Recent Advances in the Knowledge of the Climatic Variations in the Tropical Pacific Ocean

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ABSTRACT

Owing to the free participation of a ships of opportunity network, the SURTROPAC (SURvey of the TROpical PACific) programme is continuously surveying the Western Tropical Pacific with surface measurements (meteorology, temperature and salinity) and subsurface measurements (0-400m thermal profile). The Western Pacific Ocean may be described from two perspectives: (a) as a heat pool that concentrates thermal energy and returns it to the atmosphere and the rest of the ocean; and (b) as an area particularly sensitive to inter-annual oscillations associated with the El Nino phenomenon.

The heat content of the Western Tropical Pacific seems to be only slightly affected by seasonal variations. Rather, a large amount of its energy is exchanged with the atmosphere particularly in the vicinity of the intertropical convergence zone of the winds. A small thermal variation can induce a strong variation of energetic transfer to the atmosphere, mainly in the form of latent heat. From this latent heat transfer and the estimated rainfall, an evaporation-precipitation balance can be calculated that is similar to the balance computed from the surface salinity.

In the Western Pacific Ocean, the interannual variation is usually connected to the appearance of the El Nino phenomenon and consequently to the Southern Oscillation. Conditions related to El Nino can be distinguished as pre-El Nino conditions, conditions occurring during the El Nino itself and post-El Nino conditions. These different phases of El Nino are described through consideration of the thermal profiles, sea levels and surface temperatures in the whole Pacific. The climatic consequences in the Western Pacific are also pointed out.
INTRODUCTION

Until 1970, oceanography was practiced with the help of research vessels that carried out "oceanographic cruises" only when available. These cruises gave a very accurate snapshot description of the ocean, valid only for localised areas and for short time-scales of a few weeks. Repetition of these cruises was necessary to follow the evolution in time of the oceanographic conditions, and any such programme was very expensive.

Between 1970 and 1980, the energy crisis rendered the fuel necessary for research vessels costly. On the other hand, the energy shortage also encouraged climatic research, since the forecasting of seasonal fluctuations helps to optimise fuel consumption. Also during this period, climatic research was given further impetus by the occurrence in the Pacific of two strong El Nino episodes (1972 and 1976), which seemed to have climatic effects on the entire planet, ranging from droughts in Australia, the southwest Pacific and northeast Brazil, to severe winters in the USA.
The first El Nino studies pointed out the great influence of the tropical ocean on the climate and particularly the influence of the Pacific Ocean. But in order to understand the climate and then to forecast it, it became evident that it would be necessary to follow the thermal state of the Pacific Ocean continually, and to find a monitoring method to measure parameters relevant to the understanding of the climate. Satellite monitoring was the first candidate, but for oceanographers, satellites are still not able to measure anything useful other than sea surface temperature. On the other hand, a simple, economic and very flexible way of obtaining data was to make use of the thousands of merchant ships continuously plying the oceans. Thanks to the goodwill of the ship-officers, a great variety of measurements can be done (physical or biological, surface or sub-surface) without the need of scientists aboard. Therefore, as present satellites cannot measure anything below the surface, the use of merchant ships seems to be the most suitable tool for the continuous study of the ocean which may lead to the understanding and perhaps prediction of world climatic conditions.

For this reason, as early as 1969, the officers of five Japanese ore-carriers were persuaded to aid scientists voluntarily by performing, free of cost, the measurement of sea surface temperature and the gathering of surface samples between New Caledonia and Japan. The climatic features observed in the south west Pacific during the 1972 El Nino induced the extension of this observational network in 1974 to other shipping routes, until the present network was reached. The whole study is now called SURTROPAC (SURvey of the TROPical PACific).

SURTROPAC has profited from a good oceanographic heritage. Through the work of LEGAND and ROTSCHI, numerous data have been obtained from 1956 to 1964 by Centre ORSTOM de Nouméa, leading to determination of water masses and seasonal variations in the Coral Sea. Two major experiments, EQUIPAC in 1956 and the International Geophysical Year (IGY) in 1958 described the conditions of the tropical Pacific; the IGY cruises permitted, by chance, the determination of the conditions that prevailed during the 1957-58 El Nino. From 1965 to 1972, the arrival of the Research Vessel CORIOLIS gave a new impulse to the research at the Centre ORSTOM de Nouméa. It made ten visits to the 170°E meridian from 20°S to 5°N between 1965 and 1968, as an oceanographic shuttle (MAGNIER, ROTSCHI, RUAL and COLIN, 1973). During this same time frame (1967-68), the international EASTROPAC expeditions described the water masses and seasonal variations in the Eastern Pacific. In addition, some surface sampling experiments were conducted: in the Tasman and Coral seas by CSIRO Australia and in the Pacific between Hawaii and Tahiti by the Bureau of Commercial Fisheries of Honolulu.

2. METHODS OF THE SURTROPAC PROGRAMME

The tropical Pacific Ocean is monitored from merchant ships-of-opportunity that report to the Centre ORSTOM de Nouméa and to the Centre ORSTOM de Papeete, where the data are gathered and processed (Fig.1).

The following shipping routes are operated:

- Noumea - Hong Kong
- Noumea - Japan
- Noumea - California
- Noumea - Tahiti - Panama
- Noumea - New Zealand
- Noumea - Australia
- Papeete - Honolulu
- Papeete - California
- Papeete - Panama
- Papeete - New Zealand

Surface and subsurface data are gathered including meteorological parameters, temperature, salinity, and biological measurements (phytoplankton, chlorophyll, zooplankton) at the surface, and a temperature profile from the surface to 400m depth using XBTs.

The contacts with the observers are essential to maintain the experiment and ensure its continued success. As the experiment is based on the voluntary work of the officers and crews of the ships-of-opportunity, it is essential to maintain their enthusiasm and motivation (DONGUY, 1980).
3. THE WESTERN TROPICAL PACIFIC AS A HEAT POOL

The tropical ocean strongly influences the Earth's climate due to its capacity both to store the heat provided by the sun and to export it to other geographic regions. Owing to its great size, the Pacific Ocean accomplishes a major part of this function. In tropical areas, the usual westward currents concentrate much of this heat in the western Pacific. Schematically, it seems that in the eastern part of the Pacific the ocean is absorbing energy from the atmosphere (WEARE, STRUB and SAMUEL, 1981a,b), whereas in the western part the ocean is transferring the heat to the atmosphere. A study of the mechanisms and the annual variations of this heat transfer is the main goal of the international programme TROPICAL OCEAN and GLOBAL ATMOSPHERE (TOGA). As a preliminary step to this study, it is appropriate to review the present status of knowledge about heat content in the Western Tropical Pacific.

3.1 Thermal pattern

When trade winds blow along the equatorial Pacific, the westward wind stress balances an eastward pressure gradient that is associated with an eastward decrease in the depth of the thermocline. Consequently, the thermocline is much deeper in the western Pacific than in the eastern Pacific (150m in the west, less than 50m in the east). In the western Pacific, energy is stored between the surface and 150m depth in the form of heat, and is partly returned to the atmosphere by air-sea exchange supplemented by atmospheric convection.

In the western equatorial Pacific, the seasonal fluctuations of the heat content are small (Fig.2) but they are large in the eastern equatorial Pacific (DONGUY and HENIN, 1981a). In the northern summer, however, there is a zone of relatively large heat content which extends eastward along 5°-10°N whereas in the southern summer it extends eastward near 8°S. The heat is transported to the east by the countercurrents characteristic of these latitudes (HENIN and DONGUY, 1980).

3.2 Ocean-Atmosphere relations

The western Tropical Pacific exchanges a large quantity of energy with the atmosphere through the sea surface. These exchanges occur mainly close to the Intertropical Convergence Zone of the wind (ATKINSON and SADLER, 1970) (Fig.4). West of 180° longitude, this convergence appears at 10-15°N latitude in the Northern Hemisphere during the northern summer, and at 10-15°S latitude in the Southern Hemisphere during the southern summer. However, the convergence does not disappear completely in the Southern Hemisphere in the northern summer when it constitutes the South Pacific Convergence Zone.

East of 180° longitude, the convergence usually remains between 5-10°N, but during February-March in the vicinity of 100°W it lies either close to the equator or just south of it.

The position of the Convergence Zone of the wind is also well defined in the charts of wind divergence (O'BRIEN & GOLDENBERG, 1982) which show that a minimum of divergence occurs in the
FIG. 2. Mean temperature (0-100m) between New Caledonia and Japan for 1979-1981. (from DONGUY and HENIN, 1981a).

south during January and in the north during August. This kind of seasonal variation had been already noticed by WYRTKI and MEYERS (1976) and MEYERS (1979).

The influence of these convergence zones is important for the ocean-atmosphere exchange and consequently for the climate itself. They are areas of strong atmospheric convection: the warm humid air resulting from the contact with the ocean, rises up and its water vapour condenses releasing latent heat and reinforcing the convective activity. However, the role of the convergence zone is obscure: the wind there is weak, whereas the evaporation is proportional to the wind speed. Furthermore, the results of WEARE, STRUB and SAMUEL (1981a,b) do not indicate a great heat exchange within these zones. Probably most of the latent heat released in the convergence zones comes from water vapour carried in from outside areas rather than from local evaporation.
FIG. 3. (a) Mean temperature (0-100m) in August-September 1956. (b) Mean temperature (0-100m) in October-December 1961. (from HENIN and DONGUY, 1980a)
3.3 Thermal balance

Several studies have been reported on this subject (HASTENRATH, 1980; WEARE, STRUB and SAMUEL 1981b; ESSENSEN and KUSHNIR, 1981; STEVENSON and NILER, 1983; REED, 1983; OORT and VON DER HAAR, 1976; TALLEY, 1984; and ENFIELD, 1986).

The net heat transferred between ocean and atmosphere can be expressed by:

\[ Q_n = Q_s - Q_i - Q_h - Q_l, \]

where
- \( Q_n \) = net heat transfer
- \( Q_s \) = incident short-wave energy flux
- \( Q_i \) = net infra-red radiation transfer
- \( Q_h \) = sensible heat lost from the ocean
- \( Q_l \) = latent heat lost by evaporation

These parameters are estimated by empirical bulk aerodynamical formulae but in the western Pacific only a few data are available to compute them. According to WEARE, STRUB and SAMUEL, (1981b), in the western Equatorial Pacific the ocean is absorbing energy but at a much lower rate than in the eastern Equatorial Pacific. There is also an energy transfer from the

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**FIG.4.** Seasonal oscillation of the Convergence Zone of the wind in the Western Pacific. Above: convergence zone at 10°N in September-October. Below: convergence zone at 15°S in February-March. (From ATKINSON and SADLER, 1970).
ocean to the atmosphere south of 15°S. ENFIELD (1986) made a critical analysis of the climatologies used by the previous studies.

The main mechanism of the energy transfer from the ocean to the atmosphere is probably the transfer of latent heat by evaporation, which is maximum in the Tropical Western Pacific (Fig. 5) and minimum in the Tropical Eastern Pacific (WEARE, STRUB and SAMUEL, 1981a).

![Figure 5. Flux of latent heat through the sea surface by evaporation in Wm⁻², 1957-76 average (from WEARE, STRUB and SAMUEL, 1981b)](image)

3.4 Influence of sea surface temperature variations on the atmosphere

The response of the atmospheric circulation to sea surface anomalies was modelled by WEBSTER (1981, 1982) with interesting results. However, the anomalies used in his model were strong (5°C), and although such anomalies can be observed in the central and eastern Pacific, they do not occur in the western tropical Pacific. There, the thermal anomalies are weak, always less than 2°C, but their persistence is long, possibly as long as one year at the equator (REBERT and MORLIERE, in press). Through a simulation experiment, the assumption has been tested that a weak surface thermal variation may induce a strong change in the energy transfer into the atmosphere, particularly in the form of latent heat. Along the shipping route between Noumea and Japan, sea surface data from ships-of-opportunity have been used by 2° latitude bands and a sea surface temperature anomaly of the same size as the annual variation (0.5-0.7°C) has been applied, the other parameters (wind velocity, water-air temperature difference, coefficient of turbulent exchange) being equal to the annual mean. The results (Fig. 6) show that the mean value of $Q_0 + Q_1$ (120 Wm⁻²) is less than the one computed by WEARE, STRUB and SAMUEL (1981b). The mean value has maxima at 5°S and 14°S. The ratio of transfer between a warm state and a cold state relative to the average is 50%, whereas the variation in temperature is only between 1-1.4°C. This result applies when the sea surface temperature is above 25°C due to the non-linear nature of $Q_1$ as a consequence of the exponential behaviour of the saturation vapour pressure with the temperature (GILL and RASMUSSON, 1983). This simulation seems to confirm that a weak variation of the surface
temperature above 25°C leads to strong variations of energy transfer. NEWELL (1979) showed also that "tropical sea surface temperature cannot go above about 30°C because at higher temperature loss of energy by evaporation far exceeds energy input".

3.5 Evaporation - Precipitation balance

By comparing the annual estimation of rainfall in the tropical Pacific (TAYLOR, 1973) to the latent heat transfer, WEARE et al. (1981b) established an annual Evaporation - Precipitation balance expressed in flux units (Wm$^{-2}$). The balance may be also expressed in millimetres per year using the equivalence 1Wm$^{-2}$ = 12.7mm y$^{-1}$. Figure 7 shows that the E-P balance is negative west of 180° and with two extrema of 1500mm y$^{-1}$, one on the equator north of New Guinea, the other at 10°S, 170°E. Consequently, in this area, there is a supply of moisture coming from surrounding regions, corresponding to the Intertropical Convergence Zone of the Wind. East of 180°, the balance is positive with a maximum (+1000mm y$^{-1}$) located at the equator, corresponding to a source of moisture for the atmosphere.
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The surface of moisture can be considered as the result of the Evaporation-Precipitation balance and of a zonal, meridional and vertical advection of moisture. The surface salinities gathered by the ships-of-opportunity since 1969 are valuable data which were used by DONGUY and HENIN (1978a) to draw the annual mean of the surface salinity (Fig. 8) and to compare it to the E-P balance. An E-P minimum located at 10°S coincides with a salinity minimum (DONGUY and HENIN, 1975), and east of 180°, the equatorial maximum coincides with the equatorial salinity maximum. If the salinity increases linearly from the surface to a depth $d$, where a constant salinity $S_1$ exists, the height of fresh water $h$ that can induce the surface freshening is expressed by:

$$ h = \frac{S_1 - S_2}{d} $$

$S_2$ is the surface salinity (DONGUY and HENIN, 1976a, 1980a). The distribution of the computed fresh water (Fig. 9) is not quantitatively in agreement with the E-P balance (Fig. 7). However, the equatorial E-P maximum coincides with a fresh water minimum, and the E-P minimum at 10°S with a fresh water maximum. The discrepancy observed north of New Guinea between the amount of fresh water from latent heat (Fig. 7) and the one from sea surface salinity...
FIG. 8. Mean surface salinity in the Western Pacific (from DONGUY and HENIN, 1978a).

(Fig. 9) is possibly due to the river runoff. Surface salinity depends upon three factors:

1. Evaporation - Precipitation balance
2. Horizontal and vertical advection
3. River run off

Neglecting the last factor and the diffusion terms, one may write the following equation for stationary regime:

$$\frac{\partial S}{\partial t} + u \frac{\partial S}{\partial x} + v \frac{\partial S}{\partial y} = \frac{P-E}{d}(x, y)$$

which means that advection is compensated by the P-E balance.

At the equator, with easterlies blowing (Fig. 8), a high salinity tongue appears (>35×10^{-3}) because of the combined influence of the equatorial upwelling, westward advection and a positive E-P balance (DONGUY and MORLIERE, 1983). On the other hand, when westerlies blow, low salinity water extends over the equatorial area due to the lack of equatorial upwelling, the eastward advection and a negative E-P balance (DONGUY and HENIN, 1976b; LEMASSON and PITON, 1968). The salinity maxima are subject to seasonal variations: the salinity maximum
in the central South Pacific is usually less extensive in February-March (weaker SE trades) than in August-September (stronger SE trades). The salinity maximum located southwest of New Caledonia seems to have an opposite regime (DONGUY and HENIN, 1977).

In conclusion, it seems that in the Western Pacific the surface salinity can be considered as a mirror of climatic conditions (DONGUY and HENIN, 1978b).

4. INFLUENCE OF THE WIND IN THE WESTERN PACIFIC

The equatorial and tropical western Pacific Ocean seems not to have strong seasonal variations like those of the eastern Pacific because the heat budget is dominated by different processes in the two regions. In the west, the balance of the exchanges is primarily between surface flux and vertical diffusion, while to the east, advection and meridional diffusion are also important (ENFIELD, 1986). In addition, in the west, the heat content from the surface to the thermocline is regulated by a balance between insolation and evaporation. The effect of the solar forcing is to increase the sea surface temperature but simultaneously latent heat flux changes in a non linear way and consequently the increase is counteracted and the sea surface temperature does not exceed 30°C (NEWELL, 1979). This mechanism explains
why the heat content does not seem to be affected by seasonal variations (Fig.2) (DONGUY and HENIN, 1981a). On the other hand, the surface salinity, which reflects the climate conditions, shows clear seasonal variations (DONGUY and HENIN, 1978a; HENIN and DONGUY, 1970b; DONGUY and MORLIERE, 1983).

On the contrary, the year-to-year variations of heat content are important in the western tropical Pacific and are able to mask seasonal variations (DONGUY and HENIN, 1976b). They are usually connected to the occurrence of El Nino.

4.1 El Nino

El Nino is the name of the weak seasonal current that appears along the Pacific coast of South America at Christmas. This current usually brings southward, nutrient-poor, warm and low salinity waters from the equatorial region and northern hemisphere and it temporarily displaces the Peru Current, which normally carries cold, high salinity and nutrient-rich waters northward. On the land anomalous rainfalls occur, and birds and fishes either migrate out of the region or suffer high mortalities through starvation.

Until 1970, El Nino was considered an important event but only of local significance. Now, we know that this phenomenon observed along the South America coast is only a small part of a worldwide event. Numerous climatic anomalies appear at the same time on the entire planet such as droughts in Australia, Indonesia (QUINN, ZOFF, SHORT and KOU YANG, 1978), the southwest Pacific (DONGUY and HENIN, 1980b), and northeast Brazil (KOUSKI, 1984); severe winters in USA (RASMUSSON and WALLACE, 1983; PHILANDER, 1983), and Japan (YAMANAKA, 1984; WHITE and YOU HAI HE, 1986), and the anomalous occurrence of cyclones in the central Pacific (DONGUY, BEGAUD, EBSTEIN and CALVEZ, 1979; REVELL and GOULTER, 1986).

4.2 Southern Oscillation

The worldwide appearance of the oceanic El Nino is connected to the atmospheric Southern Oscillation, and consequently the two phenomena are linked in what are now termed El Nino/Southern Oscillation (ENSO) episodes. The Southern Oscillation may be indexed by the anomaly from the long-term monthly mean of the difference between the atmospheric pressure at Tahiti, near the centre of the South Pacific anticyclone, and at Darwin Australia, in the area of the Indonesian low pressure centre.

The difference of atmospheric pressure from one extremity of the equator to the other in the Pacific Ocean is related to the strength of the Walker circulation (Fig.10), which includes ascending movements of the air masses above the warm area of the western Pacific (low pressure) and the descending movements above the cool area of the eastern Pacific (high pressure) (WYRTKI, 1979a, 1982; BJERKNES, 1969). Above the Indian Ocean another Walker Cell exists but with a reversed circulation; the ascending branch is above the warm areas of the Eastern Indian Ocean and joins the Western Pacific ascending one.
FIG. 10 Schematic representation of atmospheric and oceanographic conditions for the pre and post El Nino periods at the equator (from DONGUY, DESSIER, ELDIN, MORLIERE and MEYERS, 1984c).

FIG. 11. Atmospheric pressure anomaly at Darwin (Australia) and at Tahiti computed for a composite including 9 El Ninos (from VAN LOON and SHEA, 1984).
When the Southern Oscillation Index (SOI) is high, easterly trade winds are intense along the equator (Fig. 11). This occurs during the period preceding El Nino, called the pre-El Nino period. According to WYRTKI, STROUP, PATZERT, WILLIAMS and QUINN (1976), when the pressure difference between Tahiti and Darwin reaches 13 millibars, a critical point is reached when relaxation of trade winds is observed in the equatorial zone and an El Nino event appears; however, this statement is no longer fully comprehensive as the 1982-83 El Nino started without this condition. When the SOI is low or negative easterly trade winds are weak or even west wind may blow on the equator, as happens during El Nino (Fig. 11) (VAN LOON and SHEA, 1984).

4.3 Pre-El Nino conditions

The Pre-El Nino conditions are characterised by a well-developed Walker Cell in the atmosphere (Fig. 10) and consequently by a period of prevailing trade winds associated with a high SOI. These trade winds blowing along the equator induce a strong equatorial upwelling. WYRTKI (1981) has described the eastern Pacific aspect of this upwelling, but in the Western Pacific, equatorial upwelling occurs in a different way. The trade winds, pushing waters westward from the eastern Pacific along the equator, induce a piling up of the water in the western Pacific which is detectable as an increase of sea level. By baroclinic adjustment, isotherms and thermocline deepen by several tens of metres. Simultaneously, trade winds at the equator induce equatorial upwelling with lifting up of the isotherms. In pre-El Nino in the Western Pacific, there is the following perplexing situation: thermocline deepens by baroclinic adjustment but near-surface isotherms rise by upwelling. However, the equatorial upwelling occurs only in a 100m thick surface layer so that the negative temperature anomaly at the surface is either weak or non-existent.

During the pre-El Nino period, the heat content in the equatorial area of the western Pacific is very large and the ocean atmosphere exchanges are particularly intense. A tongue of high surface salinity extends along the equator, due to a combination of westward advection, the lack of rainfall and a mixing with the more saline subsurface water. Surface high salinity water is connected to a maximum salinity core located at the thermocline level as shown by data collected during the cruise on BORA 2 carried out by the Centre ORSTOM de Noumea in 1966 (Fig. 12) (ROTSCHI, HISARD and JARRIGE, 1972).

The prevalence of trade winds in the equatorial area (Table 1) induces the existence of a climatic scenario characteristic of the pre-El Nino in the western Pacific (DONGUY, 1982). So, easterly winds blowing in the equatorial area (5°N-5°S) induce equatorial upwelling and high surface salinity. Surface current is westward, the sea level increases westward and the thermocline deepens with a great heat content. The equatorial area is characterised by a positive Evaporation-Precipitation balance and consequently by an arid climate. The tropical area (5°N-15°N, 5°S-15°S) on the other hand, is characterised by the presence of the convergence of the winds. The curl of the wind stress is affected and consequently the divergence of Ekman transport may vary, inducing a change in the depth of the mixed layer: this effect is termed Ekman pumping. During the pre-El Nino period, Ekman pumping
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Table 1. Sequence of events during pre El Nino and post El Nino phases in the western Pacific Ocean

<table>
<thead>
<tr>
<th>EQUATOR 5°N-5°S</th>
<th>EAST WIND</th>
<th>UPWELLING</th>
<th>HIGH SURFACE SALINITY</th>
<th>WESTWARD CIRCULATION</th>
<th>HIGH SEA LEVEL</th>
<th>DEEP THERMAL UNION</th>
<th>GREAT HEAT CONTENT</th>
<th>NO RAIN</th>
<th>DROUGHT</th>
</tr>
</thead>
<tbody>
<tr>
<td>TROPICAL ZONE 5°N-15°N 5°S-15°S</td>
<td>PRESENCE ITBG</td>
<td>DOWNWELLING</td>
<td>LOW SURFACE SALINITY</td>
<td>EASTWARD CIRCULATION</td>
<td>HIGH SEA LEVEL</td>
<td>DEEP THERMAL UNION</td>
<td>GREAT HEAT CONTENT</td>
<td>TROPICAL STORMS RAIN</td>
<td>NORMAL RAINY SEASON</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>EQUATOR 5°N-5°S</th>
<th>PRESENCE ITGB</th>
<th>NO UPWELLING</th>
<th>LOW SURFACE SALINITY</th>
<th>EASTWARD CIRCULATION</th>
<th>LOW SEA LEVEL</th>
<th>SHALLOW THERMAL UNION</th>
<th>SMALL HEAT CONTENT</th>
<th>RAIN</th>
<th>EXTRA RAIN</th>
</tr>
</thead>
<tbody>
<tr>
<td>TROPICAL ZONE 5°N-15°N 5°S-15°S</td>
<td>EAST WIND</td>
<td>HIGH SURFACE SALINITY</td>
<td>WESTWARD CIRCULATION</td>
<td>EASTERLY</td>
<td>SHALLOW THERMAL UNION</td>
<td>SMALL HEAT CONTENT</td>
<td>FEW TROPICAL STORMS</td>
<td>RAIN</td>
<td>NO DROUGHT</td>
</tr>
</tbody>
</table>

**FIG. 12.** Vertical distribution of salinity at 170°E observed along a transect from 5°N to 20°S during the cruise BORA 2 (March-April 1966), during pre El Nino conditions.
induces a negative displacement of the isotherms (DONGUY, HENIN, MORLIERE and REBERT (1982) and consequently deepening of the thermocline, in phase with the deepening observed at the equator. A great heat content builds up in the whole western Tropical Pacific. Tropical cyclones, which need a large heat source, atmospheric instability and sufficient Coriolis force can originate there. Their influence, together with the presence of the convergence zone of the wind induces a rainy season in the tropical western Pacific.

The salient features of the circulation at 170°E have been pointed out by MERLE, ROTSCHI and VOITURIEZ (1969), but they vary with the hydroclimatic conditions (DONGUY, ELDIN, MORLIERE and REBERT, 1984d). During the pre-El Nino period, the circulation is characterised by the prevalence of the westward currents from 15°N to 15°S (Fig.13). The zonal surface circulation is in geostrophic balance with the meridional topography of the thermocline (DONGUY and HENIN, 1983). From 5°N to 5°S, the Equatorial Current* is westward. From 5°N to 10°N, the North Equatorial Counter Current (NECC) is eastward, as is the South Equatorial Counter Current from 5°S to 10°S. From 10°N to 15°N, the North Equatorial Current is westward, as is the South Equatorial Current at 10°S to 15°S. The transport of the North Equatorial Current is more important than that of the South Equatorial Current. WYRTKI and KILONSKY (1984) have studied the meridional circulation during the Hawaii to Tahiti Shuttle Experiment in the central Pacific; they describe an equatorial current system roughly similar but more complicated in its details.

* The terminology used for the Equatorial Pacific currents is slightly different according to usage by French than other scientists. For French scientists, the westward current located on the equator is called Equatorial Current, and the westward one located south of the equator is the South Equatorial Current.
4.4 El Niño phenomenon itself

El Niño seems to start with a decrease of the SOI and consequently with the relaxation of the trade winds in the equatorial area. This relaxation occurs usually between November and January and the equatorial trade winds may reverse from easterly to westerly. The water with its enormous heat content stored in the Western Equatorial Pacific is, therefore, no longer contained, and sea level therefore falls. A small warming of the surface water close to 180° longitude, due probably to the cessation of equatorial upwelling, is observed. The drainage of the water stored in the western Pacific occurs initially via an equatorial jet (Donguy, Eldin, Morlière and Rebert, 1984a) and then by the North Equatorial Counter Current (Wyrtki, 1979b; Meyers and Donguy, 1984) which is strongly intensified at this time (Fig.13).

The relaxation of the trade winds in October-December in the western Pacific and sometimes in the central Pacific, combines its effect with the reversal of the normally easterly trade winds into westerlies, to induce both a trapped Kelvin wave at the equator which propagates eastward with a velocity up to 3\( \text{ms}^{-1} \), and also equatorial Rossby waves which propagate westward with velocities less than 1\( \text{ms}^{-1} \). This incident Kelvin wave increases the sea level and induces a deepening of isotherms and thermocline. The Kelvin wave reaches the South American coast 2-3 months later where a deepening of the thermocline leads to an increase of heat content.

The models of Busalacchi and O'Brien (1981) and Busalacchi, Takeuchi and O'Brien (1983) describe the scenario through consideration of the variations of sea level and the pycnocline depth. Kelvin waves reflect off the eastern boundary as Rossby waves. Consequently, a reflected Rossby wave would propagate westward and would induce a rise of the equatorial thermocline in the western Pacific. In agreement with these models, correlations between wind stress in the western Pacific and the thermocline depth in the eastern and then in the western Pacific have been pointed out (Rebert, Donguy, Eldin and Morlière, 1983).

In 1972-73 and in 1982-83, a deepening of the thermocline has been observed from the west to the east along the equator, propagating eastward with a velocity of 0.5-1\( \text{ms}^{-1} \) (Fig.14), closely associated to the propagation of west wind anomalies (Donguy, Eldin, Morlière, Rebert, & Meyers, 1984b). This deepening could result from a combination of free and forced equatorial waves.

However, it is possible that El Niño does not originate only in the Western Pacific. Rasmussen and Carpenter (1982) have noticed in the oceanic area off northern Chile a trend for the surface waters to warm in November-December in the year before El Niño (El Niño-1), whereas during the El Niño year the surface waters usually reach a maximum temperature in April.
Looking at the evolution of a typical El Nino either in 1972 or 1976, it is possible to analyse the sequence of events by distinguishing the characteristic phases of the ENSO development and by using the following available data:

(a) The wind stress (BUSALACCHI et al 1983)
(b) The sea level (WYRTKI, 1977)
(c) The thermal structure along the South American coast (WYRTKI, 1975).
(d) The thermal structure along the equator from MBT, XBT and hydrocasts averaged from 1°N to 1°S.
(e) The composite sea surface temperature from six El Nino events over a 30 year period (RASMUSSON and CARPENTER, 1982).

These data are numerous enough to describe the following stages already defined by RASMUSSON and CARPENTER (1982):

4.4.1. The onset phase. The onset phase is from November of the year before El Nino (El Nino-I) to January of El Nino year, which is in fact the end of the pre-El Nino period just before the sudden decrease of the winds west of 180° (Fig.15). The sea level (+10cm) is high in the Western Pacific and low (-8cm) in the eastern Pacific. The vertical thermal structure (Fig.16) along the equator shows a great heat content west of 180° in agreement with high sea level. The sea surface temperature is normal but in December a light maximum close to 180° and the beginning of a warming in the south eastern Pacific are noticed. However, until at least September-October, the upwelling is observed along South American coast.
and the thermocline shoals toward shore (Fig. 15). The trade winds are blowing along most of the equator pushing warm water to the western Pacific.

![Diagram](https://example.com/diagram.png)


![Diagram](https://example.com/diagram.png)

**FIG. 16.** Left: Schematic representation of the equatorial conditions during the onset phase of El Niño. Right: Thermal profile at the equator in November 1971, from data averaged from 1°N to 1°S.
The peak phase (March-May) covers the period when the El Nino phenomenon first appears on the South American coast (Fig. 17). Instead of having a sloping sea surface in the Western Pacific, the sea level seems to lack any gradient from 150°E to 80°W. The vertical thermal structure (Fig. 18) along the equator still shows a strong heat content in the Western Pacific, whereas the thermocline deepens off the coasts of Ecuador and Peru, and a positive anomaly of the sea surface temperature appears. In the Western Pacific, the trade winds have been abruptly replaced by westerlies in December. Consequently the slope of the sea surface disappears and the corresponding water mass has been drained off eastward by the Equatorial jet and the North Equatorial Counter Current. The Walker cell which occupied the whole equatorial Pacific during the preceding November, has been displaced eastward with an ascending branch at 180° associated with strong rainfall.

Transition phase. The transition phase (August-October) covers the period when the El Nino effects are progressing in the Pacific (Fig. 19). The sea level is low in the Western Pacific and high in the Eastern Pacific. The vertical thermal structure (Fig. 20) along the equator shows a maximum anomaly of heat content in the Eastern Pacific. A new phenomenon appears which is the shallowing of near-surface isotherms in the Western Pacific. The positive temperature anomaly is strengthened, extending along the equator, reaching 180° in the west and linking up with the one already existing there. However, near the South American coast, the temperature anomaly has progressively been diminishing since June. West winds blow in the Western Pacific as far east as 170°W, whereas light trade winds persist east of 170°W. The Pacific Walker cell shifts eastward (Fig. 20), and the Indian Ocean cell appears in the Western Pacific. Between them, an ascending branch located near 170°W brings strong precipitation.

Mature phase. During the mature phase (December-February), the phenomena associated with El Nino reach their maximum development in the Pacific (Fig. 21). The sea level is very low in the Western Pacific and the sea surface is almost horizontal from the Central Pacific to Eastern Pacific (differences are noticed between 1973 and 1977). The vertical thermal structure (Fig. 22) along the equator implies a high heat content in the Central Pacific with isotherms close to the surface in the Western Pacific. A positive 1°C sea surface temperature anomaly extends along the equator east of 180°. Along American coasts, the sea surface temperature anomaly which has reached a second peak in November-December, has returned to near normal. The ascending branch of the Walker Cell is located in the central Pacific and coincides with the heat content maximum. It is also associated with a rainfall maximum.

1982-83 El Nino

The above scenario is valid for a typical El Nino such as the strong 1972 or the moderate 1976 event. The 1982-83 event was atypical. Although the composite scenario has been observed as for prior episodes, the 1982-83 El Nino showed many peculiarities:

FIG. 18. Left: Schematic representation of the equatorial conditions during the peak phase of El Nino. Right: Thermal profile at the equator in February 1972, from data averaged from 1°N to 1°S.

FIG. 20. Left: Schematic representation of the equatorial conditions during the transition phase of El Nino. Right: Thermal profile at the equator in October 1972 from data averaged from 1°N to 1°S.

FIG. 22. Left: Schematic representation of the equatorial conditions during the mature phase of El Nino. Right: Thermal profile at the equator in January 1973, from data averaged from 1°N to 1°S.
(a) The build up to the event through a warm water accumulation in the Western Pacific was not as obvious at the equator as for the preceding episodes (MEYERS and DONGUY, 1983) but extended throughout the whole tropical Western Pacific from 15°N to 15°S (WYRTKI, 1985).

(b) At the start of the event, the SOI was weak with a small pressure difference between Darwin and Tahiti.

(c) The 1982-83 event appeared in May as the consequence of the late relaxation of the trade wind (CANE, 1983) and not in December as suggested by the El Nino name. In contrast, the preceding events seemed to be connected to the seasonal cycle, as each year from December to March, at the El Nino time, warm waters appear in the equatorial area of the Eastern Pacific (PHILANDER, 1985).

(d) The 1982-83 event seems to have started by a wind relaxation and a sea surface heating in the Central Pacific, instead of in the Western Pacific (KERR, 1982; RASMUSSON and WALLACE, 1983). The temperature anomaly then propagated eastward. Usually, warm water appears first off American coasts and seems to propagate westward.

(e) The intensity of the event greatly exceeded the observations for prior episodes off South American coast and the sea surface temperature anomaly reached 7°C (GILL and RASMUSSON, 1983), easterly trade winds were replaced by west wind along the entire Equatorial Pacific and the slope of the ocean remained negligible during several months (DONGUY, ELDIN, MORLIERE, ROBERT, ROUGERIE and MEYERS, 1984e).

The 1981-83 time variations of monthly mean temperature profiles from 1.5°N to 1.5°S (Fig.23) have been contoured for three longitudinal zones by averaging all the XBT data collected between 140°E and 160°E (Western Pacific), 150°W-170°W (Central Pacific) and 95°W-105°W (Eastern Pacific). From these profiles, the preceding scenario from RASMUSSON and CARPENTER (1982) can be visualised for the 1982-83 El Nino, although it is the most deviant from the composite norm (CANE, 1985):

(a) The onset phase appeared in February-March 1982 when relaxation of easterlies happened in the vicinity of the date line in the Central Pacific. The heat content of the Western Pacific was large because of high sea surface temperature and a deep thermocline. The thermocline was also deep in the Central Pacific, whereas in the Eastern Pacific, the seasonal high sea surface temperature was observed.

(b) The peak phase took place in May 1982, at which time a Kelvin wave generated in the Central Pacific by relaxation of easterly winds arrived in the Eastern Pacific, inducing a deepening of the isotherms. At the same time, isotherms started to shoal in the Western Pacific, a phenomenon caused by the arrival of Rossby waves also generated in the Central Pacific by the easterly wind relaxation.
FIG. 23

(A) Time variation of the vertical distribution of temperature averaged at the Equator (1°5'N-1°5'S) in the Western Pacific (160°E) during 1981-83.

(B) As above except for Central Pacific (160°W).

(c) As above except for Eastern Pacific (100°W).

The El Nino phases are marked by thin vertical lines.
The transition phase occurred in December 1982, when a maximum anomaly of heat content due to the completion of the isotherms deepening was located in the Eastern Pacific. In the west, the shoaling of the isotherms was also close to completion. In the Central Pacific, heat content which had been at a maximum in July-August 1982 was decreasing as the isotherms shoaled.

The mature phase occurred probably during the first months of 1983. At this time, in the Eastern Pacific there was both a second deepening of the isotherms and a second maximum of the sea surface temperature. In the Western Pacific there was a second maximum shallowing of the isotherms. However, in the Central Pacific heat content continued to decrease in contrast with the other events. The 1982-83 El Nino was so active that the great heat content, located usually during the mature phase in the Central Pacific, shifted as far eastward as 140°W.

4.6 Post-El Nino conditions

Post-El Nino conditions are the prolongation of the mature phase over one year or more following El Nino. Usually there is a rapid recovery of normal conditions after El Nino, as in 1972-73. However, the best example of post-El Nino conditions having persisted for a long time was after the 1976-77 event.

In the post-El Nino period, the Walker Cell in the Pacific has usually shifted eastward, whereas the corresponding Cell in the Indian Ocean moves into the Western Pacific (Fig.10). In the Eastern Pacific, easterlies persist but in the Western Pacific, westerlies have replaced the trade winds. The descending branch of each Walker Cell characterised by a subsidence zone occurs at each extremity of the Pacific.

In the Western Pacific, post-El Nino conditions consist of either westerly or weak easterly winds in the equatorial zone, associated with a small SOI and with the presence of the intertropical convergence zone of the winds close to the equator (Fig.24). The convergence zone of the winds ceases its seasonal 15°S-15°N movement and stays close to the equator in the Western Pacific (DONGUY and HENIN, 1981b), in the vicinity of the ascending branch of the Walker Cell. The westerlies in the equatorial area have two main oceanographic consequences. They neither induce equatorial upwelling, nor do they pile up the water in the Western Pacific. On the other hand, the sea level falls by several tens of centimetres (WYRTKI, 1979b) and the thermocline rises by several tens of metres by baroclinic adjustment (DONGUY, HENIN, MORLIERE and REBERT, 1982). The thickness of the mixed layer has been reduced whereas the surface temperature anomaly is usually weak or even nonexistent, as shown by an equatorial transect obtained after the 1965 El Nino (Fig.25).

During the post-El Nino phase it rains in the equatorial zone of the Western Pacific owing to the presence of the convergence zone. The precipitation is combined with eastward advection and with the absence of upwelling results in the formation of low salinity water (DONGUY and HENIN 1976b, 1978b). This low salinity water is separated from the high salinity cell by
FIG. 24. Schematic representation of the wind field when the convergence zone of the winds is located in the vicinity of the equator.

FIG. 25. Vertical distribution of temperature at 170°E along a transect from 5°N to 20°S observed during the cruise BORA 1 (December 1965), during post-El Nino conditions.
FIG. 26. Vertical distribution of salinity at 170°E along a transect from 5°N to 20°S observed during the cruise BORA 1 (December 1965), during post-El Nino conditions.

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FIG. 28 Different zones of correlation between the intensity of El Nino and the ratio of the observed and mean rainfall (from DONGUY and HENIN, 1980b).

An intense pycnocline is observed during several cruises carried out by Centre ORSTOM de Nouméa: BORA 1 (November-December 1965) after the moderate 1965 El Nino (ROTSCHI, HISARD and JARRIGE, 1972) (Fig. 25, 26), EPONITE 2 (August-September 1976) after the moderate 1976 El Nino, for the Western Pacific (OUDOT, FERRER, HENIN, GARBE, DE GEOFFROY, JARRIGE, ROUGERIE, RUAL and SUPRIN, 1979). The equatorial zone, arid during the pre-El Nino period is particularly wet during the post-El Nino period.

Due to the presence of the convergence zone close to the equator during the post-El Nino period (DONGUY et al., 1982), in the tropical area north and south of the equator the Ekman pumping induces a positive upward displacement of the isotherms, similar to the vertical displacement at the Equator due to the baroclinic adjustment characteristic of the western equatorial Pacific.

The climatic scenario in the Western Tropical Pacific during post-El Nino period is different from that prevailing during the pre-El Nino period (Table 1) (DONGUY, 1982). During post-El Nino, anomalous rainfall on the equator due to the westerly winds or to the convergence zone of the wind replaces the aridity (Fig. 27), and surface salinity is low. However, easterlies blow outside the equatorial band (5°N-15°N, 5°S-15°S), and surface salinity becomes high due to the westward advection from Central Pacific, the lack of rainfall and the Ekman pumping bringing high salinity water up to the surface. The thermocline is shallow and consequently the heat content is small. There are then few tropical cyclones that originate in the Western tropical Pacific, instead they are formed in the Central Pacific where the heat content is large. As there is no convergence zone of the winds and no tropical cyclones,
drought occurs in the southwest Pacific where there is a negative correlation between the El Nino strength and the excess of precipitation. On both sides of the equator, this correlation is negative (DONGUY and HENIN, 1980b) (Fig.28) whereas it is positive within the equatorial band.

During the post-El Nino period, the flow of the eastward Countercurrents is stronger than during the pre-El Nino period (Fig.13) (DONGUY, ELDIN, MORLIERE and REBERT, 1984d). The South Equatorial Current flow is stronger in the post-El Nino than in the pre-El Nino, and the South Equatorial Countercurrent may extend as far eastward as Polynesia, whereas during the pre-El Nino it reaches only as far as 180° (DONGUY and HENIN, 1981b).

In the Central Pacific, post-El Nino conditions are characterised by the presence of the eastward-displaced ascending branch of the Walker cell (Fig.10). As in the Eastern Pacific the slope of the thermocline associated with the wind stress does not change, and in the Western Pacific the thermocline slope is reversed in agreement with wind stress, the thermocline is the deepest in the Central Pacific where the heat content and the ocean-atmosphere exchanges are maximum (HENIN and DONGUY, 1980) (Fig.22). This feature was obvious in 1957-58 after the strong 1957 El Nino, and also in 1976-77 after the moderate 1976 El Nino (Fig.29) according to data collected during the cruise DANAIDES 2 carried out by the Centre ORSTOM de Nouméa (ANON, 1980). This feature is also corroborated by the sea surface temperature in the Central Pacific, where a positive anomaly was noted in January 1977 (MIYAKODA and ROSATI, 1982). Moreover, empirical orthogonal analysis along the Fiji-Honolulu shipping route revealed a positive anomaly of sea surface temperature in 1976-77 (DONGUY and DESSIER, 1983). During the 1982-83 El Nino, the deepest thermocline occurred in the Central Pacific east of 150°W. At this location, maximum heat content and ocean-atmosphere exchanges were clearly evident when the tropical cyclones originated in the Central Pacific instead of in the Western Pacific (DONGUY, BEGAUD, EBSTEIN and CALVEZ, 1979; ELDIN and DONGUY, 1983; REVELL and GOULTER, 1986).

In the Eastern Pacific, post-El Nino conditions hardly differ from the pre-El Nino ones. Following the interesting 1976-77 episode, equatorial upwelling was normal (DONGUY and HENIN, 1980a) and empirical orthogonal analysis along the Tahiti-Panama shipping route during 1977 did not reveal positive anomalies of sea surface temperature (DONGUY and DESSIER, 1983). However, according to MIYAKODA and ROSATI (1982) in January 1977 the south-eastern Pacific was occupied by a positive anomaly of sea surface temperature.

In conclusion, the post-El Nino conditions are well developed in the Western and Central Pacific, but not in the Eastern Pacific.
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FIG. 29. (A) Mean temperature (0-100m) from November 1957 to March 1958 from HENIN and DONGUY (1960).

(B) Mean temperature (0-100m) from November 1976 to March 1977 (from HENIN and DONGUY, 1980).
5. CONCLUSION

Since 1979, the SURTROPAC Group of the Centre ORSTOM de Nouméa has contributed to the study of the coupled ocean-atmosphere system in the tropical Pacific using both hydrographic cruises and surface-XBT network by ships-of-opportunity. From the analysis of the observations, the following main statements can be made:

1. In the Western tropical Pacific even weak variations of the surface temperature above 25°C lead to strong variations in energy transfer through the sea surface.

2. In the Western tropical Pacific, sea surface salinity can be considered as a mirror of climatic conditions, both for seasonal and interannual variations connected with El Niño events.

3. The heat pool usually present in the Western equatorial Pacific shifts eastward during El Niño and consequently tropical cyclones originate in the Central Pacific.

4. Two types of climatic conditions occur during the pre-El Niño and post-El Niño periods in the tropical Pacific Ocean.

5. By consideration of the scenario of the 1982-83 El Niño, it seems that this event started in the Central Pacific instead of the Western Pacific.

On the other hand, the SURTROPAC network has shown the capacity to monitor not only the heat content available in the Pacific Ocean but also the surface circulation that disseminates it. This network will probably constitute one of the major data sources for the TOGA programme.

6. REFERENCES


J. R. DONGUY


