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Nitrate and chlorophyll distributions in relation to thermohaline and current structures in the western tropical Pacific during 1985-1989

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Abstract—Fourteen transects carried out from 1985 to 1989 permit us to describe the nitrate and chlorophyll *a* distributions and their temporal variability in the tropical western Pacific (165°E, 20°S–10°N). Sections representative of the moderate 1986–1987 El Niño, the strong 1988–1989 La Niña and an equatorial westerly wind burst were compared to the January 1986 transect, which we define as a “reference” section. Along 165°E, there appears to be four characteristic distributions of nitrate and chlorophyll relative to the thermohaline structure. (1) In the south (20–17°S), a seasonal cycle was observed in both nitrate and chlorophyll distributions. Variability of surface chlorophyll and depth of the deep chlorophyll maximum (DCM) was associated with variations in the vertical mixing and available amount of light. (2) In the north (6–10°N), seasonal and interannual changes remained in the subsurface layer because of the strong stratification. Ekman pumping is one cause of the changes of the DCM depth. (3) In the 2°S–2°N band, the seasonal variability of the DCM depth was associated with variations in precipitation and eastward advection of low salinity water. During the 1986–1987 El Niño, elevation and intensification of the pycnocline and shoaling of the nutrient reservoir were the result of basin-wide changes. Consequences included abnormally low surface chlorophyll concentrations and an increase of the primary production. Intra-El Niño variations of the DCM depth were associated with changes of the thickness of the low salinity surface layer. During the subsequent La Niña, upwelling developed and vertical nitrate and chlorophyll distributions were strongly modified. Surface nitrate and chlorophyll concentrations for this period were the highest of the dataset, and primary production values were about twice the El Niño values. The main differences between upwelling in the western and central Pacific are attributed to the existence of two atmospheric convergence zones in the western basin, and especially to the tendency of the Intertropical Convergence Zone to migrate south of the equator. Intraseasonal Kelvin waves influence the variations of nutrient concentration and phytoplankton biomass. During the December 1989 westerly wind burst, there were strong modifications of the thermohaline and nitrate distributions, associated with a geostrophic adjustment, which were not echoed on the chlorophyll distribution. (4) At 10°S, a doming of the different properties was frequently observed. Transient surface enrichments in nitrate or chlorophyll result from combined effects of the divergence between the South Equatorial Counter Current and the southern branch of the South Equatorial Current, Ekman pumping favourable to upwelling and local wind events. In all cases, nitrate and chlorophyll distributions were closely linked to the density structure for which salinity can be the controlling factor. Low-frequency variability (seasonal, intraseasonal and interannual scales) of the nitrate and chlorophyll distributions were determined primarily by local or remote physical processes which controlled stability of the water column that governs the vertical displacements of phytoplankton.

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INTRODUCTION

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The hydrological and biological properties of the central and especially the tropical eastern Pacific are often contrasted with the situation of the western tropical Pacific. Upwelling

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results in cooler surface waters along the coast and equator in the tropical eastern Pacific and higher biological productivity. The western tropical Pacific is known to have warm surface waters and low plankton biomass. El Niño Southern Oscillation (ENSO) events induce considerable alterations of the physical environment in the central and eastern Pacific. Numerous studies of ENSO have focused on the eastern Pacific because of the observed large sea-surface temperature (SST) anomalies (Hayes *et al.*, 1986) and because of the ecological and economic upheavals they cause along the South American coast (Barber and Chavez, 1983). In the early 1980s, Gill and Rasmusson (1983) described the relationship between warm SST areas and convection zones in the atmosphere. Since then, the western Pacific, which is the strongest heating source of the atmosphere, has received great attention from physical oceanographers and meteorologists. Understanding the mixed-layer dynamics and the ocean-atmosphere interactions in this region is a key to the understanding of the early stages of ENSO (Enfield, 1989). These studies contributed a better description of the mixed-layer structure (Lukas and Lindstrom, 1991) and the annual and interannual changes of the physical environment (Toole *et al.*, 1988; Delcroix and Henin, 1989; Delcroix *et al.*, 1992; McPhaden *et al.*, 1992).

Nutrient and biological patterns in the western Pacific are generally well known, but their response to variation of physical forcing is still the subject of much debate. Maps of primary production produced by Fleming (1957) (see Fig. 2 in Berger, 1989) and Koblentz-Mishke *et al.* (1970) showed an oligotrophic (surface layer with no nitrate and a low chlorophyll content) tropical western Pacific with a cessation of equatorial biological enrichment around 170°E. More recently, Berger (1989) suggested that the western Pacific was less oligotrophic, with areas of very low production restricted to the subtropical gyres and the border of the clearly defined equatorial enrichment zone, which extends as far west as 155°E. These maps actually represent average situations, and a permanent oligotrophic state in the western Pacific is debatable as it can be altered in response to physical forcing, notably during ENSO. A series of meridional transects along 170°E undertaken by ORSTOM-Nouméa in 1965–1968 and 1970–1975 showed that upwelling resulting in high surface nitrate and increased phytoplankton biomass was one of the situations encountered in the region (Rotschi, 1973; Oudot and Wauthy, 1976). Recently, Blanchot *et al.* (1992) described significant changes in nutrient structure, plankton biomass and productivity at 165°E during two extreme situations of the ENSO cycle in 1987 and 1988. “A periodic cessation of oligotrophy” (Dandonneau, 1988) was reported by Dandonneau (1992) using large-scale data of sea-surface chlorophyll concentration (SSCC) obtained through a ship-of-opportunity programme. Particularly during the 1982–1983 ENSO warm event, the SSCC decreased in the central and eastern Pacific while it increased in the west. Considering the absence of a well-defined relationship between biological variability in the eastern Pacific coastal upwelling and local physical forcing, Barber and Chavez (1983) proposed that the observed biological signal was the result of both local and remote physical processes. This concept was more thoroughly described in Barber (1988, 1990), where the basin-wide modification of the east–west nutricline and thermocline slope was identified as an important process to explain the large-scale variation in primary production observed in the eastern and western Pacific during the 1986–1987 ENSO event. Despite these new insights, the variability of the nutrient environment and the associated biological responses in the western Pacific are still insufficiently described in the literature.

The purpose of this paper is to summarize a large number of physical, nutrient and phytoplankton observations in the western tropical Pacific over a fully developed ENSO

event. This is the first time that such an investigation has been carried out for this region. These data were collected during 14 transects along 165°E, from 20°S to 10°N, from January 1985 to December 1989. This period includes the moderate 1986–1987 El Niño and the following cold phase in 1988. Recently, Delcroix *et al.* (1992) produced a thorough description of the variability in physical parameters in the western Pacific during the 1986–1988 ENSO. Adding nutrient and chlorophyll data to their physical framework enables us to determine if variations in physical parameters, observed on seasonal or interannual timescales, are reflected in nutrient and/or biological structures. During the 1985–1989 period, we also investigated more northern (to 10°N) and southern (to 20°S) regions where nitrate and chlorophyll distributions had previously been poorly documented.

DATA AND METHODS

The 14 cruises presented in Table 1 were undertaken in the western Pacific by the Institut Français de Recherche Scientifique pour le Développement en Coopération (ORSTOM) and constitute the central dataset used in this study. Since 1984, the SURTROPAC (Surveillance Trans-Océanique du Pacifique) group has carried out a series of semi-annual (summer and winter) expeditions along 165°E, mainly from 20°S to 10°N. Standard hydrographic and horizontal current velocity profiles were collected at nearly every degree of latitude. In January 1985, as part of the PROPPAC (Production Pélagique dans le Pacifique) programme, nutrients, chlorophyll and zooplankton biomass measurements were added to the physical observations. Similar transects run by the PROPPAC group in September and April, supplement the SURTROPAC series in 1987–1988. When necessary, additional observations, such as temperature from expendable bathythermographs (XBT) and moorings, were used.

Table 1. Cruises along 165°E in 1984–1989. First leg of SURTROPAC 13 is along 169°E (SU and PR stand, respectively, for SURTROPAC and PROPPAC cruises). Chlorophyll data from the first six cruises are deduced from the Dandonneau method (see Data and Methods)

Cruise	Date	Physical data	Nitrate data	Chlorophyll data
SU 03	9–21 January 1985	20°S–10°N	20°S–10°N	20°S–10°N
SU 04	28 June–9 July 1985	20°S–17°S, 14°S–8°N	20°S–17°S, 14°S–8°N	20°S–17°S, 14°S–8°N
SU 05	10–26 January 1986	20°S–10°N	20°S–10°N	20°S–10°N
SU 06	17–27 June 1986	20°S–10°N	20°S–10°N	20°S–19°S, 17°S–7°S, 5°S–10°N
SU 07	10–27 January 1987	20°S–7°N, 9°N–10°N	20°S–7°N, 10°N	no data
SU 08	2–15 July 1987	20°S–3°S, 1°S–10°N	19°S–3°S, 1°S–10°N	19°S–3°S, 1°S–10°N
PR 01	9–20 September 1987	20°S–6°N	20°S–6°N	20°S–6°N
SU 09	16–28 January 1988	20°S–10°N	20°S–10°N	20°S–10°N
PR 02	28 March–8 April 1988	20°S–6°N	20°S–6°N	20°S–6°N
SU 10	14–27 June 1988	20°S–10°N	20°S–10°N	20°S–10°N
PR 03	11–22 September 1988	20°S, 18°S, 14°S–6°N	19°S–18°S, 15°S–6°N	19°S–18°S, 15°S–6°N
SU 11	6–19 January 1989	20°S–10°N	19°S–12°S, 10°S–10°N	19°S–12°S, 10°S–10°N
SU 12	29 June–12 July 1989	20°S–10°N	20°S–10°N	20°S–10°N
	17–23 July 1989	5°N, 2°N–10°S	2°N–11°S	0°, 2°S, 8°S
SU 13	1–10 December 1989	20°S, 18°S–9°S, 3°S–4°N	10°S–9°S, 3°S–4°N	10°S–9°S, 3°S–4°N
	15–26 December 1989	8°N–16°S	8°N–16°S	8°N–16°S

Physical parameters

Temperature and salinity profiles were acquired with rosette-mounted STD (salinity–temperature–depth) or CTD (conductivity–temperature–depth) probes. The raw data were binned and averaged every 5 dbar from 0 to 1000 dbar. Almost simultaneous zonal velocity profiles were deduced from measurements of a profiling current meter system to 600 m. More information about measurement procedures and data processing can be found in previous papers (Delcroix *et al.*, 1987, 1992).

The XBT ship-of-opportunity programme (White *et al.*, 1985) made possible the monitoring of the upper layer temperature with better time resolution than from the cruises. Casts along the Japan–New Caledonia track were used to compute several isotherm depths in the 20–17°S and 6–10°N regions from 155°E to 165°E. Details about this programme run by ORSTOM, Nouméa, can be found in Delcroix and Masia (1989).

Daily temperature data from the TOGA–Tropical Atmosphere Ocean (TAO) equatorial mooring at 165°E were used for the study of the 1988–1989 La Niña.

Nutrients and chlorophyll

Seawater samples were collected from 12 sampling depths from 0 to 200 m. Nitrate + nitrite ($\text{NO}_3^- + \text{NO}_2^-$) were analyzed immediately with a Technicon Auto-Analyzer II using the method described by Strickland and Parsons (1972). The limit of detection was initially 0.050 μM but was reduced to 0.010 μM in June 1988, when we started to use a more sensitive method developed by Oudot and Montel (1988). For the sake of convenience, nitrate + nitrite ($\text{NO}_3^- + \text{NO}_2^-$) will be termed nitrate (NO_3) in this paper.

For most cruises, chlorophyll measurements were made in the laboratory. Before July 1987, the samples were analyzed according to a non-extractive method developed by Dandonneau (1982). Twenty centimetres of seawater were filtered through membrane filters HA type (Millipore), 13 mm in diameter. The resulting values were related to the sea chlorophyll concentration by an empirical linear relation after calibration with classical fluorometric methods (Dandonneau, 1988). After discussion of bias and incertitude, Dandonneau (1988) stated that the results must be considered only as an index of the chlorophyll concentration. They will be used to show the general features and to retrieve the depth of the chlorophyll maximum. Since July 1987, chlorophyll *a* has been extracted and measured using a fluorometric technique proposed by Herbland *et al.* (1985) and slightly modified by Le Bouteiller *et al.* (1992). Samples (100 ml) were filtered onto 25-mm Whatman GF/F filters, and chlorophyll *a* was extracted in methanol (95%) within 3 h. Replicate analysis gave coefficients of variation ranging from 2 to 5%.

Criteria for homogeneous layers depth

The top of the thermocline and mixed-layer depths were determined using vertical gradient criteria ($0.05^\circ\text{C m}^{-1}$ for temperature and 0.01 kg m^{-4} for density) as described in Lukas and Lindstrom (1991) and formerly tested along 165°E (Delcroix *et al.*, 1992). South of 10°S, vertical density gradients of the order of 0.1 kg m^{-4} that existed in the mixed layer on a small vertical scale (about 5 m) have been rejected. These small-scale vertical structures were especially present in winter. Instead of a gradient criteria to define the nitracline, 0.1 μM was used to determine the threshold of the nitrate-rich and biologically active layer

(Herbland *et al.*, 1985; Le Bouteiller *et al.*, 1992). It represents the lower limit of the layer where nitrate is assumed to be limiting. Of course, the limited number of sampling bottles resulted in a coarse vertical resolution that sometimes induced uncertainty in determination of the nitracline depth. For the same reason, the chlorophyll maximum, which is determined from direct measurement, was sometimes undersampled.

TYPICAL OBSERVED FEATURES

Southern Oscillation Index and winds

The Southern Oscillation Index (SOI) is an index of the basin-wide oceanic and atmospheric variability. Three main phases emerge from its fluctuations during 1985–1989 (Delcroix *et al.*, 1992) (Fig. 1). We define the reference period as the 1.5 years from 1985 to mid-1986 when SOI fluctuations were minimal and the mean value was close to zero. SOI decreased to a minimum in May 1987 during the El Niño year (mid-1986–end of 1987), then rose to a maximum in September 1988 during the La Niña episode (1988–1989). By the end of 1989, SOI was back to zero.

The 1985–1989 time-series of the zonal pseudo-wind stress issued by Florida State University (Fig. 2) was described in Delcroix *et al.* (1992). The data contain both seasonal and interannual variability. During the reference period, the trade winds strengthened in the winter hemisphere due to the seasonal meridional shifts of the Intertropical Convergence Zone (ITCZ) and South Pacific Convergence Zone (SPCZ). In the early months of the year, the ITCZ may extend to the southern hemisphere, and northeasterly winds are common in the equatorial band. The SPCZ activity is stronger during the austral summer when southeast trade winds give way to moderate westerlies south of the equator to 15°S. Statistical studies (Harrison and Giese, 1991) show that in the equatorial band, westerly wind events are more frequent from September to February. Zonal winds remained remarkably weak along the equator until September 1986. In July 1986, when the SOI fell below zero, the emergence of westerlies north of the equator announced a period of

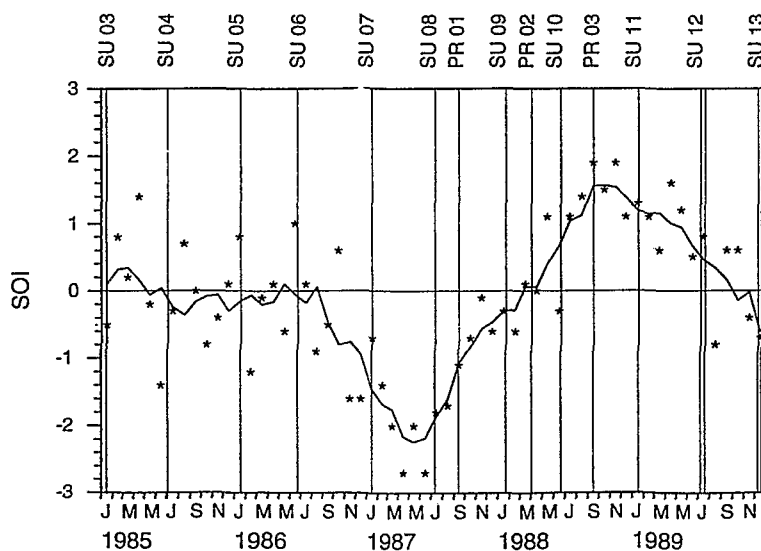


Fig. 1. Southern Oscillation Index for 1985–1989 (courtesy of the Climate Analysis Center, Washington, DC, U.S.A.). Vertical lines denote research cruises (SU for SURTROPAC, PR for PROPPAC) discussed in this report.

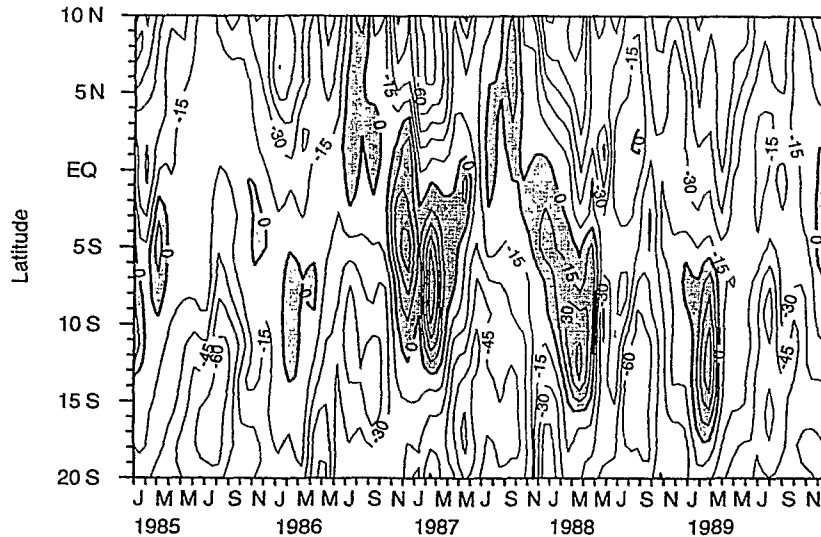


Fig. 2. Zonal pseudo-wind stress averaged from 163°E to 167°E in 1985–1989. Contour interval is $15 \text{ m}^2 \text{ s}^{-2}$; positive values of westerly winds are shaded.

enhancement of the seasonal cycle (Delcroix *et al.*, 1992). Strong westerly wind bursts occurred from 5°N to 10°S at the end of 1986, preceding another period of persistent positive zonal anomalies extending from the equator to 12°S that lasted until March 1987. The features in the following year were similar, although somewhat weaker. Beginning in April 1988, when the SOI became positive, particularly intense easterlies resumed and persisted until the end of 1989 when westerly winds occurred at the equator. From mid-November 1989 to the end of the year, the wind was mostly westerly, with two 10-day long westerly wind events from 20 November to 1 December and from 9 to 21 December. During these wind events, the daily zonal wind ranged from 6 to 8 m s^{-1} (McPhaden *et al.*, 1992).

Meridional distributions of temperature, salinity, zonal current, nitrate and chlorophyll

The general hydrographic features and circulation in the tropical western Pacific already have been described extensively (e.g. Toole *et al.*, 1988; Delcroix *et al.*, 1992), while nutrient and chlorophyll distributions have been only partially examined (Rotschi, 1973; Oudot and Wauthy, 1976; Barber and Kogelschatz, 1989). In this section, we first describe the physical, nutrient and chlorophyll distributions in January 1986 (SURTROPAC 05) during the reference period. It has been chosen as representative of the more usual structures of these properties in this region. Then, the changes that occurred from 1986 to 1989 will be illustrated using the meridional transects of three other ORSTOM cruises: SURTROPAC 08 in July 1987 during the 1986–1987 El Niño; PROPPAC 02 in April 1988 during the 1988–1989 La Niña; and SURTROPAC 13 in December 1989 after the westerly wind burst.

Reference conditions (January 1986). The situation considered “normal” for the western Pacific is shown in Fig. 3: the thermocline, pycnocline and nitracline are deep, and the surface layer is warm, nitrate-depleted ($[\text{NO}_3] < 0.1 \mu\text{M}$) with a low chlorophyll content. The absence of nutrient enrichment or cold water at the surface at the equator is consistent with the low zonal wind conditions and contrasts with the commonly described equatorial upwelling in the central and eastern equatorial Pacific.

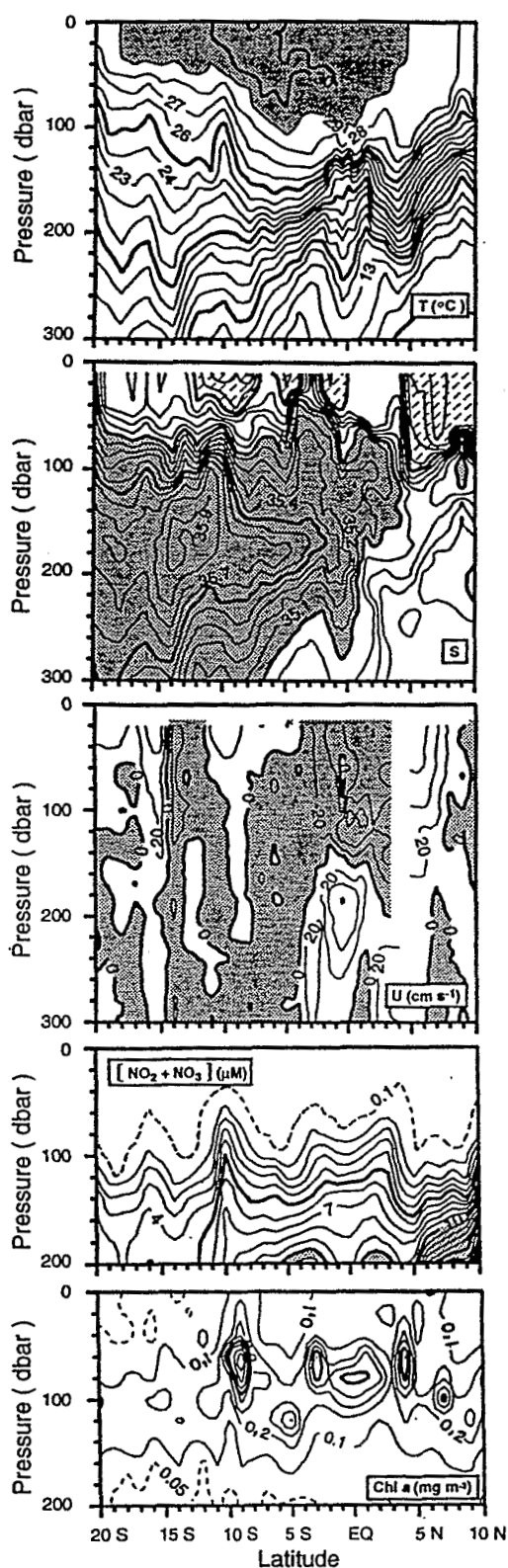


Fig. 3. Meridional sections from 20°S to 10°N at 165°E in January 1986 (SURTROPAC 05). From top to bottom: temperature (contour interval is 1°C; temperatures greater than 29°C are shaded), salinity (contour interval is 0.1; salinities greater than 35.0 are shaded, salinities lower than 34.5 are hatched), measured zonal current (relative to 600 dbar, contour interval is 20 cm s⁻¹, and westward flow is shaded), [NO₃⁻ + NO₂⁻] (contour interval is 1 μM, dashed line is the 0.1 μM detection limit; concentrations greater than 10 μM are shaded), chlorophyll *a* (contour interval is 0.1 mg m⁻³, dashed line is the 0.05 mg m⁻³ isoline). Note that there are no data in the 4–5°N band.

The major structures of the equatorial current system were encountered along 165°E. At the surface, the southern limit of the westward-flowing North Equatorial Current (NEC) was located around 9°N, the eastward-flowing North Equatorial Counter Current (NECC) extended from 9°N to 4°N, and the westward flowing South Equatorial Current (SEC) lay from 4°N to 15°S. The SEC was split into an equatorial branch and a southern branch by the South Equatorial Counter Current (SECC) that flowed eastward in the surface layer from 5°S to 11°S. The eastward current, expanding from 15°S to at least 20°S, is part of the southern branch of the large-scale anticyclonic gyre (Wyrтки, 1975; Levitus, 1982; Kessler and Taft, 1987). At depth, the eastward Equatorial Undercurrent (EUC) was centred between 150 and 200 m at the equator, and the north and south subsurface counter currents were located at 3°N and 3°S, respectively.

The temperature section shows that the well-known warm pool ($T > 29^{\circ}\text{C}$) covered the 15°S–5°N region with a thick isothermal layer (80 m in the central part). A shallower isohaline layer is often observed in the western Pacific (Delcroix *et al.*, 1992; Sprintall and McPhaden, 1994), and the nearly isothermal and salinity stratified region between the top of the thermocline and the base of the mixed layer has been defined as the “barrier layer” by Lukas and Lindstrom (1991). The vertical structure in salinity controls the mixed-layer depth, which is shallower than the isothermal layer almost all the way from 10°S to 5°N. Low-salinity surface waters ($S < 34.5$) north of 5°N, and between 5°S and 10°S, are associated with areas of maximum annual precipitation under the Intertropical and South Pacific Convergence Zones (Dorman and Bourke, 1979; Delcroix and Hénin, 1991). Intermediate salinity waters, which present a moderate maximum, are usually situated a few degrees south of the equator. The January 1986 cruise coincided with a period of maximum southern displacement of the warm, fresh surface water. Indeed, its extension toward the south is maximum from January to April and minimum from July to October as the consequence of two processes: the seasonal variation of the solar incident radiation (Tournier, 1989; Delcroix and Hénin, 1989) and the meridional displacement of salinity minima associated with rain fall under ITCZ and SPCZ (Delcroix and Hénin, 1991).

The general thermohaline characteristics of subsurface waters were similar to those described in the central Pacific by Wyrтки and Kilonsky (1984) and Feely *et al.* (1987). The core of high-salinity water (up to 35.9) of subtropical origin (Rotschi, 1973) spreads from the south toward the equator in the upper part of the weak thermocline and is associated with low nutrient concentrations. In the south of the transect, the nutrient reservoir resides below our deepest sampling depth ($[\text{NO}_3] < 5 \mu\text{M}$ at 200 m), in the lower thermocline, as previously observed both in the central (Wyrтки and Kilonsky, 1984, Figs 2 and 5) and western (Rotschi *et al.*, 1972, Fig. 7a–m) Pacific. The vertical nitrate gradient, just as the temperature gradient, was the weakest of the transects, and the nitracline was located much deeper than the mixed layer. In contrast, north of 4°N, the thermocline shoaled and was sharper. This is where the highest nitrate concentrations (15–21 μM at 200 m) and strongest vertical gradients were observed. This nutrient-rich water is associated with two water masses: below 120 m, water from the eastern tropical North Pacific (ETNP) (Wauthy, 1986) overlays the salinity intermediate waters of the north Pacific (Toole *et al.*, 1988; McPhaden *et al.*, 1992). ETNP waters compose the southern edge of the NEC and partly supply the NECC (Wauthy, 1986). Their equatorward extension was indicated by a meridional nitrate gradient around 5°N. Between these two regions, from 9°S to 5°N, mixing of southern hemisphere subtropical water, northern hemisphere water and water advected by the SEC contributed to the formation of waters of intermediate nitrate content (7–11 μM at 200 m).

The encounter of southern and northern hemisphere waters is depicted in the salinity section by a region of meridional salinity gradient and complicated interleaving structures between 3°S and 4°N. The equatorial spreading of nitrate isolines below 160 m is consistent with the geostrophic signature of the EUC. Feely *et al.* (1987) also attributed the relatively low nitrate concentrations (7–9 μM) of the core of the EUC to eastward advection of low nutrient water. The chlorophyll section showed a subsurface maximum all along the transect situated in the 60–120 m depth range. Its location in the vicinity of the nitracline is an essential feature of the Typical Tropical Structure (TTS) originally described by Herbland and Voituriez (1979) in the Atlantic.

The vertical distributions of thermohaline and biological parameters are strongly influenced by the zonal circulation (Merle *et al.*, 1969). The spreading of isotherms below 160 m at the equator is associated with the EUC. The convergent flow within the EUC is depicted by the meridional salinity front, which is more distinct between 100 and 200 m than above. Below the core of the EUC, the depression of the nitrate isolines suggests that the nitrate distributions are controlled by downwelling. The equatorial branch of the SEC is flanked (about 5°S and 5°N at the time of observation) by convergences that also depress the nitrate isolines. Most important, because of their potential biological consequences, are the divergences. Although much weaker than in the central Pacific, divergence is present at the equator, but upward advection is confined in a thin surface layer (Wyrтки, 1981; Philander *et al.*, 1987; Barber and Kogelschatz, 1989) and entrainment is insufficient to mix cold, nitrate-rich water from the thermocline into the surface layer (Meyers *et al.*, 1986; McPhaden *et al.*, 1990; Sprintall and McPhaden, 1994). This explains the absence of surface indications of equatorial upwelling in temperature and salinity. A weak upwelling was noted at 165°E only during the first half of 1984 (Delcroix *et al.*, 1987). For the same reasons, most of the time, no nitrate is reported at the surface in the western Pacific (Fig. 6). Interestingly and in contrast with the other parameters, the influence of equatorial upwelling was seen in the chlorophyll distribution, which had a slight surface chlorophyll enrichment from 3°S to 7°N (Fig. 3). Such an enrichment also was observed by Barber and Kogelschatz (1989), although neither temperature nor nitrate indicated upwelling. Surface chlorophyll enrichment also was reported around 10°S where nitrate rose to within 50 m of the surface. Temperature and salinity indicate that doming occurred between 50 and 180 m, and that its upper limit matched the bottom of the lower salinity surface layer. This feature, already identified by Merle *et al.* (1969) and Oudot and Wauthy (1976), will be discussed more fully in a later section.

El Niño conditions (July 1987). The major current modification during El Niño is the reversal of the SEC in response to strong westerly winds and equatorial Rossby waves (Delcroix *et al.*, 1992). In July 1987 (Fig. 4), during the mature phase of the 1986–1987 El Niño, the eastward surface flow was firmly established over a broad latitude band (5°S–6°N), whereas, away from the equator, counter current transports diminished. In the equatorial zone, significant surface features also included a slight cooling and a decrease in salinity. The sea-surface temperature (SST) cold anomaly, also observed during the 1982–1983 El Niño event, may have been the effect of evaporation (Meyers *et al.*, 1986) while sea-surface salinity (SSS) fluctuations were probably related to changes in precipitation and eastward zonal advection (Donguy and Hénin, 1976; Delcroix and Hénin, 1991). Surface nitrate remained undetectable, and chlorophyll concentrations were low (of the order of 0.1 mg m^{-3}). Even though there was interannual variability in SST and SSS in the equatorial region, there was no El Niño signal in surface nitrate (Fig. 6). At 10°S, surface chlorophyll concentrations greater

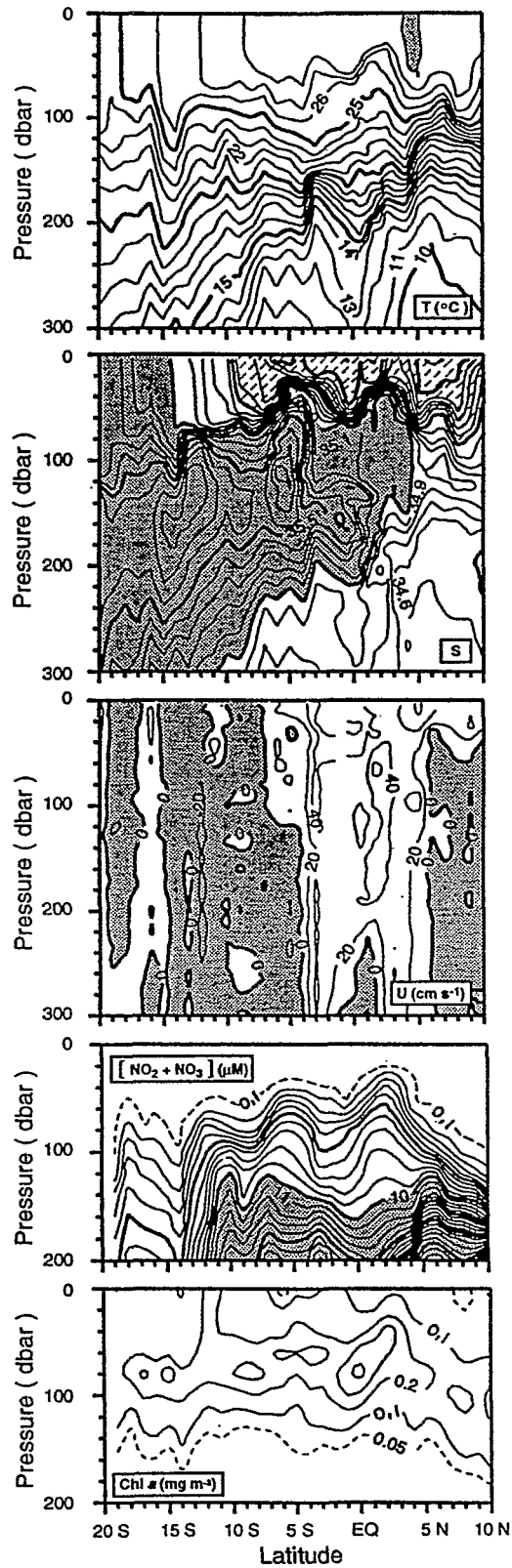


Fig. 4. As in Fig. 3, for July 1987 (SURTROPAC 08).

than 0.1 mg m^{-3} were reported for the two cruises during the 1986–1987 El Niño, however, surface nutrient enrichment appeared only in September 1987.

Remarkable changes in the thermohaline structure also were observed. A shallow thermocline, consistent with rising of the warmer isotherms of the thermocline during the first months of El Niño (Delcroix *et al.*, 1992), were characteristic of this period. The isothermal layer, which was 50–70 m thick between 10°S and at least 10°N , coincided with the low-salinity surface water layer. Variations in the nutrient distribution approximately followed the changes in the thermohaline structure. From 10°S to 10°N , the nutrient isolines shoaled by 20–40 m relative to distributions in the reference period. This process was initiated in mid-1986 (SURTROPAC 06). The $0.1 \mu\text{M}$ nitrate isoline was only 20 m deep from 1°N to 3°N , and nitrate concentrations at 200 m were the highest recorded during the 1984–1989 period ($15 \mu\text{M}$ from 7°S to the equator and more than $25 \mu\text{M}$ in the northern part of the section). The equatorial depression of the nitrate isolines reflected the surface convergent flow, already evidenced by other parameters. The high meridional nutrient gradient south of 10°S was the result of the shoaling of the nutrient reservoir. All along the transect, the depth (40–120 m) of the deep chlorophyll maximum (DCM) was 0–20 m above the nitracline depth.

La Niña conditions (April 1988). The distributions of the properties observed in the equatorial region in April 1988 (Fig. 5) were remarkably different to the two preceding transects and gave evidence of the great diversity of situations encountered in this part of the western Pacific. Strong SEC and EUC were observed and, as a consequence of intense trade winds in the western Pacific (Fig. 2), equatorial upwelling developed as far west as 165°E . The ensuing features were significant SST cooling and increased SSS, as well as surface nutrient and chlorophyll concentrations. The April 1988 cruise occurred during the early stage of the La Niña period, shortly after the initiation of upwelling (McPhaden *et al.*, 1990). Chosen for pragmatic reasons because of the completeness of the measurements, it shows the main characteristics of the upwelling at 165°E during the 1988–1989 La Niña. Upwelling diminished and finally disappeared in mid-1989 when trade winds eased. As pointed out by Donguy and Dessier (1983), surface temperature is not as reliable an indicator of upwelling in this region as surface salinity or nitrate. The nutrient enrichment zone (evidenced by the $0.1 \mu\text{M}$ nitrate surface isopleth in Fig. 6) reached $2\text{--}5^{\circ}$ poleward of the equator, and its limits matched the northern and southern limits of the equatorial branch of the SEC. The greatest northern extension of nutrients was observed at the beginning of the year and it extended southward in the middle of the year. The maximum equatorial nitrate concentration was $3.23 \mu\text{M}$ in March–April 1988, while surface chlorophyll exceeded 0.35 mg m^{-3} in September 1988. Upwelling affected the vertical structures of all the described parameters. The top of the thermocline, the base of the mixed layer and the top of the vertical nitrate gradient matched as a result of the development of a quasi-homogeneous layer. The vertical distribution of chlorophyll was strongly modified, with chlorophyll at the surface and a smooth chlorophyll maximum, which was sometimes absent. Poleward of the upwelling region, no major changes were observed, except for the temperature and nitrate isolines which deepened at the edges of the upwelling. By the end of 1989, the general features were comparable to the structures encountered during the reference years.

Westerly wind burst conditions (December 1989). Philander *et al.* (1987) emphasized the complexity of the vertical structures and temporal variability in this region where westerly

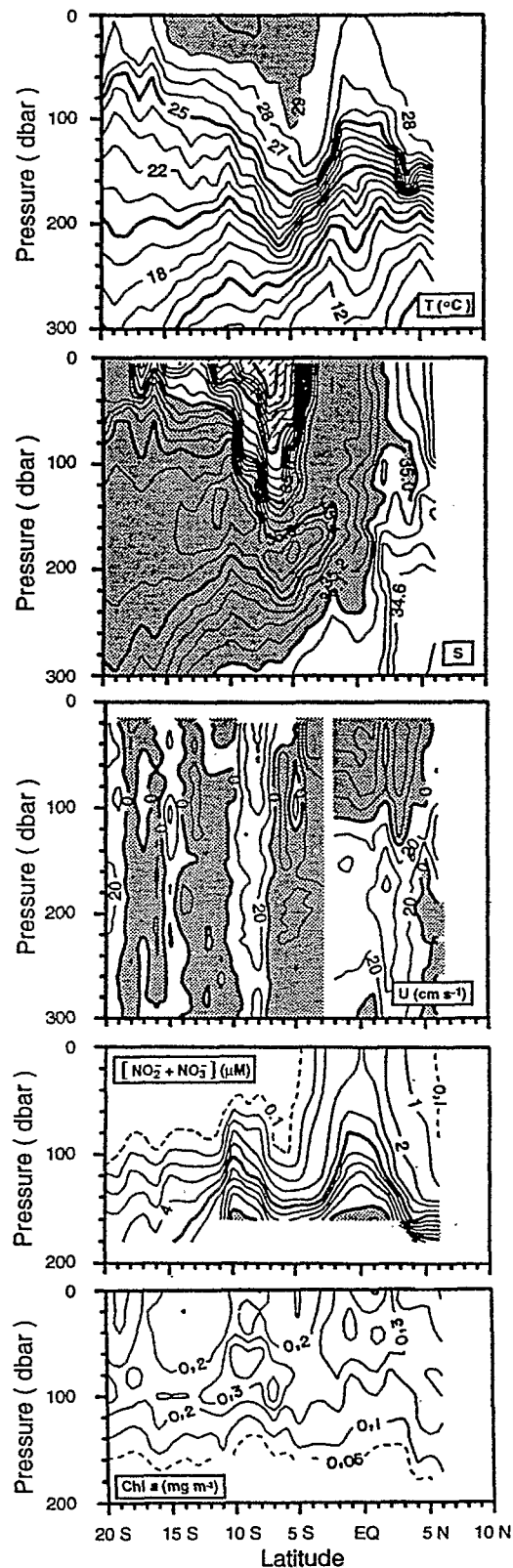


Fig. 5. As in Fig. 3, for March–April 1988 (PROPPAC 02). Note that there are no data in the 2–3°S band.

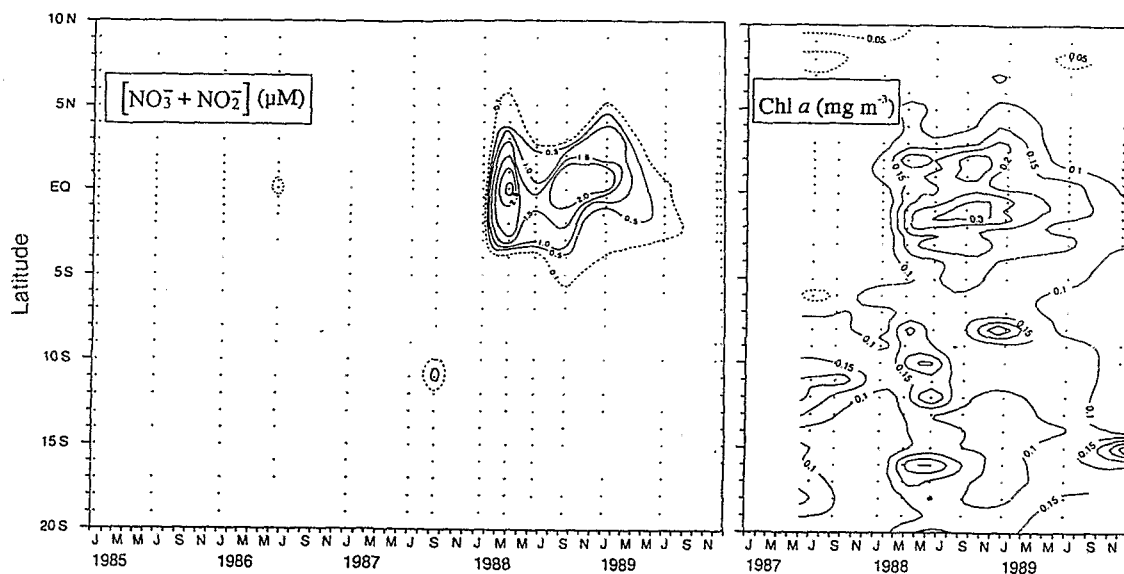


Fig. 6. Sea-surface concentrations (dots represent positions of stations). Left panel: $[\text{NO}_3^- + \text{NO}_2^-]$ (contour interval is $0.5 \mu\text{M}$, dashed line is the $0.1 \mu\text{M}$ detection limit). Right panel: chlorophyll *a* (contour interval is 0.05 mg m^{-3} , dashed line is the 0.05 mg m^{-3} isoline).

and easterly winds meet seasonally along the equator. On timescales of less than 1 month, surface currents may reverse following the reversal of the zonal component of the wind. Hisard *et al.* (1970) and McPhaden *et al.* (1990) related rapid reversals of the equatorial branch of the SEC to episodes of westerly wind. In this dataset, several cruises (January 1985 and 1987, December 1989) took place during, or shortly after, westerly wind episodes. The December 1989 cruise occurred during a period of successive westerly wind events in November 1989–January 1990. The response of the ocean to the December 1989 wind burst is exemplified in Fig. 7. A striking change in the upper layer was a 2° -broad eastward jet from the surface to 150 m overlying a westward counter flow and the undercurrent (McPhaden *et al.*, 1992). It resulted in deepening of the isothermal layer to 150 m and sharpening of the top of the thermocline. The SST was not strongly affected by the wind burst, but SSS decreased significantly due to meridional advection of fresher water from the north (McPhaden *et al.*, 1992), which created a shallow halocline and confined the mixed layer to 40–50 m in the vicinity of the equator. Surface salinity lower than 34.5 extended from the north to 1°S . The salinity front observed around 3°N in November 1989 was displaced to 1°S in December 1989 (McPhaden *et al.*, 1992). Surface chlorophyll concentrations over 0.10 mg m^{-3} were observed from $0^\circ30'\text{S}$ to $2^\circ30'\text{S}$, in the possible meridional divergence between the eastward-flowing equatorial current and the westward flowing SEC. The nitracline deepened considerably and roughly matched the top of the thermocline in a narrow band of latitude (1°S – 0°). Similarity of temperature and nitrate distributions supports the argument that nitrate was also subordinate to geostrophic adjustment. Interestingly, the chlorophyll distributions were very different, and it is the only example in this dataset of a DCM lying within the nitrate depleted layer (about 50 m above the nitracline). For the timescale under consideration (the westerly burst began about 10 days earlier; McPhaden *et al.*, 1992), no striking deepening of the chlorophyll structure was observed, although a slight depression of the isolines occurred below 100 m in the vicinity of the equator. The DCM (80–100 m) persisted in a layer where light appeared to be the dominant factor for production and where nitrate was seemingly unavailable. Siegel *et al.*

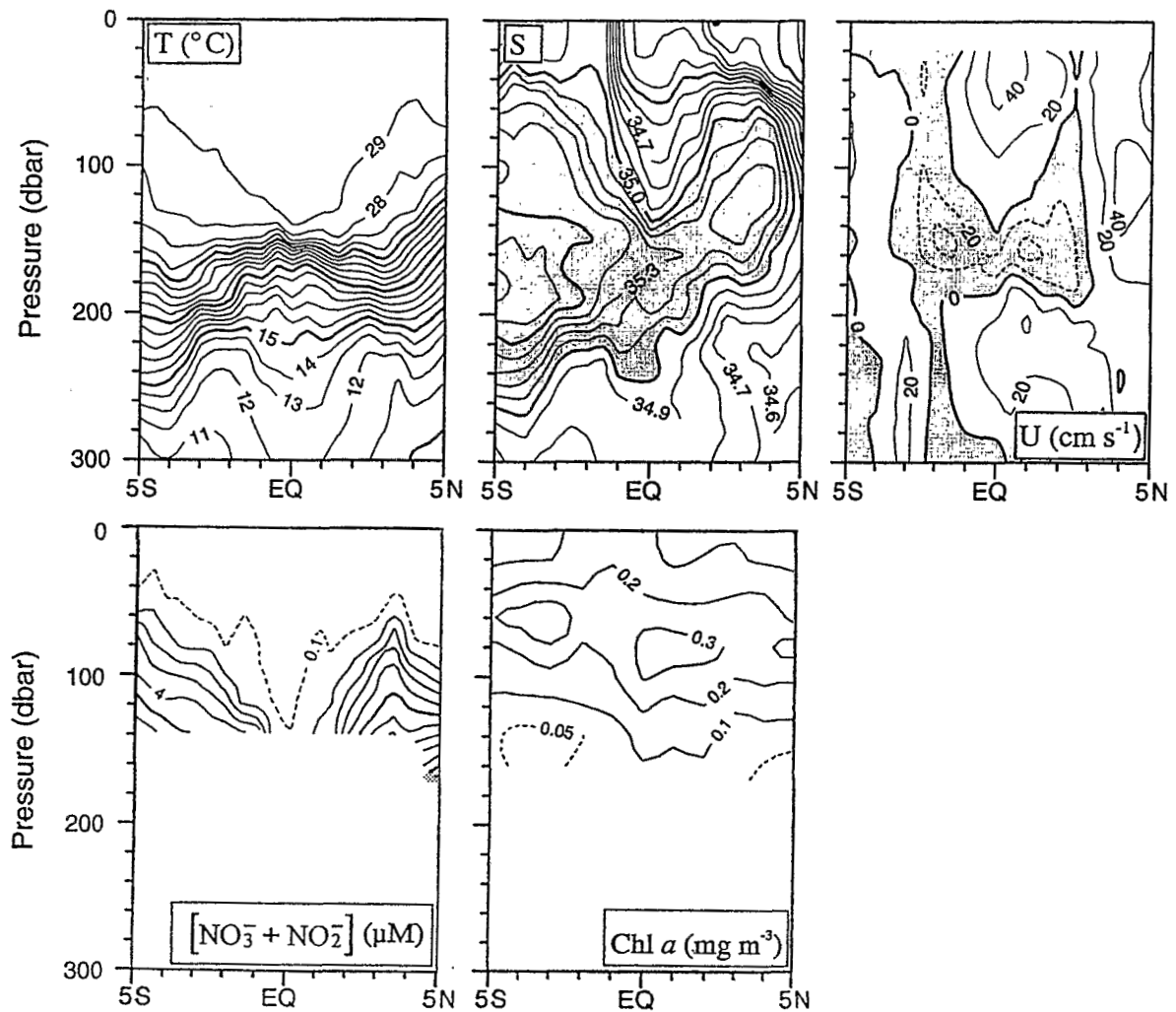


Fig. 7. Meridional sections from 5°S to 5°N at 165°E in December 1989 (SURTROPAC 13). Top panel: temperature, salinity, measured zonal current. Bottom panel: $[\text{NO}_3^- + \text{NO}_2^-]$ chlorophyll *a*. (Contour intervals are the same as in Fig 3.)

(1995) reported similar results after a westerly wind burst in December 1992–January 1993 south of the equator at 156°E. They argued that nutrients were entrained into the mixed layer during the wind event and that a phytoplankton increase occurred at cessation of the wind event. During the December 1989 wind event, it is highly probable that entrainment was reduced because of a barrier layer that tended to reduce the efficiency of vertical mixing in insulating thermocline and nutricline from the surface (McPhaden *et al.*, 1992). Non-negligible meridional advection (equatorward velocity of 10 cm s^{-1} estimated by McPhaden *et al.*, 1992) of water may have been instrumental in maintaining the DCM. From this dataset, we can draw no conclusion for other sources of nitrogen.

DISCUSSION

The south and the north of the transect: two oligotrophic states

The southern and northern regions of the transect are both oligotrophic; however, they have presented different thermohaline features. Reasons for oligotrophy in the south are the deep nutricline, while in the north the nitrate fluxes toward the euphotic zone are reduced. In

the south, the depth of the DCM and the concentration of the sea-surface chlorophyll vary seasonally. In the north, the DCM depth is subject to both annual and interannual variability.

The 20–17°S region. During 1985–1989, stratification in the upper 200 m was stronger (0.04 kg m^{-4} around 50 m) in austral summer coincident with surface layer warming and with the southward incursion of fresher surface waters. Seasonal variability of the vertical density structure was clearly present during the reference years (1984–mid-1986) when the base of the mixed layer was located around 40 m in austral summer and deepened to 60 m in austral winter (Fig. 8). The annual cycle of the DCM depth (about 110 m in summer and 80 m in winter) tracked the nitracline depth (less than 10 m apart) and was in opposite phase to the mixed-layer depth. As a result, a large gap between the base of the mixed layer and the DCM depth existed in summer. Furthermore, variations of the DCM depth were poorly related to variations in the upper thermocline, represented here by the 24°C isotherm depth (deduced from XBT data). The seasonal variation of the DCM depth is due to variations in the available amount of light and vertical mixing. The DCM is deeper in the summer because the amount of light is greater than in winter. In addition, vertical mixing in the summer is limited to a superficial layer and has no influence on the nutrient input toward the surface. By contrast, in winter, owing to weaker stratification and stronger winds, vertical mixing is stronger and deeper and is responsible for nutrient supply toward the euphotic zone. Phytoplankton then find adequate nutrient and light conditions at a shallower depth. A simple calculation of the attenuation of downward solar radiation with depth [Paulson and Simpson (1977): their equation (4) using Type I water according to the classification of Jerlov (1968)] shows that during 24 h in summer, phytoplankton at the DCM depth would receive only 50% of the winter light quantity. Another result of strong winter vertical mixing is erosion of the DCM, leading to larger sea-surface chlorophyll concentrations in winter

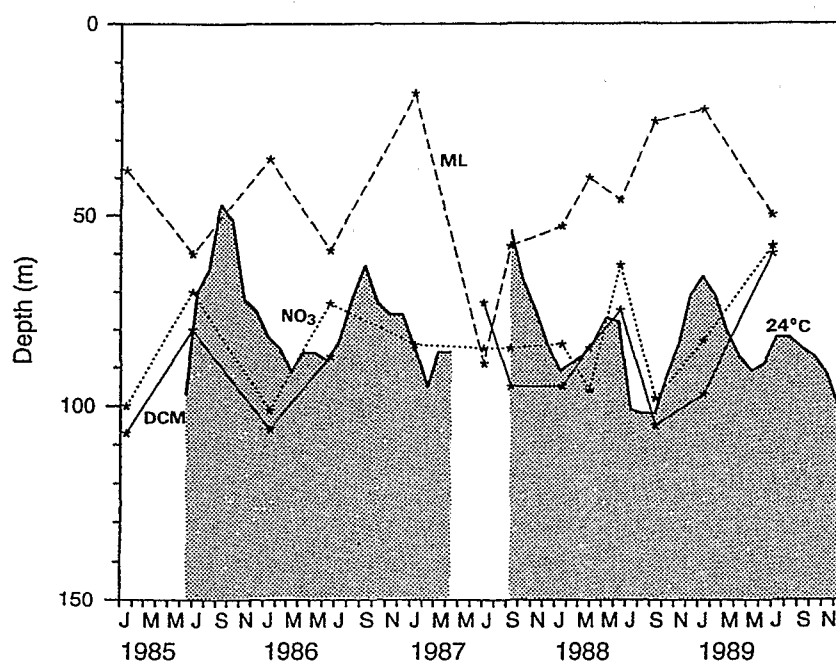


Fig. 8. Time-series of the mean depths of the mixed layer (dashed line), the nitracline (dotted line), the deep chlorophyll maximum (solid line) and the 24°C isotherm (area below this isoline is shaded) in the 20–17°S region.

than summer. Because chlorophyll data were unreliable prior to July 1987, we can only propose annual cycles of chlorophyll during El Niño and La Niña periods. The results were very consistent with the previous results of Dandonneau and Gohin (1984) who observed an increase of the surface chlorophyll concentration south of 20°S in the tropical western Pacific, using samples taken from voluntary observing ships from 1978 to 1982. Expanding on these authors' results, we found the northern limit of the winter increase of surface chlorophyll around 17°S along 165°E. The mean surface chlorophyll concentration was 0.07 mg m⁻³ in summer and it doubled in winter. On the contrary, annual changes of the DCM concentration (0.25–0.30 mg m⁻³) or the integrated chlorophyll concentration (23 mg m⁻²) were not significant.

No clear interannual variations in the depths of the nitracline and DCM were observed during El Niño and La Niña. However, the annual cycle was somewhat unclear, and only a remnant of the seasonal cycles of the mixed layer, nitracline and DCM depths was discernible. Deviation from the seasonal cycle may be partly the result of the complicated and rapidly changing alternation of divergences and convergences in the southern part of the transect. South of 15°S, a succession of eastward and westward surface currents led to an intricate system, changing drastically from one cruise to another. This trend seems to have been enhanced from July 1987 to the end of the period under consideration and may have been responsible for the alteration of the annual cycle of the homogeneous layers' depths.

The 6–10°N region. In spite of the fact that the highest nitrate concentrations of the transect were found at 200 m, sea-surface chlorophyll concentrations remained low (0.10 mg m⁻³ or lower) throughout the year (Fig. 6). The vertical density gradient (0.10 kg m⁻⁴) was strong enough to enhance static stability, reducing the possibility of upward nitrate flux and trapping the DCM at depth. The chlorophyll profile in all seasons was similar to the summer chlorophyll profiles for 20–17°S, although the DCM concentration was somewhat lower (0.15–0.25 mg m⁻³). The mixed layer varied both seasonally and interannually (Fig. 9). Its weak annual cycle was the result of stronger winds in winter and slightly higher surface warming in summer. During El Niño, the mixed layer was reduced by 20–30 m and reached its minimum depth of 40 m. In 1988, the first year of La Niña, the mixed-layer depth deepened to 70–95 m. In 1989, it returned to normal. The variation of the mixed-layer depth was not echoed in the nitracline and DCM depths, which were always deeper. The amplitude of the seasonal solar cycle is weak in this region, thus the daily solar radiation also had little influence on vertical motions of the DCM. The annual and interannual vertical displacements of the DCM were consistent with the general trend of the upper thermocline motion represented by the 25°C isotherm. Shallowest and deepest depths were reached in boreal autumn and spring, respectively, during reference years (Fig. 9). Apparent discrepancy between the thermocline and DCM depths is also attributed to an unsatisfactory sample resolution.

The seasonal and interannual variations in the depths of thermocline and DCM were partly explained by Ekman pumping. This is a result of a balance between two terms:

$$-\partial h/\partial t = \text{curl}_z(\tau/\rho f),$$

where h is the 25°C isotherm depth (positive down; derived from monthly XBT data in the 6–10°N, 155–166°E region), τ is the wind stress (derived from monthly FSU wind values in the 5–11°N, 164–166°E region), ρ is the density of sea water and f is the Coriolis parameter.

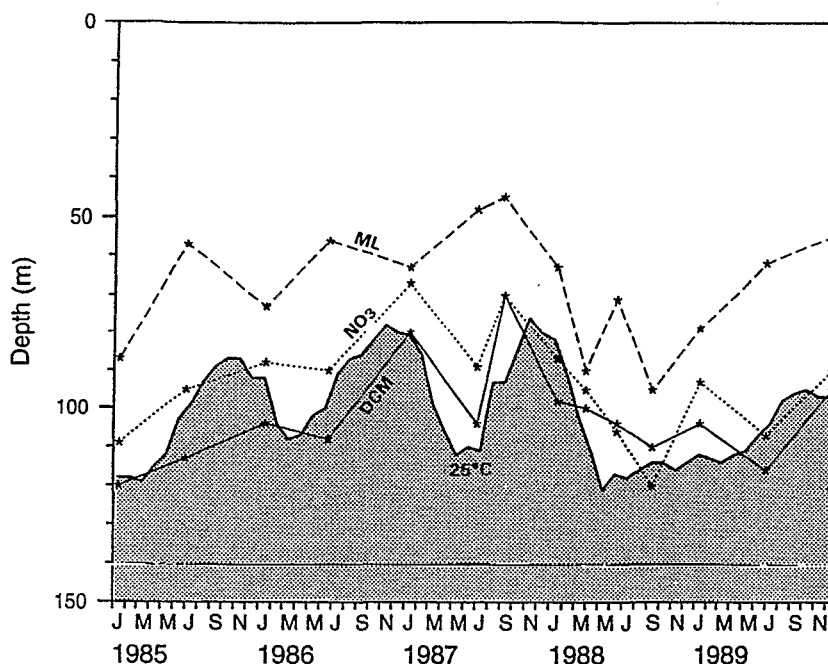


Fig. 9. Time-series of the mean depths of the mixed layer (dashed line), the nitracline (dotted line), the deep chlorophyll maximum (solid line) and the 25°C isotherm (area below this isoline is shaded) in the 6–10°N region. Note that there is only one station (6°N) for the PROPPAC cruises (September 1987, March–April 1988 and September 1988).

A comparison between the two terms is shown in Fig. 10 (the upward motion is positive). The correlation between the two terms was high during El Niño (August 1986–October 1987). During La Niña no statistical relation was found, although the Ekman pumping effect was consistent with the sudden deepening of the mixed layer, nitracline and DCM depths at the beginning of 1988, and that, in 1989, the strong negative Ekman pumping was probably a cause of the unusually small upward motion of the thermocline. During the reference years, a correlation existed in the 7–9°N region. These results agree with previous studies of the Ekman pumping influence on the thermocline motion at these latitudes. Tournier (1989) showed that the strong depth anomaly observed in the 5–9°N region during the 1982–1983 El Niño resulted from Ekman pumping in the central as well as in the western Pacific. On an annual basis, the vertical motion of the 20°C isotherm in the central Pacific was strongly related to the Ekman pumping vertical velocity (Meyers, 1979; Kessler and Taft, 1987; Tournier, 1989), whereas, in the western Pacific, the thermocline depth variations were partly accounted for by Ekman pumping (Tournier, 1989). Meyers (1979) argued that the propagation of westward Rossby waves may be a major process affecting the vertical structure. In the central Pacific, Ekman pumping was greatest at the poleward side of the NECC, whereas its equatorward side came under other influences (Kessler and Taft, 1987). In conclusion, the nitracline and DCM depths depend essentially on the thermohaline structure, and the local wind forcing, through the Ekman pumping mechanism, has interannual or seasonal influences.

The equatorial zone and its borders: strong interannual variations

Study of the equatorial zone was restricted to the 2°S–2°N latitude band except when upwelling was effective, when the distributions in the zone under the influence of upwelling

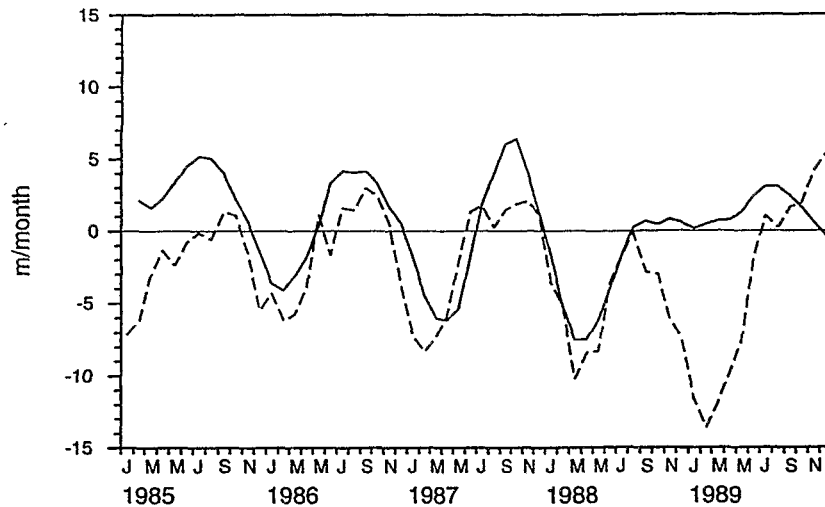


Fig. 10. Time-series of the Ekman pumping balance in the 6–10°N region. Solid line is $-\partial h/\partial t$, dashed line is $\text{curl}_z(\tau/\rho f)$. Both terms are filtered using a 5-month running mean.

were studied. The time-series of mixed-layer depths (Fig. 11) showed both seasonal and interannual variability. At the top of the thermocline, the well-known interannual signal prevailed, and at shallower depths shorter timescale processes were superimposed. These physical processes also influenced the biological distributions.

Reference conditions. During the reference period, the top of the thermocline was 80–120 m deep (except in January 1985 when the deep thermocline was associated with an eastward equatorial current) and the mixed-layer depth was only at 40 m. Thick freshwater layers were observed during boreal winter when westerly wind and ensuing eastward equatorial current were more likely to occur. Zonal and meridional advection of low-salinity water from the western Pacific contributes to the formation of the low-salinity surface layer, which is also concomitant with stronger local precipitation in December–March (Porte, 1992). In 1985–1986, the nitracline and the DCM lay in the “barrier layer” where salinity strongly influenced the density structure. Their depths followed an annual cycle not related to light variations, as all cruises took place during solstices. Variations in depths of the nitracline and DCM appeared to be related to the fluctuations of the depth of the salinity gradient just below the fresher surface water. Despite the paucity of data, agreement is discernible between vertical displacements of the DCM and salinity isopleths. Vertical density gradients remained weak (lower than 0.03 kg m^{-4}) below the mixed layer and strong gradients were observed (greater than 0.10 kg m^{-4}) below 100 m. Weak stratification probably allowed some vertical mixing over the euphotic zone, and the situation was unfavourable for phytoplankton growth (Sverdrup, 1953) and development of an intense DCM.

El Niño conditions. During the 1987 El Niño, the top of the thermocline shoaled to 40–50 m in July–September when the equatorial upwelling Rossby wave was observed in the western Pacific (Delcroix *et al.*, 1992). Combined with the nearly unchanging depth of the mixed layer, the “barrier layer” disappeared. Variability of the thermohaline structure in the western Pacific results from several processes, and an essential point is that no barrier layer is found when upwelling Rossby waves reach the western Pacific (Shinoda and Lukas, 1995).

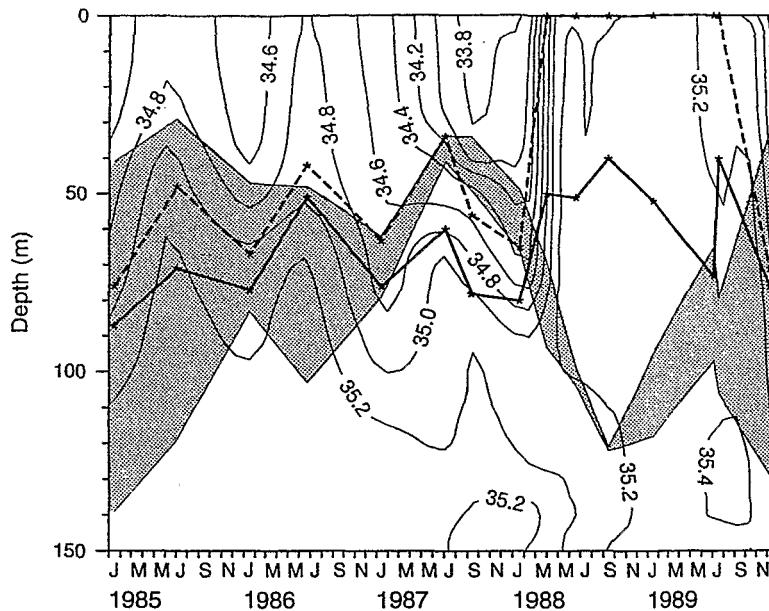


Fig. 11. Vertical salinity distribution in the 2°S–2°N region (contour interval is 0.1). Superimposed are the barrier layer (shaded area), the mean deep chlorophyll maximum depth (heavy line) and the mean nitracline depth (dashed line).

The fresh surface layer was probably the consequence of the advection of low salinity waters by the surface eastward current (Fig. 4) as well as increased local precipitation (Porte, 1992; Delcroix and Henin, 1991). A strong salinity stratification developed at the base of the freshwater layer, which, associated with the thermocline shoaling, contributed to the formation of the maximum vertical density gradient (about 0.13 kg m^{-4}) at the base of the mixed layer. In the western Pacific, shallower nitracline and DCM were expected because of the change in the basin-wide slope of the thermocline. In contrast to the eastern Pacific (Barber and Chavez, 1983), this resulted in favourable conditions for photosynthesis and an increase in primary production. Accordingly, Barber (1990) found a 25–50% increase of the average primary production in the warm pool between February 1986 and October 1987. An increase of the available nitrate to the euphotic zone resulting from the shoaling of the nutrient reservoir is an attractive interpretation for this increase in production. This assumption, however, must be viewed with caution, since, compared with the reference period, there was no significant rise of the nitracline or of the DCM (Fig. 11). In fact, as in 1985–1986, both the nitracline and DCM occurred in the strongly stratified zone at the bottom of the freshwater layer. Depth variations observed during the El Niño period remained in the same range as during the reference years. The DCM was 60 m deep in July 1987 and had shifted deeper by 20 m 2 months later. The reason for the enhanced primary production is probably due to the shallower depth of the maximum density gradient, which, in contrast with the reference years, confined the vertical mixing to a surface layer much thinner than the euphotic zone.

Other consequences of the strong density gradient were the limitation of the exchange of nutrients toward the surface layer and the low surface chlorophyll content (0.10 mg m^{-3} ; Fig. 6). This also was corroborated by ocean colour remote-sensing studies of the strong 1982–1983 El Niño (Dupouy-Douchement *et al.*, 1993) and merchant ship data from the moderate 1986–1987 El Niño (Dandonneau, 1992). Surface waters were abnormally poor in chlorophyll in 1982 and 1987.

In conclusion, the thermocline tilt emerges as a basic factor contributing to the increase in primary production during El Niño for two reasons: nutrients are lifted toward the euphotic zone, and association of the top of the thermocline with the surface halocline leads to a shallow strong stratification, which limits the vertical mixing of the surface layer. The variable depths of stratification resulted in variable depths of the DCM and, we assume, also in variable but enhanced primary production. The strong density gradient prevented exchanges with the surface where chlorophyll remained unusually low.

La Niña conditions. In 1988–1989, trade winds were significantly stronger than climatology (McPhaden *et al.*, 1990), and upwelling took place at the equator and 165°E (Fig. 6). The basic mechanism was the surface Ekman drift as illustrated by Cromwell (1953). A meridional circulation was maintained by sustained easterly trade winds that induced an equatorial divergence and an adjacent convergence situated downwind off the equator. The divergent equatorial flow was balanced at depth by the convergence of geostrophic flow resulting from the zonal pressure gradient along the equator. Variation in surface nitrate deduced from our cruises was consistent with variation of SST deduced from mooring data (Fig. 12). It is highly probable that a better resolved time-series of surface nitrate would roughly mirror SST as only a small amount of new nitrogen is used for biological consumption (Cullen, 1991) and because nitrate renewal by vertical advection occurs over short timescales. As expected, the variations in SST and surface nitrate appeared to be mostly related to the local zonal wind stress (Figs 2 and 12). After easterly anomalies developed in March 1988, low SST occurred and nitrate rapidly became non-limiting at the surface. The easterlies eased in May–June, and surface nitrate decreased substantially in June 1988. During the two subsequent cruises (September 1988 and January 1989), nitrate was more abundant, but decreased considerably in July 1989 when the easterlies returned to the pre-1986 conditions and a deep barrier layer formed again. Apparently due to the same influences, surface chlorophyll concentration varied in the same way as surface nitrate.

In the central Atlantic, Weingartner and Weisberg (1991) located an average maximum vertical velocity just above the core of the EUC. This corroborates the assertion that the depth of origin of the upwelled water is located in the upper thermocline (Wyrtki and Eldin, 1982) and, therefore, in the upper part of the nutrient reservoir.

The depth of the top of the thermocline in our study showed a strong interannual signal. Subsequent to El Niño, it deepened in early 1988, reached a maximum value in late 1988 and returned to a shallower depth in 1989 (Figs 11 and 12). A striking feature is the sharp uplift of the thermocline in early 1989, consistent with the signature of an upwelling Kelvin wave that evolved in the western basin following an easterly wind anomaly (Delcroix *et al.*, 1994). Intraseasonal Kelvin waves affects variability of the thermocline depth (Kessler *et al.*, 1995), leading to a modification of the efficiency of the equatorial Ekman divergence. A better resolved time-series would be necessary to show the influence of intraseasonal Kelvin waves on the nutrient and chlorophyll variability. Nevertheless, analogous with studies in the central Pacific (Murray *et al.*, 1994), we assert that the nitrate distribution, like the temperature distribution, is affected by remote forcing. The 5 μM nitrate isopleth (Fig. 12, lower panel) shows the large-scale variation of the depth of the nutrient reservoir. It is probable, for example, that in January 1989, when an upwelling Kelvin wave was identified (Delcroix *et al.*, 1994), the nutrient reservoir shoaled during several weeks. We infer that nitrate and chlorophyll surface values result from both local and remote forcing, and the

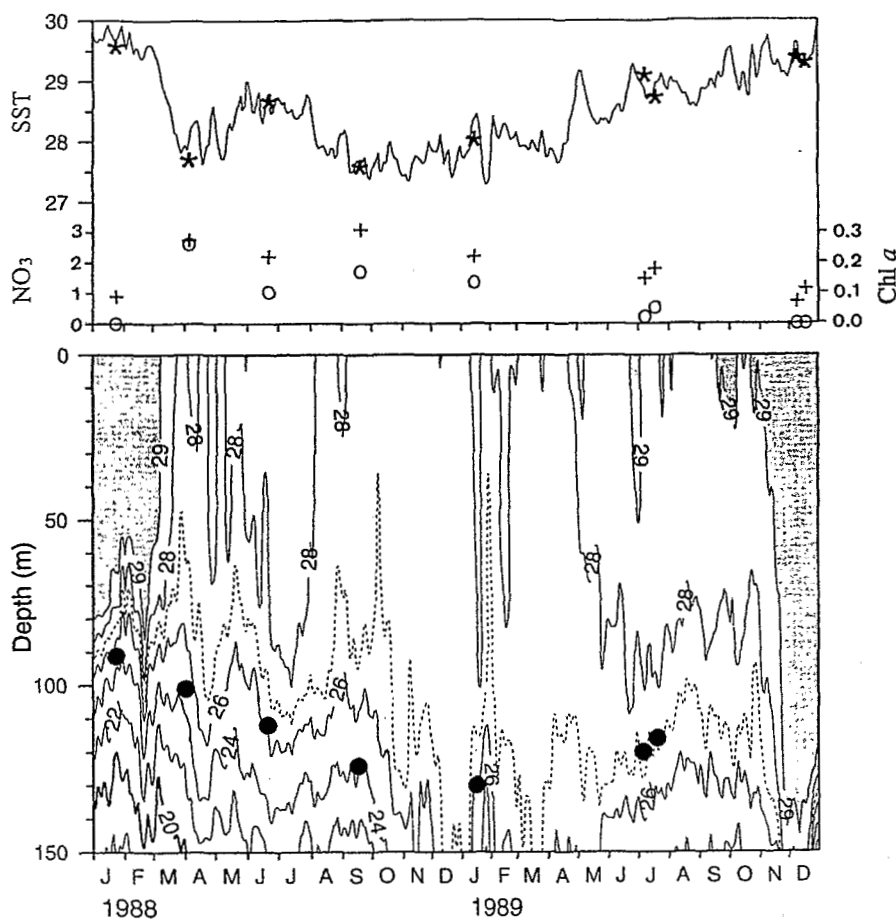


Fig. 12. Time-series of equatorial temperature, nitrate and chlorophyll *a* at 165°E in 1988–1989. Upper panel: TAO SST with SST from cruises (★), surface concentrations of $[\text{NO}_3^- + \text{NO}_2^-]$ (○) and chlorophyll *a* (+). Lower panel: TAO vertical temperature distribution with the $5 \mu\text{M}$ $[\text{NO}_3^- + \text{NO}_2^-]$ values (●). TAO temperature are deduced from the equatorial mooring at 165°E, $[\text{NO}_3^- + \text{NO}_2^-]$ and chlorophyll *a* concentrations are averaged from measurements taken during cruises from 2°S to 2°N.

increase in nitrate surface concentration probably occurred simultaneously with the SST decrease in late January.

The northern and southern boundaries of the enrichment zone, although presenting asymmetrical patterns, were sharply delimited within 5° of the equator (Figs 5 and 6). This contrasts with the strongly asymmetrical distribution in the central Pacific where the nutrient-rich surface water ends at 4–5°N but extends as far south as 10°S (Bender and McPhaden, 1990). The meridional termination of upwelling is the consequence of the wind-drift convergence and oceanic convergence caused by meridional shear between the SEC and counter currents, the dynamics of which are mainly ruled by the curl of the wind-stress (Sverdrup, 1947).

Two characteristics of the atmospheric convergence zones appear to be major reasons for differences between western and central Pacific upwelling: the existence of the SPCZ and the tendency of the ITCZ to extend to the southern hemisphere in the western Pacific. Because of the presence of the SPCZ, the SECC is a characteristic permanent feature of the western Pacific (Kessler and Taft, 1987). Thus, only in the western basin are the upwelled waters impeded by oceanic convergences at both edges of the SEC. Associated with the seasonal

migration of the ITCZ across the equator is a reversal of the meridional component of the wind. As opposed to the central Pacific, wind-drift convergence also may be situated south the equator. According to dynamic analysis by Charney and Spiegel (1971) and Cane (1979), an eastward current develops at the equatorward side of the counter current in response to the meridional component of the wind, and superimposes on the wind-stress curl driven solution. Therefore, convergences bordering the zone of upwelling may be more intense on the northern (southern) side during periods of southeasterlies (northeasterlies) (Cane, 1979; Philander *et al.*, 1987). An illustration is given by data from the April 1988 cruise (Fig. 5) when the strongest meridional gradients were found on the southern side of the upwelling zone (4°S) when the prevailing equatorial wind was northeasterly. We reported in the last section on the meridional displacement of upwelling toward the winter hemisphere (Fig. 6) following the alternation of northeast (December–May) and southeast (June–November) equatorial trades. This had been previously described by Oudot and Wauthy (1976) and is consistent with the model derived by Cromwell (1953). This was confirmed in terms of meridional Ekman transport, for which monthly values were computed from the monthly FSU wind data for 1988–1989 at 5°N and 5°S (not shown here). The transequatorial Ekman transport (sum of the north and south components of the Ekman transport) was directed toward the winter hemisphere as a consequence of the meridional migration of the ITCZ to south of the equator. Note that because the mean zonal wind-stress is stronger in the north than in the south (Fig. 2), the resulting annual flow from July 1988 to June 1989 in the 5°S – 5°N latitude band was northward. In addition, seasonal meridional variations of countercurrents are closely related to variations in Ekman pumping (Kessler and Taft, 1987; Tournier, 1989). The NECC (SECC) reaches its closest position to the equator in boreal (austral) summer (Kessler and Taft, 1987; Tournier, 1989; Gouriou and Toole, 1993).

Compared to situations of the reference and El Niño periods, upwelling is the main factor modifying the vertical structure of water properties. The mean chlorophyll profile had a smooth maximum (0.35 mg m^{-3}) at 50 m. Similar mean profiles of upwelling situations have been described at the equator by Fig. 11 of Chavez *et al.* (1991) from 95°W to the date line between 1980 and 1988, and by Fig. 4 of Cullen (1991) at 150°W in March 1988. Zonal variability of biological properties in the equatorial upwelling is relatively small (Cullen, 1991; Le Bouteiller and Blanchot, 1991). Individual profiles reveal diverse shapes, but remarkably, the depth of the chlorophyll maximum is relatively constant (40–60 m; Fig. 11). Although higher than during non-upwelling periods, surface chlorophyll concentrations remained within the 0.12 – 0.43 mg m^{-3} range. The question of low chlorophyll concentration despite persistent nutrients (high-nutrient–low-chlorophyll regime) is beyond the scope of this paper, and several explanations are being debated in the scientific community [see Cullen (1991) for a review]. However, it is noteworthy that because of the characteristics of the barrier layer in the western equatorial Pacific (absent from early 1988 to mid-1989 and deep in the second half of 1989 when the effects of upwelling diminished), the mixed layer was deeper than the euphotic zone, which is one of the causes of biomass and primary production limitation. Nevertheless, it was during the La Niña period when the highest primary production was observed in the western equatorial Pacific. In April 1988, it reached $130\text{ mmol m}^{-2}\text{ day}^{-1}$ which was more than double the El Niño value of September 1987 (Le Borgne *et al.*, 1989). If the ratio between the euphotic zone depth and the mixed-layer depth is a relevant factor for phytoplankton biomass and primary production, propagation of an upwelling Kelvin wave will considerably restrain the depth of the mixed layer for several weeks and will likely affect biological profiles significantly.

10°S: a privileged location for short events

The ridge of the isolines in the 10°S region was described by Merle *et al.* (1969) and Oudot and Wauthy (1976) using data from cruises along 170°E. The doming was reported as a subsurface structure, sometimes with a surface enrichment in nutrients or chlorophyll. This feature appears to be variable in both space and time, and is thus seldom sampled. For example, Dandonneau (1992) reports no special surface chlorophyll increase around 10°S in the 1978–1989 period. During the 1984–1989 cruises, the 10°S ridge was often reported without any seasonal or interannual recurrence. An example of surface enrichment was seen in mid-September 1987 when high surface nitrate (0.60 μM) and chlorophyll (0.27 mg m^{-3}) concentrations were observed. The increase of integrated chlorophyll concentration to 40 mg m^{-2} corresponded to an enhancement of 58% compared to the mean values of all cruises at this latitude. Three weeks after its first observation (Le Borgne *et al.*, 1992), surface nitrate had disappeared, although the chlorophyll augmentation was still detectable. High biomass in a nitrate-depleted upper layer has been observed a few times at this latitude, perhaps reflecting previous brief nutrient enrichments. We also note that part of this high biomass value sometimes results from manifestation of *Trichodesmium* sp., which is often observed at the surface between 8°S and 15°S. An explanation to this ridge would be the role of the vertical displacements of water, induced by the zonal circulation, in the vertical distribution of nutrients (Merle *et al.*, 1969). We recall that the SECC is a permanent feature of the western Pacific, and divergence at its southern edge causes a shoaling of the hydrographic structures. The second main explanation is that Ekman pumping is always favourable to upwelling in this region (Delcroix and Hénin, 1989). Therefore, more easily here than at surrounding latitudes, a local wind event can result in turbulent entrainment at the base of the mixed layer, allowing nutrient input into the surface layer. If this analysis is correct, surface enrichment is more probable in the middle of the year or during El Niño periods, when both the wind stress and its curl (not shown) are stronger. Unfortunately, this variability was not resolved by this data set. An additional explanation is that the effect of horizontal advection could be non-negligible in a location so close to the Solomon Islands. Despite their variability, consequences of these transient phenomena on the phytoplankton community structure are far from insignificant (Blanchot *et al.*, 1992).

SYNTHESIS

From 20°S to 10°N along 165°E three main periods were determined during 1985–1989: the reference year from 1985 to mid-1986; the 1986–1987 El Niño; and the 1988–1989 La Niña. Several situations were described: low-density stratification associated with a poor and deep nitrate reservoir in the south; high-density stratification associated with a shallow, rich nitrate reservoir in the north; and effects of westerly wind bursts and upwelling at the equator, doming at 10°S. For all of these situations, nitrate and chlorophyll distributions were closely linked to the density structure for which salinity can be the controlling factor. A notable exception occurred in December 1989 when modifications of the thermohaline and nitrate distributions due to a westerly wind burst were not echoed on the chlorophyll distribution at the considered timescale.

In the south (20–17°S), the vertical gradients are the weakest of the transect and the nutrient reservoir is deep. During austral winter, because of stronger vertical mixing and lower available amount of light, phytoplankton find adequate nutrient and light conditions

at a shallower depth than in summer. The winter mean surface chlorophyll concentration (0.16 mg m^{-3}) doubled the summer value. During the 1986–1987 El Niño, and the subsequent La Niña, no clear interannual variations in the depths of the nitracline and deep chlorophyll maximum were evidenced.

The north ($6\text{--}10^\circ\text{N}$) is the most oligotrophic region of the transect (surface chlorophyll always lower than 0.10 mg m^{-3}). Seasonal changes occurred in the subsurface layer, as strong density stratification always insulates the surface from deeper layers. Ekman pumping was a major factor controlling seasonal or interannual variations in the depth of the deep chlorophyll maximum.

At the equator, seasonal, intraseasonal and interannual variability was seen. During the reference period (from 1985 to mid-1986), a warm pool structure with a barrier layer occurred. The surface layer was nitrate depleted, and the surface chlorophyll concentration was low ($0.10 \pm 0.05 \text{ mg m}^{-3}$). Variations in the salinity structure strongly influenced seasonal variability of the density stratification of the upper part of the barrier layer where the nitracline and deep chlorophyll maximum were observed. During boreal winter, the density gradient, which traps the deep chlorophyll maximum at the base of the freshwater layer, was deeper because of stronger probability of eastward freshwater transport and intense local precipitation. The deep chlorophyll maximum varied from 70 to 80 m in the beginning of the year to 40–50 m in the middle of the year. During the 1986–1987 El Niño, the barrier layer vanished in July–September 1987, concurrently with the presence of an upwelling Rossby wave in the western Pacific (Delcroix *et al.*, 1994; Shinoda and Lukas, 1995). During this period, a strong and shallow pycnocline trapped the deep chlorophyll maximum at the base of the mixed layer. A direct consequence was the abnormally low surface chlorophyll concentrations (0.10 mg m^{-3}) because of limited exchange of nutrients toward the surface layer. Intra-El Niño variations in the depth of the deep chlorophyll maximum were caused by variations in the depth of the freshwater layer. These depth variations remained in the same range as during the reference period. Although enhanced compared to the reference period, we assume the primary production to be variable during the El Niño. During the 1988–1989 La Niña, the vertical structure was completely modified (surface nitrate up to $3.23 \mu\text{M}$; surface chlorophyll greater than 0.20 mg m^{-3} ; primary production values about doubled compared to the El Niño values) as equatorial upwelling extended far to the west of the date line.

In contrast with the central Pacific, upwelling was sharply bounded within 5° of the equator. Indeed, in the western Pacific, convergence between the South Equatorial Current and the South Equatorial Counter Current is a permanent feature. The meridional displacements of the surface enrichment zone toward the winter hemisphere followed the alternation of northeast (December–May) and southeast (June–November) equatorial trade winds. Variability of surface nitrate and chlorophyll were consistent with local wind forcing. In addition, the efficiency of the upwelling was enhanced (reduced) because of the shoaling (deepening) of the thermocline and nutrient reservoir. Similar to studies in the central Pacific (Murray *et al.*, 1994), we assert that the nitrate distribution was influenced by the passage of intraseasonal Kelvin waves. The December 1989 cruise took place during a period of successive westerly wind events and shows the persistence of a deep chlorophyll maximum in a layer where nitrate were apparently depleted. The deep chlorophyll maximum was about 50 m above the nitracline. During this wind event, a barrier layer probably insulated the nutricline from the surface, and meridional advection of richer water may be partly responsible for the persistence of the deep chlorophyll maximum.

The 10°S region is characterized by a doming of the thermohaline structure and transient nitrate and chlorophyll surface enrichments. Turbulent entrainment from the nutrient reservoir is easier in this region where divergence between the South Equatorial Counter Current and the southern branch of the South Equatorial Current (Merle *et al.*, 1969) and Ekman pumping favourable to upwelling (Delcroix and Hénin, 1989) bring the nutrient reservoir closer to the surface. Probability of surface enrichments is higher in the middle of the year or during El Niño.

Mixing processes at different time and space scales are responsible for modifying the depth of the mixed layer or the intensity of the mixing. Favourable or unfavourable consequences of vertical mixing on biomass and production are still debated (Sverdrup, 1953; Prieur and Legendre, 1988; Belyaev, 1992), and it needs to be better resolved among the many factors (grazing, iron limitation) that influence biomass distribution. One aspect is the study of the importance of local and remote forcing on nutrients and biological distributions in this region where Kelvin waves are generated.

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