure of the eastern equatorial regions have been predicted by these theories before being observed in the Pacific and the Atlantic (KNOX and HALPERN - 1982 ; SEQUAL group - personal communication). Then the same shallow water models were used with realistic climatologic winds to simulate the mean seasonal response of the ocean (BUSALACCHI and O'BRIEN - 1981 ; BUSALACCHI and PICAUT - 1983 ; DU PENHOAT and TREGUIER - 1984). Such modelling results for the Atlantic are presented Fig.7. Both the amplitude and phase of the thermocline depth seasonal variability are correctly

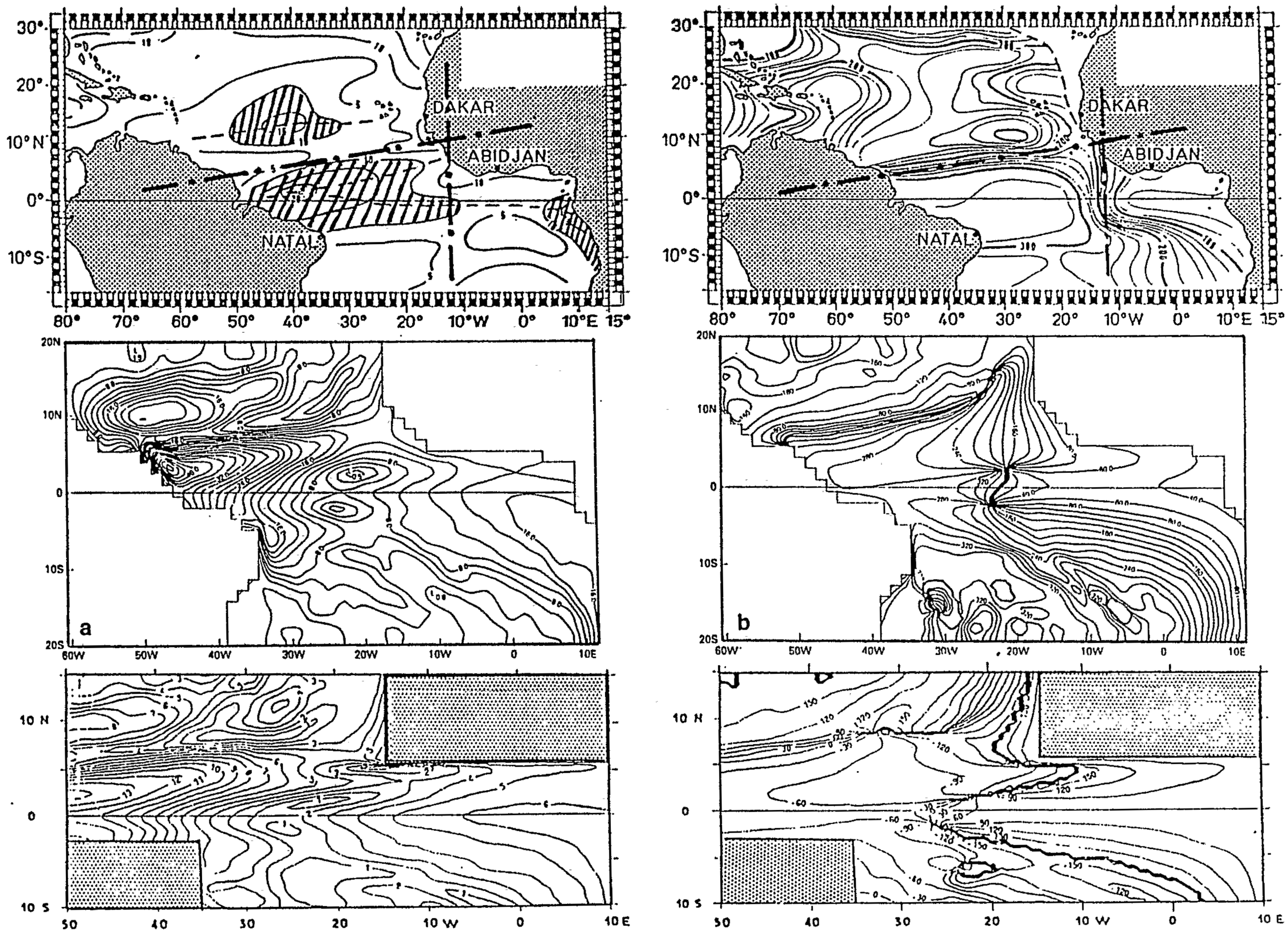


Fig. 7 Upper panels : amplitudes (left) and phase (right) of the seasonal variation of the depth of the thermocline (20°C isotherm)
Intermediate panels : Amplitude and phase of the seasonal variation of the depth of the surface separation in a 2 layers model forced by climatological winds (From BUSALACCHI and PICAUT - 1983).
Lower panels : Amplitude and phase of the surface dynamical topography in a two layers model forced by climatological winds (From DU PENHOAT and TREGUIER - 1984).

simulated. more recently, realistic winds and surface thermodynamic fluxes have been used to force sophisticated three-dimensional models (PHILANDER and PACANOWSKI - 1984) that simulate most of the known features of the seasonal variability of the tropical Atlantic ocean and particularly the surface currents (Fig.10).

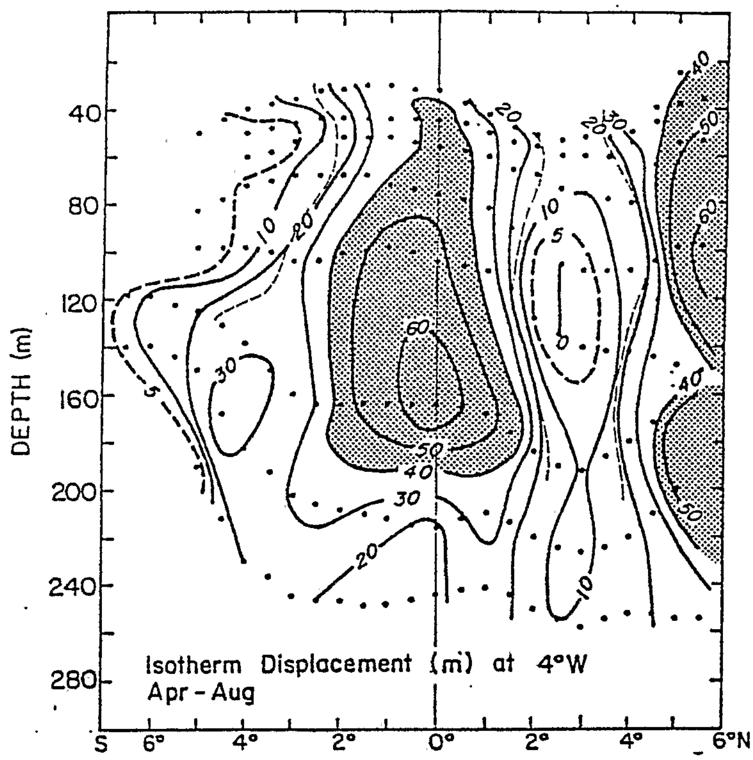


FIG. 8 Isotherm displacement beetween April and August at 4°W in the Guinea Gulf (From HOUGHTON - 1983)

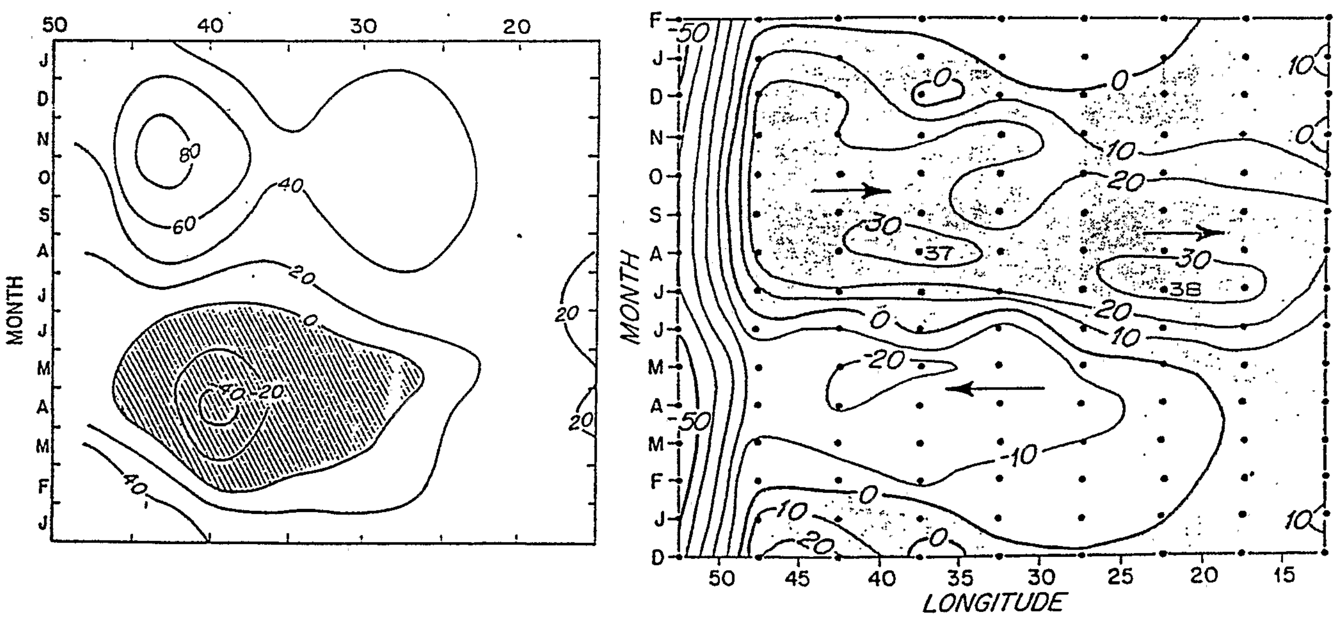


FIG. 9 Seasonal longitudes changes of the depth of the thermocline from 3°N to 7°N (From GARZOLI and KATZ - 1983) compared with the surface currents changes obtained by ship-drift (from RICHARDSON - 1984)

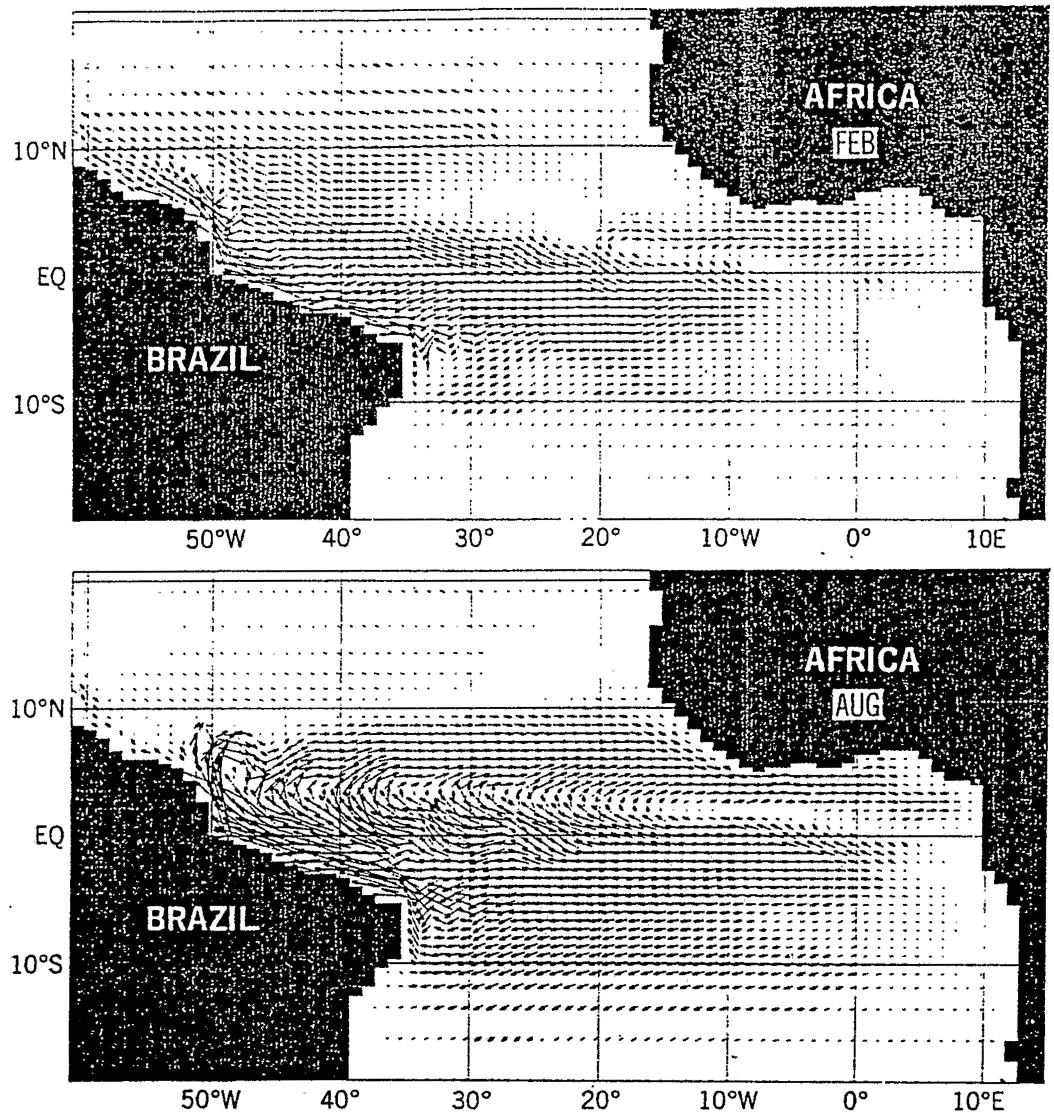


FIG. 10 Surface currents in the tropical Atlantic Ocean, February and August, as simulated with a multi-level numerical model (From PHILANDER and PACANOWSKI 1984)

4. HEAT BUDGET AND HEAT TRANSPORT

Seasonal variations of heat content follow the seasonal variations of the thermocline depth (Fig.11 a and b). The 0-300 meters heat content shows a large maximum of amplitude in the West and at the North of the equator with values close to 6.10^8 J/m^2. An eastern equatorial maximum is also observed with values over 10.10^8 J/m^2 at the African coast close to the equator. The phase map shows also a phase difference between the Eastern and the Western part of the equatorial band with a transition point at about 15°W. Another change phase is observed along a zonal line near 6-8°N which is similar to the one described for the 20°C isotherm depth and situated under the mean position of the ITCZ.

These differences in the seasonal variation of heat content are also marked in the heat storage (So) that is the difference of the time rate of heat content change (ΔHC) minus the net heat flux exchanged at the air - sea interface (F)

$$So = \Delta HC - F$$

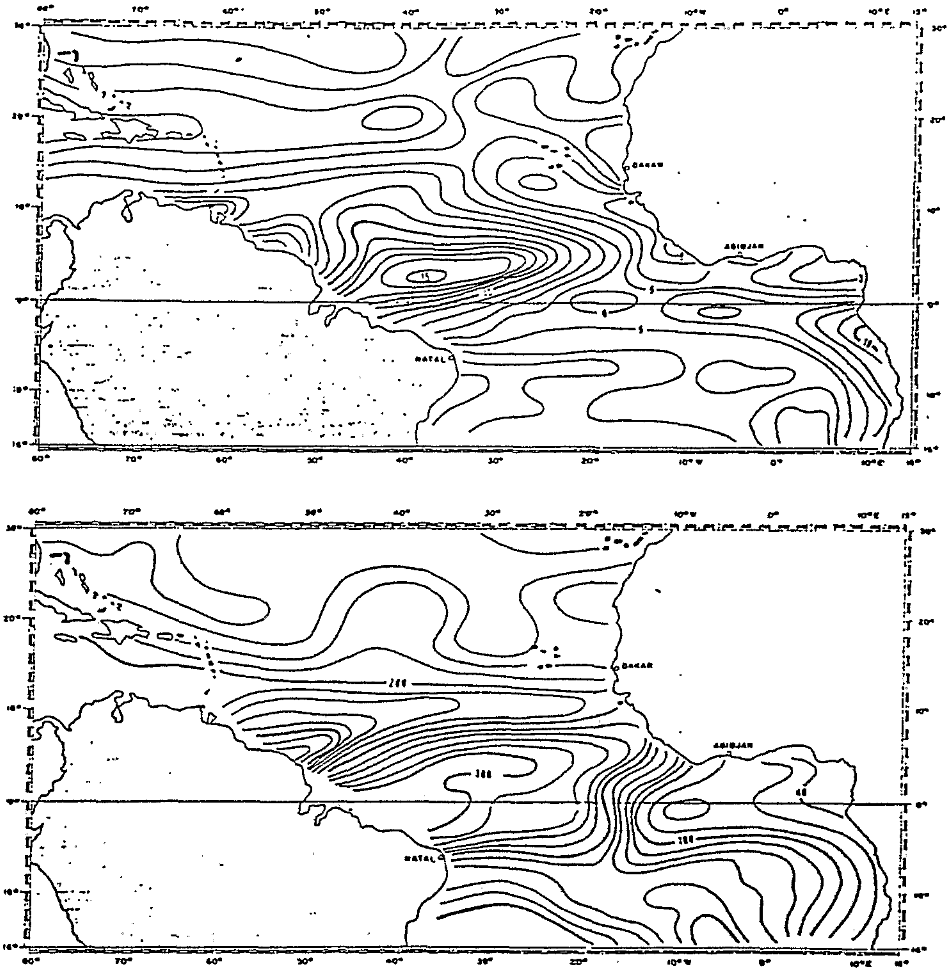
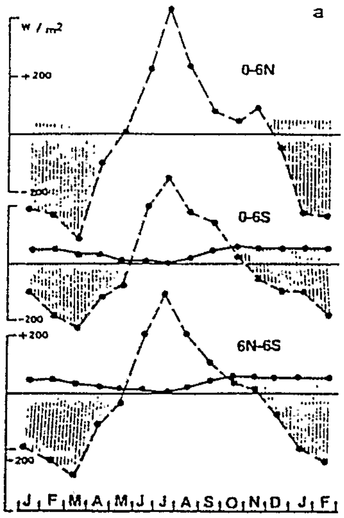


FIG. 11 a and b,amplitude and phase of the seasonal variation of heat content in the layers 0-300 meters.

In the near Equatorial band (6°N - 6°S), the seasonal variation of the time rate of heat content is about 10 times larger than the seasonal variation of the net heat flux through the surface given by HASTENRATH and LAMB (1978) (Figs.12a-12b). In the western region a divergence (or export) of heat is observed from October to May with a maximum in March ; a heat convergence (or import) occurs from June to September in the 6°N-6°S region as a whole. In the eastern region (Gulf of Guinea) heat divergence occurs in two periods : the main period is April-August and the secondary period is November-December. Thus in an annual mean the equatorial Atlantic region is exporting heat but not throughout the year (MERLE 1980 a).

Furthermore these zonal and meridional phase changes in heat storage imply that heat is sloshed East-West but also redistributed merionally. Fig.13a and b present the time rate change of heat content for two months (June and September) that frame the summer equatorial upwelling in the Gulf of Guinea. heat convergence (divergence) appears at a rate of the order of 300 watts per meter square.

LAMB and BUNKER (1982) computing the 0-500 meters heat content and using the net heat surface flux provided by BUNKER (1976) integrated the merional component of the heat transport throughout the Atlantic from 70°N to 20°S. He found an annual mean meridional heat transport of the order of 150.10^{13} W in the tropical Atlantic compatible with previous estimation of EMIG (1967), STOMMEL (1979) and also BRYDEN and HALL (1980) that used direct hydrographic observations along the 27°N parallel.



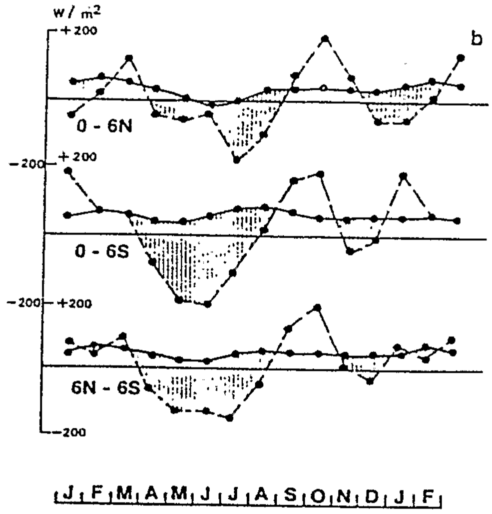


FIG. 12a Annual variation of time rate of change of heat content (interrupted line) and annual variation of net heat oceanic gain (full line) in the western equatorial Atlantic region. Units are W/m^2. Dashed areas represent divergence of heat.

FIG. 12b As in Fig.10 a, except for the eastern region (From MERLE - 1980 a)

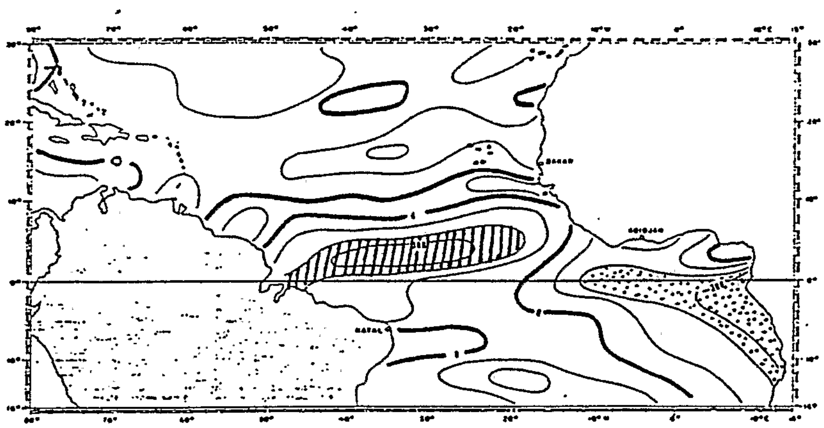


FIG. 13a Rate of change of heat content from 0 to 300 Meters in June (in W/m^2)

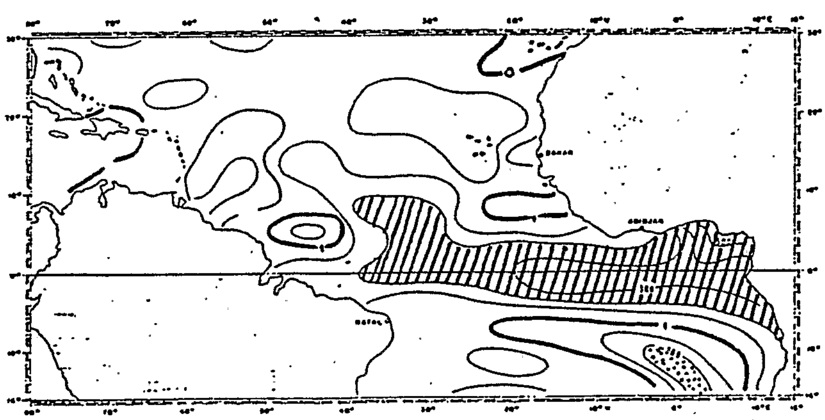


FIG. 13b As in Fig.1 a for September.

Furthermore LAMB and BUNKER presented a seasonal variation of the meridionnal heat transport indicating a surprising reverse of the heat transport flowing southward in fall all along the Atlantic from 70°N to 20°S (Fig.14).

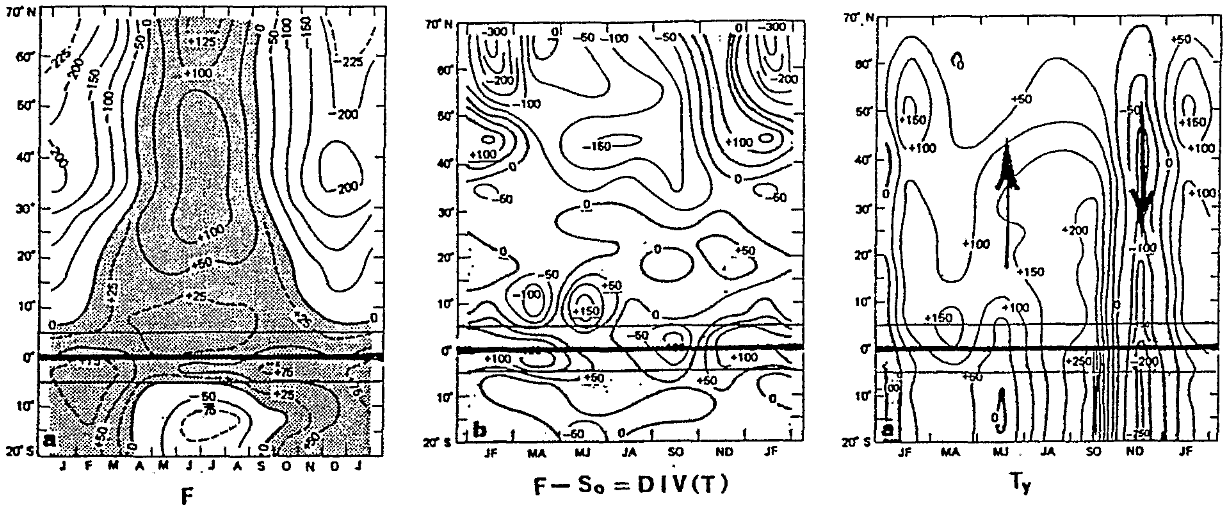


FIG. 14 Seasonal variation of zonal integration in the Atlantic of :
- Net heat flux at the air-sea interface (F)
- Divergence of heat transport (T). So being the time rate change of heat content in layer 0-500 meters.
- Meridionnal component of heat transport in the layer 0-500 meters (Ty). (From LAMB - 1983).

5. THE SEA SURFACE TEMPERATURE RESPONSE

5.1 Observations of sea Surface temperature variability.

The sloping of the thermocline which rises from west to east is partly responsible for the coolest waters that are found in the eastern tropical Atlantic. The spatial structure of the amplitude of seasonal SST variability ressembles the mean state (Fig.15 - SERVAIN et al. - 1985). Regions with the greatest seasonal variability coincide with regions of low SST. The shallow thermocline in the eastern tropical Atlantic together with vertical mixing induces maximum seasonal SST variability (σ = 1.4-3.4°C) in the seasonal upwelling zones along the coasts of Mauritania and Senegal (NW Africa), the northern and southern coasts of the Gulf of guinea, and along the equator near 10°W. The location of the minimum SST variability (σ = 0.4°C) is coincident with the thermal equator and the mean Intertropical Convergence Zone (ITCZ).

The interannual variability of SST, as depicted in Fig.15 is maximum (σ = 0.6 -1.0°C) in the coastal and equatorial upwelling regions but minimum offshore the northern coast of Brazil.

5.2 A remote forcing mecanism to explain the SST variability in the Gulf of Guinea ?

MOORE et al (1978) suggested that remote forcing by winds in the western Atlantic may provide an explanation of the SST variability in the Gulf of Guinea.

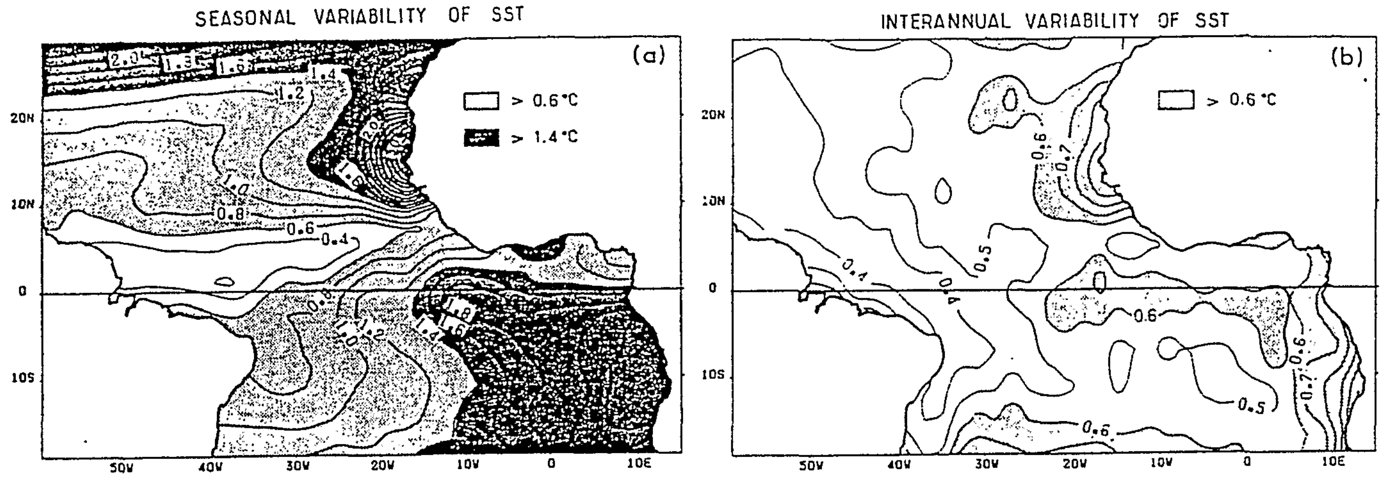


FIG. 15 Seasonal and Interannual variability of SST (From SERVAIN et al - 1985).

An analysis of SST and surface winds in selected areas of the tropical Atlantic by SERVAIN et al (1982) indicated 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 (Fig.16). A negative (positive) anomaly of the wind stress near the north Brazilian coast is follow by a positive (negative) anomaly in the Gulf of Guinea about one month later. Furthermore, the correlation between the local wind stress anomaly and SST anomaly of SST in the Gulf of Guinea is considerably smaller. These results indicate that remote forcing in the western equatorial Atlantic ocean is affecting the eastern equatorial Atlantic sea surface temperature.

Some particular years like 1967 or 1968 impulsive wind stress fluctuations in the Western equatorial Atlantic could excite equatorially trapped waves that would induce SST forced response in the Gulf of Guinea. The warming in the Gulf of Guinea of more than 2°C in July-August 1968 was preceded by abnormal wind stress forcing off the northern Brazil coast one month before (Fig. 17). These warm events related to abnormal wind similar to the "EL NINO" phenomenon in the Eastern Equatorial Pacific have been called "Atlantic EL NINO" (HISARD - 1980 ; MERLE - 1980b). In 1984 a warm event of unusual amplitude affected the Eastern tropical Atlantic during the relaxation phase of the 1982-83 Pacific "EL NINO" . Some possible connexions beetween the two events could be searched through the global long term perturbation of the trade winds during the period 1982-1984.

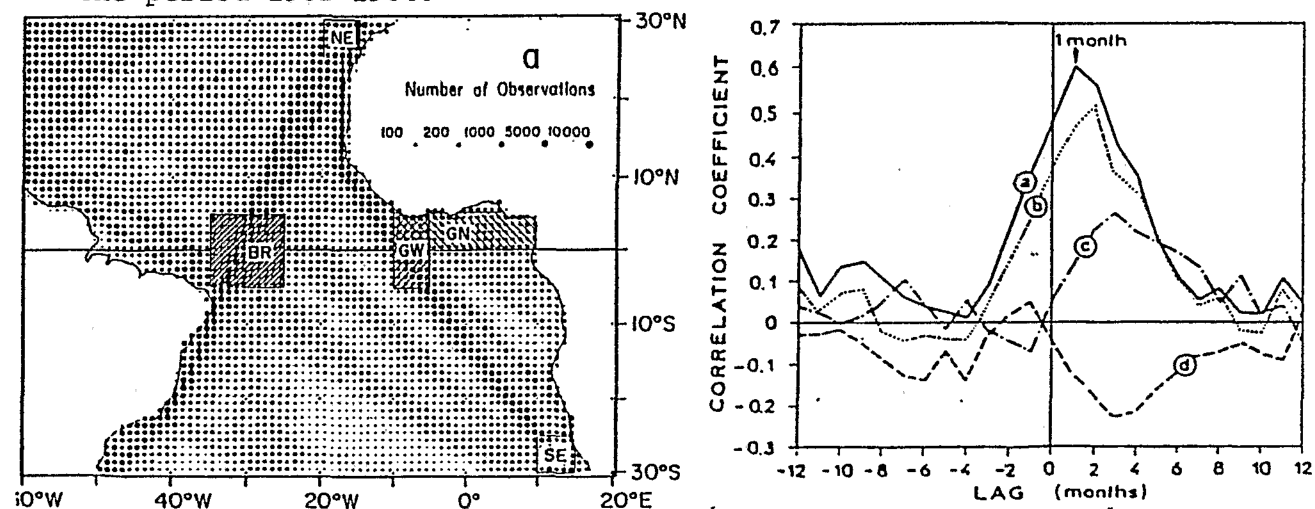


FIG. 16 Correlation with time lag. Curve a: τ^x-BR with SST-GW. Curve b: τ^x-BR with SST-GN. Curve c: τ^x-GW with SST-GW. Curve d: τ^y-GW with SST-GW. (From SERVAIN et al. - 1983).

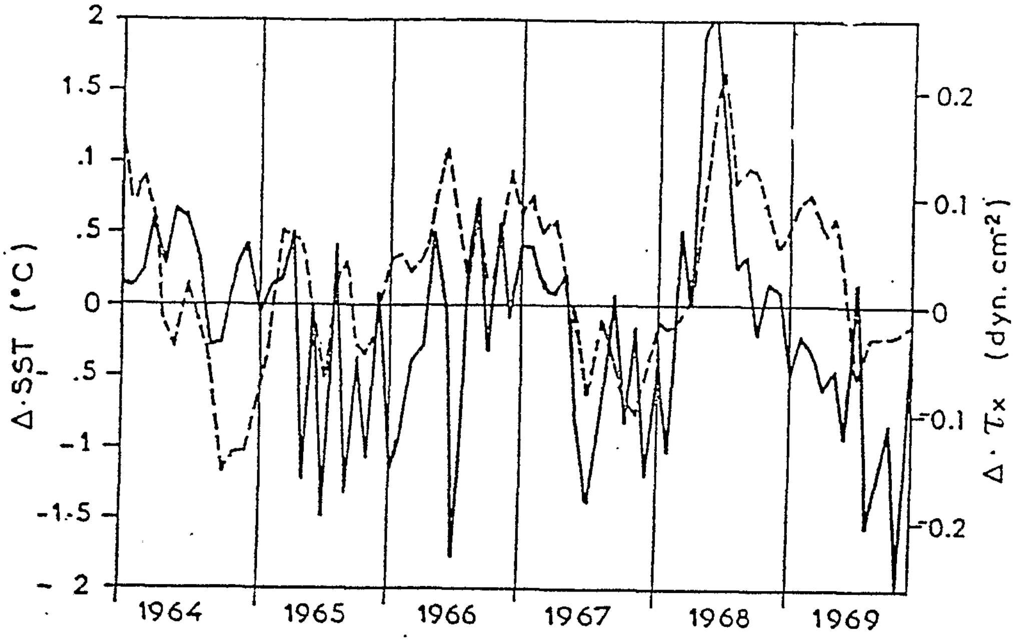


FIG. 17 . Comparison of monthly anomalies of SST in GW area(dotted line) and τ^x in BR area (solid line). From SERVAIN et al (1983)

6. CONCLUSIONS

Three times smaller than the tropical Pacific ocean the tropical Atlantic is a remarquable equatorial "laboratory" for studying the basic physics of the coupled ocean-atmosphere system. A small ocean basin has a shorter adjustment time than a large basin, because this adjustment time is proportional to the time it takes planetary waves to propagate across the basin. Thus, instead of the Pacific, the tropical Atlantic has almost no memory. The zonal slope of the thermocline along the equator, the density field and the intensity of the equatorial currents vary almost in phase with seasonal changes in the wind field. Even outside of the equatorial band a near equilibrium response of the tropical ocean is observed. The seasonal reversal of the North Equatorial countercurrent is in phase with the seasonal excursion of the ITCZ and the associeted wind changes (GARZOLI and KATZ - 1983). Because of its smaller size the tropical Atlantic shows seasonal changes that are dynamically similar to interannual changes in the Pacific and make it easier to observe.

The available data and the models simulations have provided remarkable coherents pictures of the seasonal response of the tropical Atlantic ocean to the wind forcing (Fig.7). The double oscillation of the thermocline on an East-West and North-South rotation imply seasonal redistributions of heat zonally and merionally that are an order of magnitude larger and out of phase with the heat flux across the surface (MERLE - 1980a). This means that the dynamical interior response of the tropical ocean is the most important for understanding and predicting the ocean's role in the overall climate system. Zonal equatorial heat transport may account for "EL NINO" like phenomenons on seasonal time scale or interannual time scale (MERLE - 1980b ; HISARD - 1980) like in the Pacific.

If at the seasonal and interannual time scales the Atlantic ocean has a weaker influence on the world climate than the Pacific and Indian oceans, it may have a dominant influence on time scales of decades or centuries. The Atlantic ocean is a meridional ocean functioning as a heat pipe, deriving heat to the North Atlantic from the world ocean heat circulation system (Fig.1). The narrow tropical Atlantic concentrate a huge cross equatorial merional heat flux and may offer the best place in the world to observe and understand the oceanic heat transport and its influence on climate.

BIBLIOGRAPHY

BRYDEN H.L. and M.M. HALL, 1980 : Heat transport by curents across 25°N latitude in the Atlantic Ocean. Science, 207, 884-886.

BUDYKO M. I., 1956 : The heat-balance of the Earth's surface. English translation by US Weather Bureau, Washington D.C.

BUNKER A.F., 1976 : Computations of surface energy flux and annual air-sea interaction cycles of the North Atlantic Ocean. Mon. Wea. Re., 104, 1122-1140.

BUSALACCHI A. and J.J. O'BRIEN, 1981 : Interannual variability of the equatorial Pacific in the 1960's. J. Geophys. Res., vol.86, 10901-10907

BUSALACCHI A. and J. PICAUT, 1983 : Seasonal variability from a model of the tropical Atlantic Ocean. J. Phys. Oceanogr., 13, 1564-1588.

CANE M.A. and E.S. SARACHIK, 1981 : The response of a linear baroclinic equatorial ocean to periodic forcing, J. Mar. Res., 39, 651-693.

EMIG M., 1967 : Heat-transport by ocean currents, J. Geophys. Res., 72, 10, 2519-2529.

GARZOLI S. L. and E. J. KATZ, 1983 : The forced annual reversal of the Atlantic North Equatorial Countercurrent, J. Phys. Oceanogr., 13, 2082-2090.

HASTENRATH S., 1977 : Relative role of atmosphere and ocean in the global heat budget : tropical Atlantic and eastern Pacific. Quart. J. Roy. Meteor. Soc., 103, 519-526

HASTENRATH S. and P.J. LAMB, 1977 : Climatic atlas of the tropical Atlantic and eastern Pacific oceans. The University Wisconsin Press, 112 p.

HASTENRATH S. and P.J. LAMB, 1978 : Heat budget atlas of the Tropical Atlantic and eastern Pacific ocean. The University Wisconsin Press, 140 p.

HELLERMAN S., 1980 : Charts of the variability of the wind stress over the tropical Atlantic. GATE Sup. II. Deep sea Res., 26, 63-75.

HISARD P., 1980 : Observation de réponses de type "EL NINO" dans l'Atlantique tropical oriental-Golfe de Guinée. Ocean. Acta, 3, 69-78.

HOREL J.D. and J.M. WALLACE, 1981 : Planetary-scale atmospheric phenomena associated with the Southern Oscillation. Mon. Wea. Rev;, 109, 813-829.

HOUGHTON R.W., 1983 : The subsurface thermal structure in the Golf of Guinea, J. Phys. Oceanogr. , 13, 2070-2081.

KNOX R. A. and D. HALPERN, 1982 : Long range Kelvin wave propagation of transport variations in Pacific ocean equatorial currents, J. Mar. Res., 40, Supp., 329-339.

LAMB P.J. and A.F. BUNKER, 1982 : The annual march of the heat budget of the North and tropical Atlantic ocean. J. Phys. Oceanogr., 12, 12, 1388-1410.

MC CREARY J. P., 1976 : Eastern tropical ocean response to changing wind systems : With application to El Nino, Journal of Physical Oceanography, 6, 632-645.

MERLE J., 1980 a : Seasonal variation of heat-storage in the tropical Atlantic ocean, Oceanologica Acta, 3, 209-220.

MERLE J., 1980 b : Variabilité thermique annuelle et interannuelle de l'ocean Atlantique équatorial Est.L'hypothèse d'un "El Nino" Atlantique, Oceanologica Acta, 3, 209-220.

MERLE J., 1983 : Seasonal variability of subsurface thermal structure in the tropical Atlantic ocean, Proc. 14 th LIEGE Colloquium ocean Hydrodynamics, Elsevier, 31-49.

MOORE D., PH. HISARD, J. MC REARY, J. MERLE, J. O'BRIEN, J. PICAUT, J. M. VERSTRAETE and C. WUNSCH, 1978 : Equatorial adjustment in the Eastern Atlantic. Geophysical Research Letters, 5, 637-640.

MOURA A.D. and J. SHUKLA, 1981 : On the dynamics of droughts in northeast Brazil : Observations, theory and numerical experiments with a General Circulation Model. J. Atmos. Sci., 38, (to appear).

O'BRIEN J. J., D. ADAMEC, D. MOORE, 1978 : A simple model of equatorial upwelling in the Gulf of Guinea, Geophys. Res. Lett, 5, 8, 633-636.

OORT A. H. and T.H. VONDER HAAR, 1976 :On the observed annual cycle in the ocean-atmosphere heat balance over the northern hemisphere. J. Phys. Oceanogr., 6 (6), 781-800.

PAN Y. H. and A. H. OORT., 1983 : Global climate variations connected with Sea Surface Temperature anomalies in the eastern equatorial Pacific ocean for the 1958-73 period. Mon. Wea. Rev., 111, 1244-1258.

DU PENHOAT Y. and A.M. TREGUIER, 1984 : Sea Surface Dynamic Height Topography and the North Equatorial Countercurrent as Infered from a Linear Model. Geophys. Res. Letter, vol.11, 799-801.

PHILANDER S. G. H. and R. C. PACANOWSKI, 1984 : Simulation of the seasonal cycle of the tropical Atlantic ocean. To appear in J. Phys. Oceanogr.

SERVAIN J., J. PICAUT and J. MERLE, 1982 : Evidence of remote forcing in the equatorial Atlantic Ocean. J. Phys. Oceanogr., 12, 457-463.

SERVAIN J., J. PICAUT and A.J. BUSALACCHI, 1985 : Interannual and seasonal variability of the tropical Atlantic ocean depicted by 16 years of Sea Surface Temperature and wind-stress. To appear in J. Phys. Oceanography.

STOMMEL H., 1979 : Oceanic warming of Western Europe. Proc. Nat. Acad. Sci., U.S.A., 76, 2518-21