MAJOR DYNAMICS AFFECTING THE EASTERN TROPICAL ATLANTIC AND PACIFIC OCEANS

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ABSTRACT

Despite their difference in size, the Pacific and Atlantic oceans have eastern tropical regions that are similar, especially in their dominant dynamics. Both regions are primarily influenced by remote wind stress effects through equatorial wave dynamics. However, the eastern equatorial Pacific is dominated by an interannual time scale (i.e., El Niño), whereas the eastern tropical Atlantic is dominated by a seasonal time scale. In both of these oceans, the Equatorial Undercurrent impinging upon the eastern boundary is associated with an important westward zonal pressure gradient—a gradient that is not in equilibrium with the local wind stress. Two other subsurface currents—the northern and southern equatorial undercurrents—play a role in the formation of the Guinea, Angola, and Costa Rica domes. All these common dynamic processes are important to the nutrient enrichment of both eastern boundary oceans.

RESUMEN

A pesar de sus diferencias en magnitud, los océanos Pacífico y Atlántico poseen regiones tropicales orientales semejantes, especialmente en lo que se refiere a sus características dinámicas más importantes. Ambas regiones están influenciadas principalmente por los efectos remotos de la perturbación del viento a través de la dinámica de olas ecuatoriales. Sin embargo, el Pacífico ecuatorial oriental está dominado por una escala temporal interanual (i.e., El Niño), mientras que el Atlántico tropical oriental está dominado por una escala temporal estacional. En ambos océanos la Corriente Ecuatorial Subsuperficial que avanza sobre el margen oriental está asociada con un importante gradiente zonal occidental de presión, gradiente que no está en equilibrio con la perturbación local del viento. Otras dos corrientes subsuperficiales, las ecuatoriales norte y sur, influyen en la formación de los domos de Guinea, Costa Rica, y Angola. Todos estos procesos dinámicos comunes a ambos sistemas son importantes para el aporte de nutrientes a ambos sectores oceánicos orientales.

INTRODUCTION

The eastern parts of the tropical Atlantic and Pacific oceans are among the most biologically active regions of the world oceans. This is because in these regions the thermocline is, on the average, very close to the surface and affected by strong vertical motions, which are fundamentally important in fertilizing the euphotic zones. The dramatic El Niño, now believed to be a large-scale ocean-atmospheric event, has its most spectacular effects in the eastern tropical Pacific. A similar phenomenon occurs in the eastern tropical Atlantic (hereafter referred to as the Gulf of Guinea) and despite its smaller oceanic signature can have devastating effects on the fisheries and climate of the surrounding countries (Hisard and Piton 1981).

Predicting such phenomena is the goal of numerous research groups, but this goal would be imperfectly reached without, first, a clear understanding of the mechanisms involved. A rational (cause-effect) approach is almost impossible in biological-ocean studies because of the complexity of the biological processes themselves and of their linkages with the physical environment. Medium- and high-latitude oceans are characterized by a predominance of thermodynamic effects in the surface layer and are dominated by the presence of highly energetic mesoscale vortices throughout the water column. Because of the high number of such vortices and the impossibility of linking them in a simple manner to any direct origin, an empirical approach appears more suitable at such latitudes. In the equatorial oceans the local thermodynamic effects are small compared to the dynamic ones, and these oceans have the outstanding property of responding clearly and coherently to the wind fluctuations. A rational approach is therefore much better adapted there.

In this paper I tentatively explain the major dynamics that affect the eastern tropical Atlantic and Pacific oceans. I have tried to concentrate on the cause-effect relationship, which necessarily involves some notion of the balance of forces involved. One fundamental aspect of the equatorial oceans—their ability to radiate energy quickly from one part of the basin to another—is explained through simple schemes. I hope to improve the understanding of the remote forcing

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theory, which has been applied with some success to the interannual El Niño in the Pacific and to seasonal upwelling in the Gulf of Guinea. The reversal of the zonal pressure gradient in the eastern equatorial oceans is also presented. It must be pointed out that although this reversal is not well known, it is indeed a general feature of these eastern oceans. Finally, the mechanisms involved in the formation of the highly productive thermal dome areas are discussed.

RESPONSE TO REMOTE FORCING

Most of the tropical Atlantic and Pacific oceans are under the influence of the northeast and southeast trades. In the eastern part of both oceans, the winds are predominantly from the south, with an eastward veering toward the coasts of the Gulf of Guinea and Central America. In the Gulf of Guinea the strong coastal and equatorial seasonal upwelling (Figure 1) does not seem related to the local wind via Ekman divergence (Houghton 1976; Berrit 1976; Bakun 1978; Servain et al. 1982), nor to heat exchange with the atmosphere (Merle 1980). In the same way, the interannual El Niño appearance of abnormally warm water along the coast of Ecuador and Peru is not clearly related to the weakening of the coastal southerly wind (Wyrtki 1975). Along the western coast of North America, sea level (representative of thermal variation over the water column) and local alongshore wind stress anomalies are poorly correlated south of San Francisco, but much better correlated farther north, where predominance of a local wind-driven response is suggested.

Bjerknes (1966, 1969) was the first to suggest that the El Niño phenomenon could be due to a large-scale ocean-atmosphere interaction in which the relaxation of the trades along the equator has a dominant role. Wyrtki (1975) emphasized the dynamical aspect of such an approach. The usually strong westward winds along the equator in the central and western part of the basin accumulate water in the west and build up an east-west sea-level slope (see next section for further explanation). A rapid relaxation of the trades in the central equatorial Pacific destroys the wind stress/sea-level slope equilibrium and excites an internal downwelling equatorial Kelvin wave. This wave propagates eastward and depresses the thermocline all the way to the eastern boundary. This remote forcing mechanism and its effect on the boundary has been studied analytically and numerically by Moore (1968), McCreary (1976), Hulburt et al. (1976), and Busalacchi and O'Brien (1981).

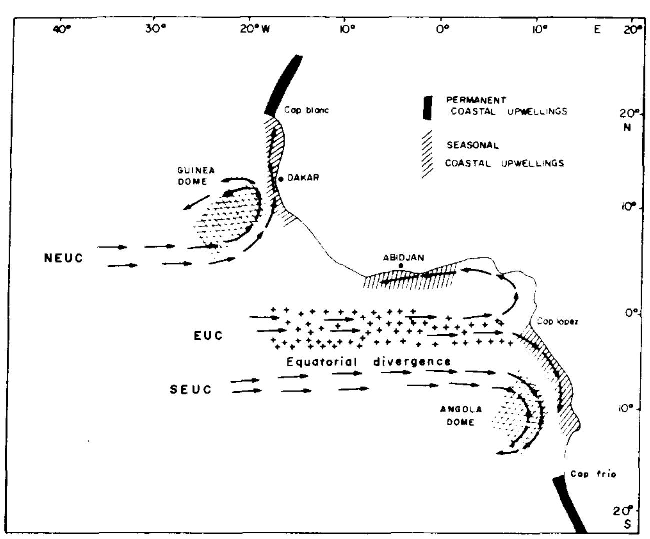


Figure 1. The productive areas in the eastern tropical Atlantic Ocean and the three branches of the Equatorial Undercurrent system (from Voituriez and Herbland 1982).

In the Pacific Ocean, the downwelling El Niño event occurs on an interannual time scale. In the Gulf of Guinea the strong upwelling signal occurs seasonally. In neither case does local forcing seem to be the main explanation. In view of such similarities, Moore et al. (1978) have suggested a similar mechanism for upwelling in the Gulf of Guinea, with the seasonal intensification of the trades in the western Atlantic at the origin of an equatorial upwelling Kelvin wave. Numerical models have also been used to detail this seasonal remote forcing mechanism (Adamec and O'Brien 1978; Busalacchi and Picaut 1983; McCreary et al. 1984).

In order to better explain this remarkable property of far-field forcing, which is characteristic of the equatorial oceans, I have extracted some figures from O'Brien et al. (1980). They represent the simplest numerical solution that clearly illustrates this Kelvin wave scenario. A reduced-gravity linear model is forced by an easterly patch of zonal wind stress acting on the western part of an idealized equatorial basin, which is limited on the east and west by solid boundaries and on the north and south by open boundaries. This model simulates the generation of an upwelling Kelvin wave in the Pacific, but has also been adapted to the Atlantic by O'Brien et al. (1978). A simple reversal of the wind direction could induce a downwelling Kelvin wave and therefore simulate, crudely, an El Niño event. Figure 2 represents the depth displacement of the pycnocline separating the two layers of the model. Initially the ocean is at rest. Within a few weeks after the easterly winds, west of 170°E, are switched on, a westward, wind-driven surface flow accumulates water against the western boundary. At the same time the Ekman divergence along the equator induces upwelling just west of 170°E. This patch of upwelling then propagates eastward along the equato-

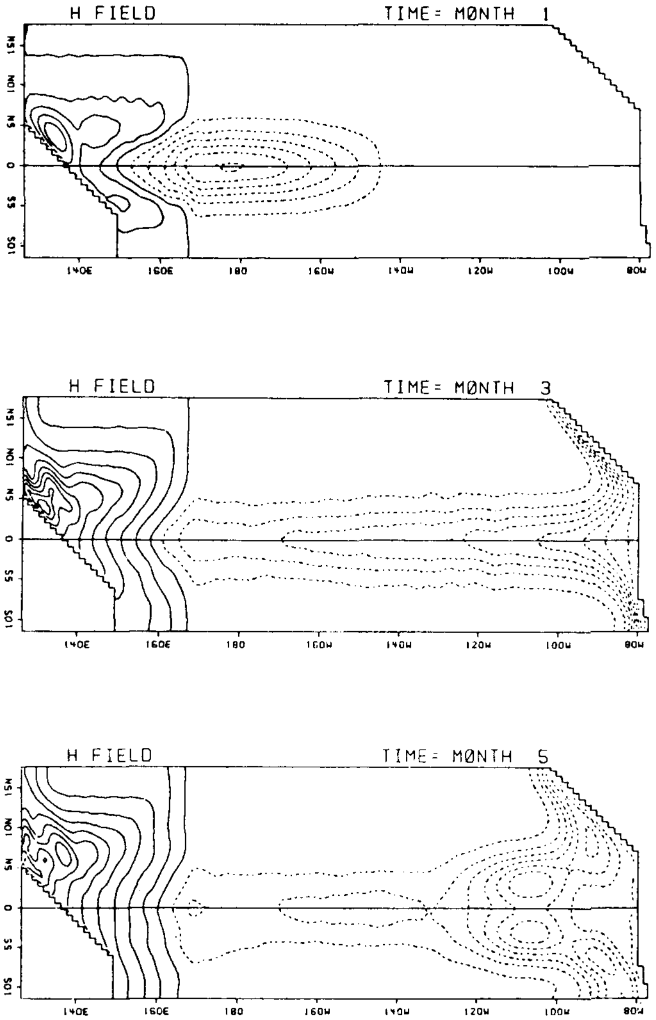


Figure 2. Interface displacement depicting the progression of an equatorially trapped Kelvin wave and resulting coastal Kelvin waves and Rossby waves. Contouring interval is 5 m, with dashed contours indicating upwelling (from O'Brien et al. 1980).

rial wave guide as a freely propagating equatorial Kelvin wave. When this upwelling disturbance reaches the eastern boundary, it reflects as packets of Rossby waves propagating westward symmetrically about the equator, and as two coastal Kelvin waves propagating poleward.

Such a scenario, where the westward equatorial wind (cause) could be easily related, a few months later, to an eastern boundary upwelling (effect) thousands of kilometers away, is one fundamental aspect of the equatorial region. At midlatitudes there is no wave mechanism to bring energy over long distances from west to east. Only Rossby waves exist, but they propagate westward so slowly, compared to their equatorial counterpart, that it is impossible to simply relate wind forcing with its effects in temperate oceans.

Reduced-gravity models assume that the ocean consists of a thin surface layer (50-200 m) of weak density overlying a deep lower layer (2,000-5,000 m) of greater density. Even if such an approximation is quite relevant in equatorial oceans, these models restrict the energy to traveling only horizontally. Figure 3, adapted from McCreary (1981), is a theoretical representation of equatorial waves in a continuously stratified model. The energy that appears east of the forcing area is the result of the radiation of many equatorial Kelvin waves. They superpose to form a beam of energy that propagates eastward but also downward. When this Kelvin beam arrives at the eastern boundary, it reflects as a collection of beams of Rossby waves that propagate energy westward and downward. This reflection may be completed, depending on the forcing frequency, by two coastal Kelvin beams, which propagate energy poleward and down-

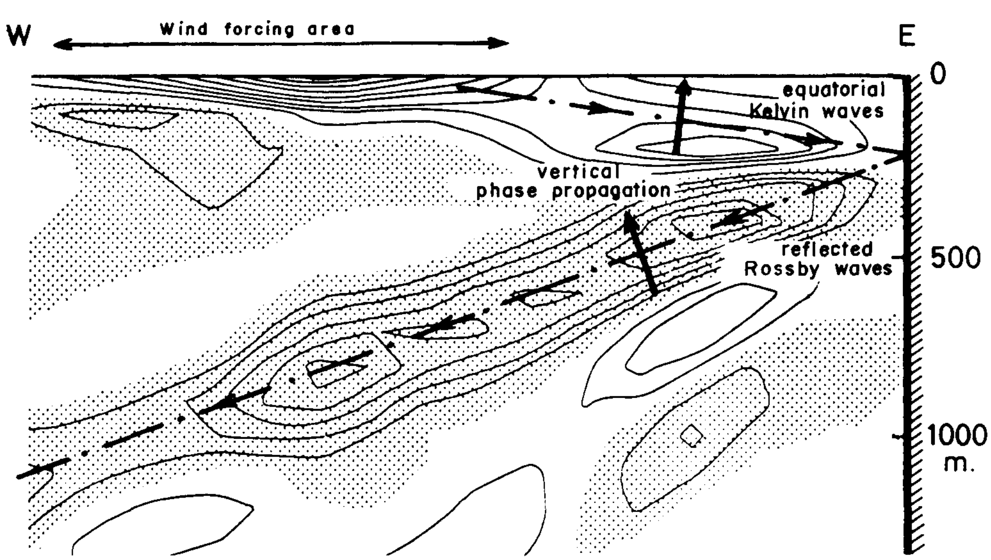


Figure 3. Vertical section of zonal velocity along the equator (contour interval = 5 cm s^{-1}, westward shaded) in a linear model forced by a zonal wind oscillating at the annual period. The response is shown at time of maximum eastward wind in the annual cycle. (from McCreary 1981).

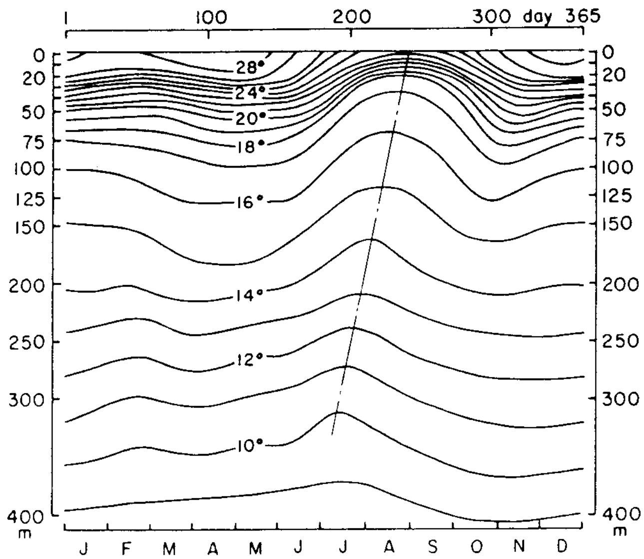


Figure 4. Mean seasonal cycle of isotherm depths deduced from 217 hydrological stations near Abidjan in the Gulf of Guinea. Straight line indicates vertical propagation of the seasonal upwelling signal (from Picaut 1983).

ward. As in the preceding example, the propagation of energy is expressed as an uplifting or a deepening of the isotherms, depending on whether the remote wind direction is westward or eastward. In addition, the downward propagation of energy implies that the energy penetrates in the deep layers faster than in the surface layers; therefore the thermal structure is displaced first in the deep layers, and then progressively toward the surface layers. This is clearly illustrated in Figure 4, where the observed seasonal upwelling signal, near the northern coast of the Gulf of Guinea, exhibits a vertical phase propagation from 300 m to the surface (Picaut 1983).

Horizontal and vertical phase propagation does not imply a mass transport from the forcing area to the eastern boundary, nor from the deep ocean to the surface. The reader has to keep in mind that many of the large-scale phenomena of the equatorial oceans are now explained in terms of wave dynamics, and the analogy to the well-known phenomena of tide and swell could improve understanding of this wave concept. The notion of phase propagation and of associated horizontal and vertical currents remains the same in both. Only the space-time scales are totally different; in these wave dynamics, the frequency is very low (periods from a few weeks to a few years) and the horizontal (vertical) wavelength is of the order of the ocean basin width (depth).

It is fascinating to note that most of the theoretical and numerical studies on El Niño and on the Gulf of Guinea upwelling were made at a time when there was no *in situ* wave evidence. Recently, Knox and Halpern (1982) have shown clear evidence in the Pacific of a pronounced pulse propagating eastward at 2.7 m/s along the equator from 150°W to 90°W; this pulse was apparently forced by an abrupt reversal of trades near 170°E. Eriksen et al. (1983) discussed propagation of such events in detail, and Lukas et al. (1984) presented direct evidence of equatorial wave propagation during the 1982-83 El Niño. In the Atlantic, Katz (1984) observed pulses of displacement of the thermocline that propagate eastward all along the equator. Enfield (1980) discussed evidence of poleward wavelike propagation of sea-level fluctuations all along the eastern boundary of the Pacific, and Picaut (1983) found poleward propagation of the seasonal upwelling along the coast of the Gulf of Guinea. In a study of 29 years of monthly sea level from Mexico to Alaska, Chelton and Davis (1982) showed that the interannual variability of all these records is closely related to El Niño phenomena in the eastern tropical Pacific and propagates poleward at a mean phase speed of 40 cm/s (Figure 5). But the interannual variability along the western coast of North America could not be explained solely by this wave-dynamical theory of El Niño. Large-scale and local atmospheric forcing are also important, especially (as previously noted) north of San Francisco. Lukas and Firing (in press) present evidence of an annual Rossby wave in the central equatorial Pacific; the wave propagates vertically from 900 m to the near-surface pycnocline. In the eastern equatorial Pacific, Lukas (1981) showed some evidence of remotely forced waves by discovering vertical phase propagation, from 500 m to 100 m, at semiannual frequency. Remote forcing of the seasonal upwelling signal along the Gulf of Guinea by zonal equatorial wind stress outside the gulf has been supported by the evidence of vertical (Figure 4) and horizontal phase propagation (Picaut 1983) and the SST-wind stress correlation of Servain et al. (1982). This remote forcing mechanism has been carefully detailed with the numerical models of Busalacchi and Picaut (1983) and McCreary et al. (1984).

WESTWARD ZONAL PRESSURE GRADIENT

Except in their eastern parts, the equatorial Atlantic and Pacific oceans are forced by the strong easterly trades. The westward wind stress appears to be roughly equilibrated by an eastward pressure gradient (Sverdrup balance), which appears as a descending slope of the sea surface from west to east (Figure 6a). This pressure gradient extends down into the subsurface layer and induces a descending slope of the isotherms (Figure 6b) from east to west (Lemasson and Piton 1968; Neumann et al. 1975). Just below the mixed layer this eastward pressure gradient is no longer in

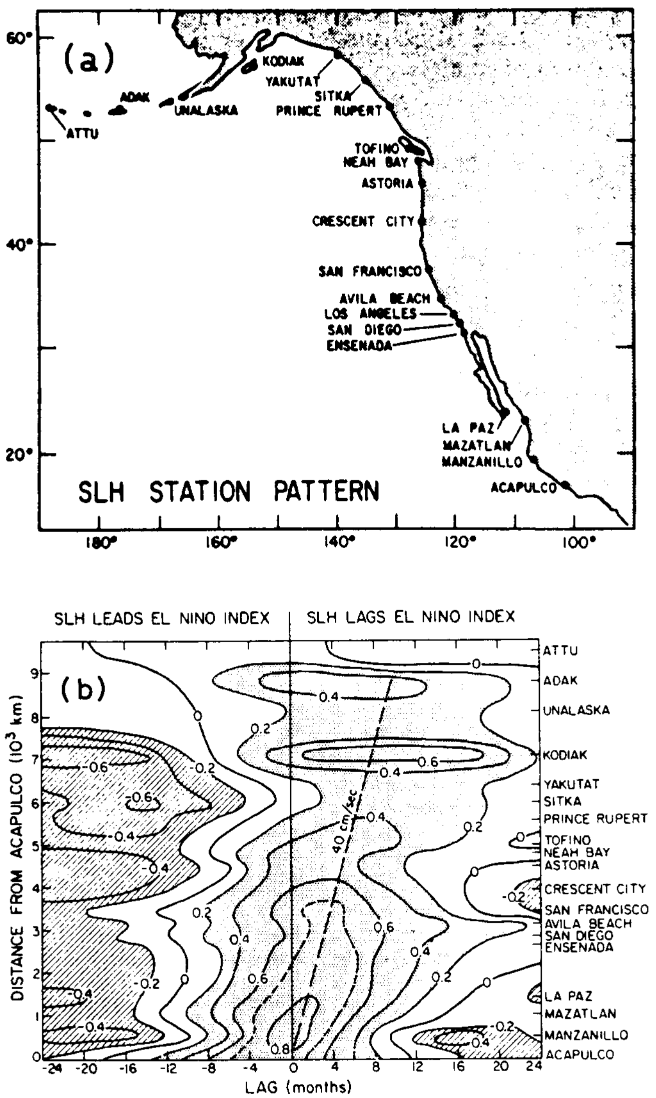


Figure 5. *a.* Location of the 50 sea-level height (SLH) tide-gage stations analyzed.
b. Contour plot of the correlation between low-frequency SLH at each of the 20 tide-gage stations and low-frequency eastern tropical Pacific SST. Dashed line represents approximately 40 cm s^{-1} northward propagation. The 95% significance level corresponds to correlations of ~0.35 (from Chelton and Davis 1982).

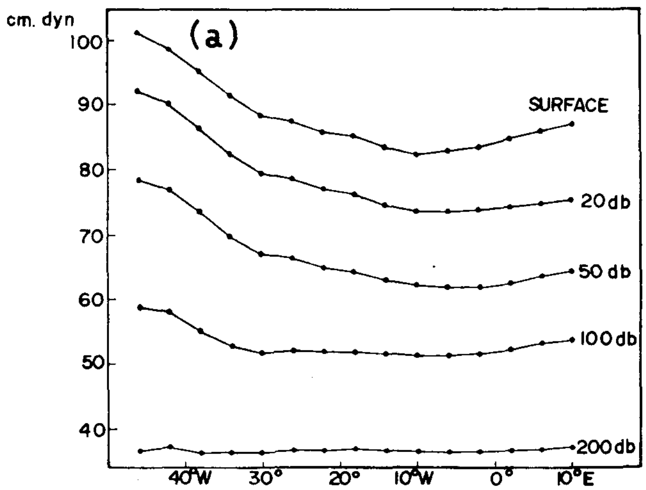
equilibrium with the zonal wind stress and therefore drives an eastward current, the well-known Equatorial Undercurrent.

A simple descriptive way to "explain" this zonal pressure gradient and associated isotherm slopes is to say that the westward winds accumulate water toward the western boundary (piling-up and downwelling), thereby elevating the sea level, while draining water from the eastern boundary and depressing the sea level (upwelling) there. Unfortunately, this common descriptive view lacks in physics, and the concepts of equatorial wave dynamics appear very useful in understanding that the zonal pressure gradient is established some time after a variation of wind in the wake of the propagating waves (Figure 2). If the wind varies regularly at low frequency, then an equilibrium between the zonal wind stress and pressure gradient could appear through a complex superposition of Kelvin waves and Rossby waves (Cane and Sarachik 1981). This appears to be the case at the annual frequency in the Equatorial Atlantic (Katz et al. 1977; Lass et al. 1982; Arnault 1984) but less so in the Equatorial Pacific (Tsuchiya 1979).

The eastern parts of both equatorial oceans are characterized by a clear reversal of sea level and thermocline slopes, which occurs east of 0°-5°W in the Atlantic (Figure 6a) and east of 90°-100°W in the Pacific (Figure 6b). This reversal is not confined to the equator but extends meridionally; for example, it appears all the way to the northern coast of the Gulf of Guinea (Verstraete and Picaut 1983). It is a permanent feature of both eastern oceans, although affected by important variations like the seasonal cycle (Neumann et al. 1975; Meyers 1979; Lukas 1981; Arnault 1984). As noted in the beginning of the previous section, in these eastern ocean regions the winds are predominantly from the south with a small westerly component. As in the central and western equatorial basins, it has been supposed that the eastward wind stress could balance the sharp westward zonal pressure gradient. Recent detailed calculations based on all historical data available, by Lukas (1981) for the Pacific and Arnault (1984) for the Atlantic, reveal that such equilibrium does not hold (by a factor of 3 to 6). In a Sverdrup balance the friction, thermodynamic effects, and nonlinear terms are neglected. Therefore, in the eastern parts of both oceans one or all of these terms must be important.

This could not be the case with the horizontal and vertical frictional effects, because they allow the Equatorial Undercurrent to run all the way from one side of the ocean to the other, and there is no specific reason why friction should increase suddenly when approaching the eastern boundary. Near this boundary, in both equatorial oceans and in their northeast parts, there is an accumulation of warm, low-salinity Tropical Surface Water (Wyrtki, 1967; Berrit, 1973). These light waters are due to heavy rainfall and freshwater outflow. They are present in, at most, the upper 20 meters of the surface layer, and their contribution to the increasing sea-surface height in the east does not appear important (Lukas 1981; Arnault 1984) even in the Gulf of Guinea where there is strong runoff from the Niger and Congo rivers. Finally, it appears that the nonlinear terms—i.e., inertia—probably are the most important contributors to the formation of this slope reversal. These terms are mainly expressed through a deceleration of the Equatorial Undercurrent as it approaches the eastern boundary.

Physically, we could summarize all these balances of forces as follows. In the western and central parts of the equatorial basin the eastward zonal pressure gradient, initiated in the surface layer by the westward winds, drives the Equatorial Undercurrent. When this current approaches the eastern boundary, it has to decelerate until it completely vanishes at the coast. This deceleration must be compensated by a new force, namely the observed eastward pressure gradient. In the first case the pressure gradient creates the current; in the second case the current creates the pressure gradient. According to the synthesis work of Khanaichenko (1974) and the dynamical investigation of McPhaden (1984), the Equatorial Undercurrent belongs to a set of triple-branched, zonal, eastward undercurrents. In the next section we will see that the northern and southern branches of these undercurrents might contribute to the formation of the highly productive tropical thermal domes.



THERMAL DOMES AND EQUATORIAL UNDERCURRENTS

According to Wyrtki (1964), the Costa Rica Dome, found in 1948 from BT observations, consists of a thermal dome situated near 9°N-89°W (Figure 7). It appears to be at the eastern terminus of a ridge in the topography of the thermocline, extending across the Pacific near 10°N, that corresponds to the northern side of the North Equatorial Countercurrent. When this eastward surface current reaches the eastern boundary, it turns north to form the Costa Rica Coastal Current. A cyclonic circulation is therefore completed with the westward North Equatorial Current. The divergence associated with this rotation creates an upward motion believed to be at the dome's origin.

The Guinea Dome, southwest of Dakar, near 10°N-22°W, was first noticed by Rossignol and Meyrueis (1964) and appears clearly, during the northern hemisphere summer, on the historical data analysis of

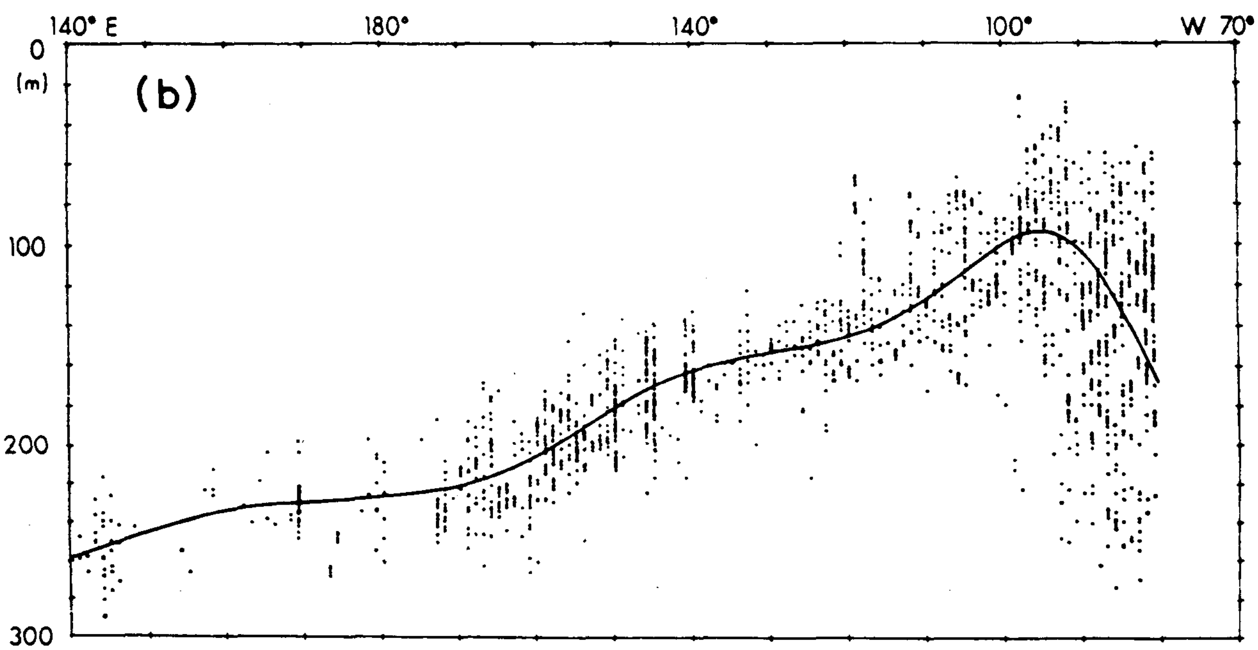


Figure 6. *a*. Mean dynamic height of the 0-, 20-, 50-, 100-, and 200-dbar surface relative to 500 dbar, calculated from a combination of all historical hydrological XBT and BT profiles along the Equatorial Atlantic and between 2°N and 2°S (from Arnault 1984).
b. Depth of the 14°C isotherm. The heavy lines fit all the scattered points obtained with all historical XBT and BT profiles along the Equatorial Pacific and between 1°N and 1°S (from Meyers 1979).

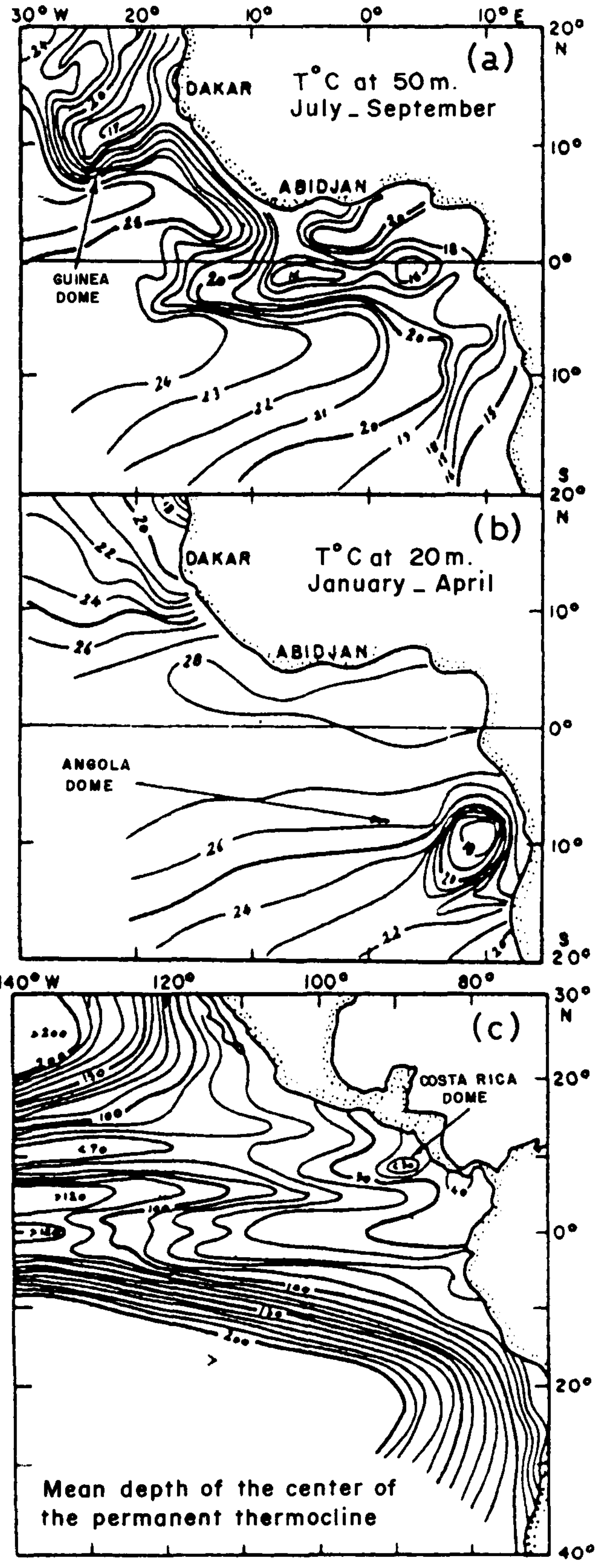


Figure 7. The thermal domes in the eastern tropical Atlantic and Pacific oceans: *a*. Guinea Dome (from Mazeika 1967); *b*. Angola Dome (from Mazeika 1967); *c*. Costa Rica Dome (from Wyrtki 1964).

Mazeika (1967). As in its eastern tropical Pacific counterpart, a cyclonic rotation induced by the surrounding current field seems associated with this dome. Such gyral circulation is not so well defined in the southeast part of the Gulf of Guinea, even though Mazeika's map indicates an Angola Dome near 10°S-9°E during the southern hemisphere summer. A hypothetical Peru Dome might exist near 8°S-85°W, but in the subthermoclinal layers (Voituriez 1981).

In a comprehensive review of phenomena associated with these domes, Voituriez (1981) contests the classical explanation. He noticed that the associated uplifting of the thermal structure is not limited to the thermocline layer but appears farther below (Figure 8). Therefore, the currents more probably involved in the formation of these domes are the North Equatorial Undercurrent (NEUC) and South Equatorial Undercurrent (SEUC). These two subthermocline undercurrents have been carefully documented in both oceans (e.g., Hisard and Rual 1970; Molinari et al. 1981). They take their energy from the Equatorial Undercurrent (McPhaden 1984) in the western and central parts of the equatorial basin and, like this central Equatorial Undercurrent, transport maxima of oxygen (Figure 8b). Near the eastern boundary they are completely detached from the central Equatorial Undercurrent, and when they impinge on this boundary they deflect poleward. Again, this turning motion creates a divergence and an associated upward movement that is probably a part of the domes' mechanism.

Divergence (or convergence) of water mass can also be produced by variations in space of the wind stress. For example, the associated Ekman transport could be greater in one area than in an adjacent area. This leads to a deficit of water mass at the boundary, which in turn is compensated by water from below, i.e., an upwelling. This phenomenon is known as Ekman pumping, and in a steady case is directly related to the wind stress curl ($\partial\tau y/\partial x - \partial\tau x/\partial y$) where τx and τy are the wind stress components. Maps of wind stress curl are therefore very useful in determining possible areas of open ocean upwelling or downwelling. The atlases of Hastenrath and Lamb (1977) and O'Brien and Goldenberg (1982) provide such maps for the entire tropical Atlantic and Pacific oceans, but the wind stress curl near the eastern ocean boundaries is more detailed in charts prepared by Bakun and Nelson[2], thanks to a finer grid resolution. These maps show that the Costa Rica and Guinea domes correspond to regions of favorable cyclonic wind stress curl, which is maximum from May to October. Near the Angola

[2]Bakun, A., and C.S. Nelson. Wind stress curl in the California, Peru, Canary, and Benguela Current systems (Poster). 1984 CalCOFI Conference, Idyllwild, California, October 29-31, 1984.

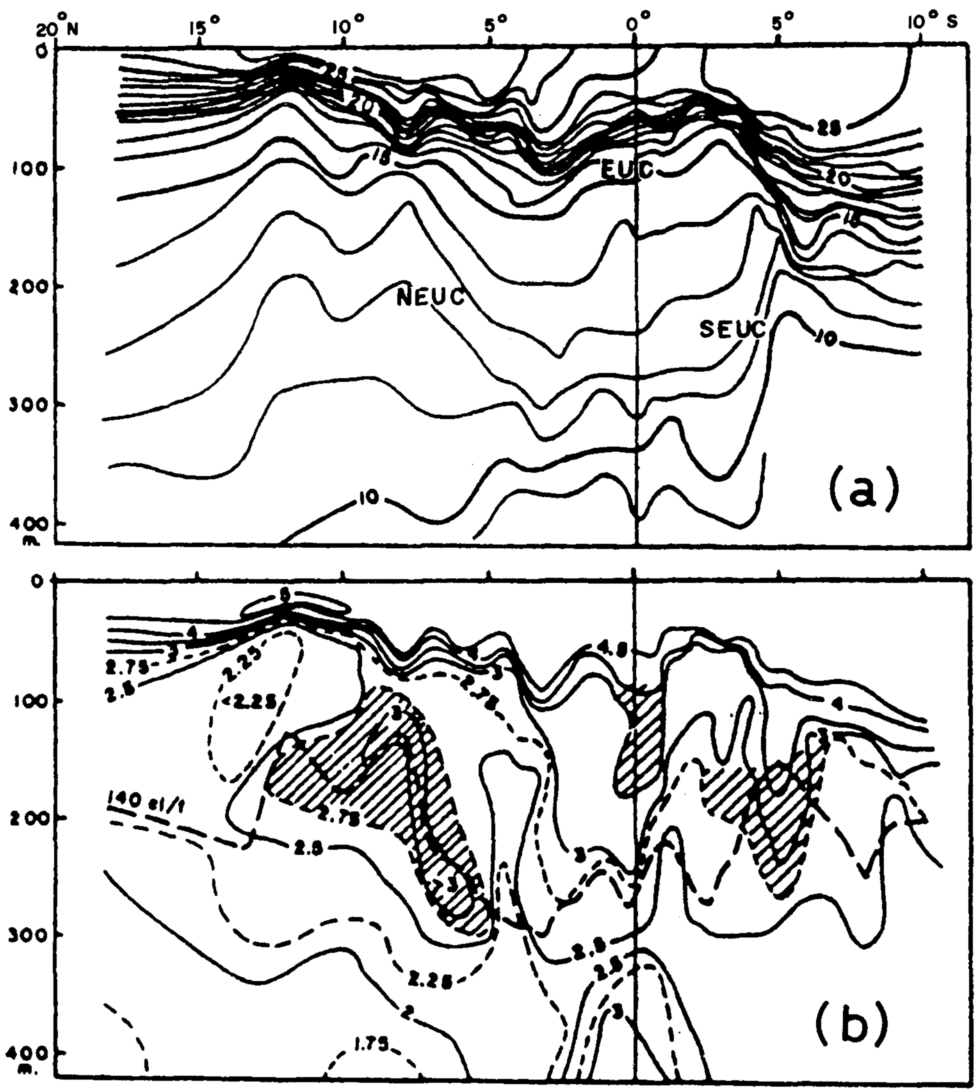


Figure 8. Transequatorial sections in summer 1964 in the Atlantic Ocean at 20°W (EQUALANT 2): a. temperature, b. oxygen. The high oxygen values of the three branches of the equatorial undercurrent system are shaded (from Voituriez 1981).

Dome the wind stress curl is also upwelling-favorable most of the year, but off Peru the wind stress distributions near the doming, noticed by Voituriez (1981) in the deep layers, are not favorable for upwelling development in the surface layer.

Simple models enable us to better understand the formation mechanisms of these domes. A simple mass transport model forced by wind stress was used by Hofmann et al. (1981) to obtain the gross features of wind-driven circulation in the tropical Pacific. The results do not indicate any surface cyclonic circulation in the Costa Rica Dome area and agree with the suggestion of Voituriez (1981) that surface currents are not necessarily the mechanism that forms these domes. The wind generation of the Costa Rica Dome was detailed by Hofmann et al. (1981) through the results of the reduced gravity model by Busalacchi and O'Brien (1980) that was forced by the observed mean seasonal wind stress. For the Atlantic, Busalacchi and Picaut (1983) discussed the results of a similar model in the Guinea and Angola dome regions. In the model, the simulated Guinea Dome appears to be governed only by Ekman pumping. The corresponding model pycnocline is shallow from August to October, and the associated upwelling-favorable wind stress curl is due, as in the Costa Rica Dome, to the northward movement of the Intertropical Convergence Zone. The contribution of Rossby waves in the seasonal variability is significant for the Costa Rica Dome and dominant for the Angola Dome.

CONCLUSION AND DISCUSSION

Thanks to a relatively clear causal relation resulting from particular equatorial dynamics, theoreticians have had great success in explaining the dominant processes at work in the eastern tropical Atlantic and Pacific oceans. Models like those presented in this paper appear to be tremendously helpful in such research. But we must bear in mind that their purpose is to explain, not to simulate. For example, the equatorial waves discussed in this paper, albeit important, are just a part of the chain that induces the global, large-scale phenomenon entitled ENSO (El Niño-Southern Oscillation). However, the recent model results of Busalacchi et al. (1983) clearly demonstrate the link between the changes of the zonal wind stress in the western and central Equatorial Pacific and the El Niño signature along the eastern boundary a few months later. The extension of the remotely forced theory to the Gulf of Guinea by Moore et al. (1978) and the predictive character of such causal relations have enabled Cury and Roy[3] to discuss a predictive model of fishery along the northern gulf coast. They first established a model using fishing effort and local upwelling index. Then they related this upwelling index with the zonal wind stress off Brazil, one month in advance (the time for an eastward Kelvin wave to cross the basin). Finally they discussed the ability of this wind stress to provide the basis for a predictive model of the fishery.

Another approach to prediction could be attempted through the upward phase propagation phenomenon. If the upwelling or downwelling event is remotely forced, we can expect that its signature will arrive first in the deep layer and then in the surface (Figures 3-4). But we must admit that even though observational evidence of horizontal phase propagations is still relatively rare, it is nonetheless more common than for upward propagation.

Most of the analytical and modeling equatorial studies have focused on purely equatorial mechanisms. As far as we know, only Hofmann et al. (1981) and Busalacchi and Picaut (1983) have looked at their model results near the dome regions. Apart from the modeling study of McPhaden (1984), little is known of the mechanism of the northern and southern branches of the equatorial undercurrents, and even less of their relation to the eastern boundary currents. Multilayer models, on a smaller scale than those usually used for large-scale studies, are needed to better understand some of the fascinating physics of the eastern tropical Pacific and Atlantic oceans.

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