Pergamon

0025-326X(94)00095-6

Marine Pollution Bulletin, Vol. 29, Nos 1–3, pp. 14–25, 1994

0025-326X/94 $7.00+0.00

The Polynesian South Ocean: Features and Circulation

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The hydroclimatic pattern of the French Polynesian zone is determined mainly by the pressure difference between the Easter Island High and the Western Equatorial Low. Trade wind activity is correlated with this gradient and drives surface oceanic water westward to produce the South Equatorial Current, which constitutes the northern part of a vast anticyclonic gyre in the South Pacific. High evaporation from the centre of the anticyclonic high results in the formation of hypersaline surface water, the Southern Tropical Water, which is dense enough to sink. This downwelling process removes plankton and particles from the euphotic zone and contributes to the strong oligotrophy of the gyre. Thermohaline pattern underlines the crucial role of vertical mixing processes in the upper mixed layer, which is isolated from the deeper Antarctic Intermediate Water by a stable pycnocline. Oceanic circulation measured in the band 140–145°W between latitudes 8–28°S during six HYDROPOL cruises (1986–1989) showed a weak (5–10 cm s^{-1}) shallow (0–300 m) westward flow, South Equatorial Current, that was often dissected by small eastward countercurrents. The eastward component was increased considerably during the ENSO event of 1987 indicating an aperiodic forcing superimposed on the dynamic instability of the zone.

Data gathered in the vicinity of barrier and atoll reefs showed that oceanic conditions (that is oligotrophic mixed layer) prevailed around these structures in spite of periodic thermal oscillations related to the tide regime. The Polynesian oceanic province being mainly in low dynamical conditions, vortex and eddies created around islands are insufficient to modify the vertical stratification or to pump up nutrients from the nutricline, that explains the absence of local upwelling or 'island mass effect'. The permanence of a clear oligotrophic, euphotic layer then allows coral and reef communities to thrive and reach the atoll stage.

The Great Southern Anticyclonic Gyre

Situated more than 5000 km from the American, Australian and Antarctic continental coastlines, the Central Southern Pacific is a maritime region with features dependent on medium and large scale ocean–atmosphere interactions, interannual variability, and aperiodic ENSO (El Nino–Southern Oscillation) anomalies.

Hydroclimate of the Central Southern Pacific

The atmospheric driving forces consist of an anticyclonic high pressure zone located in the region of Easter Island (110°W–28°S) coupled with a quasi-permanent depressive system in the central and western equatorial zone that reaches its lowest levels in the northern Australian–Indonesian zone during the Austral winter. The pressure gradient between the southeastern subtropical Pacific and the western equator causes a transfer of air mass, in accordance with the global circulation model of Walker (Ramage, 1970). This circulation, orientated diagonally from the southeast to the northwest, is fed in the lower troposphere by the southeast trade winds, and in the higher troposphere by a returning circulation (the westerly high altitude jet streams). In the equatorial zone, a powerful vertical convection of warm and humid air feeds the high troposphere (ascending branch of the Walker model); this air returns to low altitudes in the returning circulation after being transported east and south to the subtropics (Fig. 1(a)).

Because of the predominance of the austral pressure field over the boreal pressure field, the southern hemisphere trade winds are stronger than their counterparts of the northern hemisphere; consequently, the southern Pacific trade winds cross the equator and reach 5°N. Between 5 and 10°N, there is a windless zone where the trade winds of the two hemispheres converge called the Inter-Tropical Convergence Zone (ITCZ). Where the trade winds of the two types have different concentrations of water vapour and temperature, high levels of condensation occur in this zonal band and rainfall can exceed 4 m yr^{-1}; these are the famous 'doldrums', calm and rainy, dreaded by navigators. A second line of convergence, called the South Pacific Convergence Zone (SPCZ) stretches diagonally from Papua New Guinea to the southeast of French Polynesia (Cauchard, 1985). The SPCZ is a convergence front between the two types of trade winds that blow in the southern Pacific: the northeastern trade winds (controlled by the Easter Island High) and the southeast trade winds, controlled by the high subtropical pressure ridge, east of the Kermadec Islands. The SPCZ, strong in the austral summer, also constitutes a zone of weak winds and heavy rainfall. Its presence above the French Polynesian archipelago determines the strength of the rainy seasons, but its influence

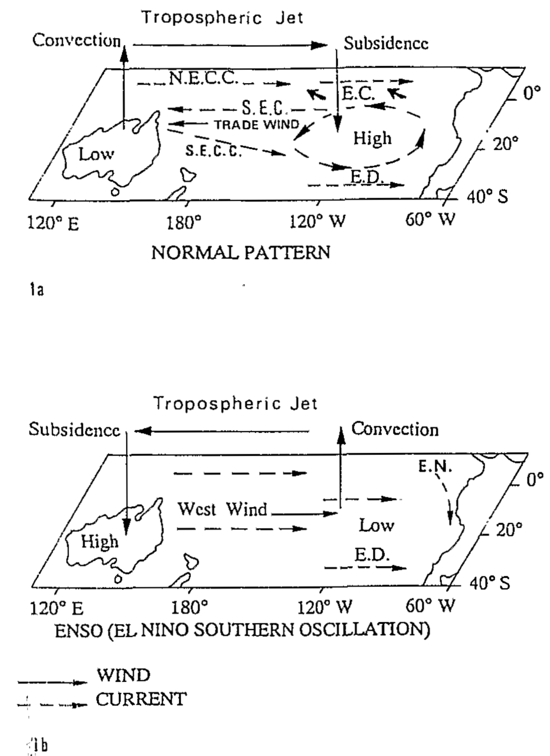

Fig. 1 Hydroclimatic features of the equatorial and South Pacific. (a) Normal pattern with dominant trade winds and correlative vertical movements (Walker circulation). Oceanic circulation in the band 4°N–20°S is mainly westwards. (b) ENSO situation with inversion of the Walker circulation. Oceanic circulation in equatorial/south tropical zone is mainly eastwards. N.E.C.C., North Equatorial Counter Current; S.E.C.C., South Equatorial Counter Current; E.C., Equatorial Current; E.N., El Nino; S.E.C., South Equatorial Current; E.D., Eastward Drift.

on the Tuamotu Archipelago is generally limited to its western border. In the Easter Island High, the descending dry air mass maintains permanent aridity and low cloud cover. Rainfall is low (in the order of 1 m yr^{-1}) and the strong evaporation can surpass 2 m yr^{-1}. This situation occurs in the Tuamotu Archipelago, especially in its southeastern part. The Evaporation–Precipitation budget (E–P), is therefore strongly positive reflecting the dominance of evaporation: E–P is in the order of +80 cm yr^{-1} around Mururoa Atoll and +40 cm yr^{-1} around Rangiroa Atoll.

This coupled ocean–atmosphere equilibrium is influenced by profound, aperiodic disturbances known as ENSO (El Nino–Southern Oscillation) episodes. A very strong event with many climatic and biological impacts was observed in 1982–1983; lesser events have since been observed in 1987 (medium strength) and 1991–1993 (strong) (Fig. 1(b)). Each ENSO period lasts about 12–18 months. Its onset is signalled by a weakening of the Easter Island High and a proportional increase of the equatorial low pressure fronts (Hansen, 1990). When the normal pressure gradient across the Pacific is absent, the trade winds weaken or stop, and the surface currents slacken. The ITCZ migrates south of the equator to 0–10°S and heavy rains fall in that region. This phenomena extends from the Central Pacific to the coast of Peru where a warm poleward current appears called the El Nino Current. The French Polynesian area is then subject to cyclones favoured by converging westerly winds up to the central Tuamotu. Drought conditions are experienced in the western Pacific and in southern Polynesia while the Marquesas (8–12°S) record heavy increases in rainfall (6 m in 1992 compared with an annual average of 1.2 m).

Oceanic circulation in the Central South Pacific

Through surface friction, the trade winds induce a large scale oceanic circulation that is directly linked to the South Pacific anticyclonic system. This geostrophic stream flows to the west and following classical nomenclature (Tabata, 1975), divides into the Equatorial Current (4°N–8°S band) and the Southern Equatorial Current (10–20°S band). Because of the Coriolis force, Ekman's rotation produces a leftward deviation in these great currents which ends in the creation of a vast anticyclonic vortex or gyre, with its oriental part located under the high pressure system from Kermadec to the Easter Island sector (Fig. 1). The southern border of the gyre consists of a returning easterly current which flows in the subtropical zone (25–35°S), favoured by the dominant westerly winds. The northern branch of the gyre consists of the South Equatorial Current, with an average speed of 10–20 cm s^{-1}. North of 10°S appears the Equatorial Current; it is more rapid and constant with an average speed of 50 cm s^{-1} (Wyrtki, 1981). This current is primarily geostrophic south of the equator; however, in the equatorial band, the direct wind-driven flow plays a major role: due to the Coriolis force this flows to the left (right) of the wind south (north) of the equator, while on the equator it flows directly downwind. Consequently there is a divergence of the surface waters and a vertical compensation through upwelling of subsurface waters. This upwelling zone can stretch 200–500 km each side of the Equator and is easily identified by its temperature which is 2–5°C lower than neighbouring tropical waters. The strength of the upwelling, and the temperature anomaly is directly dependent on the combination of trade winds and Equatorial Current. An ENSO anomaly signifies an important restructuring of the equatorial dynamics and especially the disappearance of the divergence at the surface in the Central Pacific. For example, monitoring at 140°W showed that upwelling vanished from July 1982 to May 1983 as a result of anomalous westerly winds that created an inversion of the surface current (rotation from west to east) and a convergence of warm waters towards the equatorial axis (Donguy *et al.*, 1989).

Six HYDROPOL cruises (1986–1989) were made with the R.V. *Marara* (Rancher & Rougerie, 1993) in the Polynesian waters along a track Tahiti–North of Marquesas (8–5°S)–Rapa (28°S)–Tahiti. The physical and chemical properties of water masses were determined by using the following techniques: a MORS SLS57 CTD to measure conductivity, temperature and depth with a precision of 0.1%, 0.02°C, 0.1%, respectively, with adequate calibrations before and after casts. Concentrations of nutrients were analysed on a Technicon AutoAnalyser immediately after sampling. Detection limits expressed in µM were 0.02 for nitrite; 0.1 for nitrate; 0.1 for silicate and 0.05 for phosphate. Dissolved oxygen concentrations were obtained with an analytical precision of ±0.03 ml l^{-1} with the Winkler method in using

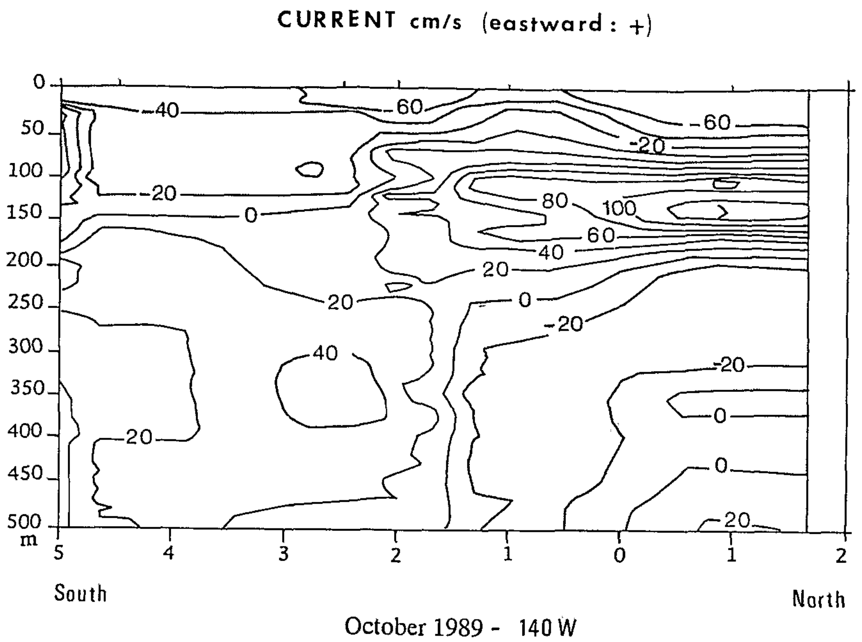

Fig. 2 Zonal currents in the equatorial band in October 1989 (N.O. *Marara* Cruise HYDROPOL 8—140°W). Direct measurements with a current profiler. Equatorial Current (Shading) reaches 60 cm s^{-1} along the equatorial line with a maximum depth of 170 m at 5° South. Eastward undercurrent (Cromwell current) is stronger with a core at 80–180 m depth.

potentiometric titration on a Metrohm 672 titroprocessor. The pH of the water was measured with an accuracy of ± 0.01 using an Orion pH meter combined with a Rossprobe. Direct current measurements were performed by the way of an AANDERAA current profiler between 0 and 500 m depth.

Hydropol data were used to reconstruct the circulation pattern, both zonally and along meridians. In October 1989, equatorial circulation (Fig. 2) was active, a rapid Equatorial Current (up to 60 cm s^{-1}) sustaining a powerful upwelling (surface temperature <25°C). In the tropical zone, south of the Marquesas Archipelago (10°S), currents were weaker with maximum values around 10 cm s^{-1} (Fig. 3). The westward component, representing the South Equatorial Current, was only present north of 14°S in April and October 1986. In 1988 and 1989, western flows were detected south of 15°S, consisting of a broad South Equatorial Current dissected by multiple countercurrents each measuring 100–200 km in longitudinal extent. These eastward countercurrents can be given the general term of South Equatorial Countercurrent (Kessler & Taft, 1987). The Marquesas zone (8–12°S) was touched by small eastward flows, running erratically from one semester to the other. Except during the 1987 ENSO year, it was difficult to define a real 'Marquesas countercurrent', supposed to follow the line of the SPCZ in the austral winter. These representations of zonal currents by calculation of the dynamic heights can perhaps minimize real drift values, but reveal the existence of whirlpools on small and medium scales (mesoscales eddies). The alternating of eastward/westward flows in this zone, and their longitudinal shift from one cruise to another gives a clear indication of the regional dynamic variability. Meridianal components showed a dominance of southerly flows, indicating a global drift towards the southwest, thus locating the western boundary of the Central South Pacific gyre in a latitudinal band 15–25°S. South of the Tropic of Capricorn, easterly flows were observed to feed the subtropical eastward drift (ED) that constitutes the southern branch of the south Pacific gyre (Levitus, 1982).

During the 1987 ENSO event, zonal currents showed intense interannual variability (Fig. 4). In March, the South Equatorial Current was cut into three sections centred around 13, 18 and 21°S. In October, at the end of this anomalous episode (which began in October 1986), eastward flow (South Equatorial Countercurrent) was dominant between the Marquesas (10°S) and Australes (22°S) archipelagoes, with speeds up to 8 cm s^{-1}. The South Equatorial Current was reduced to a weak and shallow flow between 15 and 19°S. In contrast, westward drift was stronger south of the Tropic of Capricorn. So, although mainly an equatorial phenomenon, ENSO events tend to increase the instability of the south tropical current field, which is already weak and erratic.

The physico-chemical properties of the water masses

The thermal field of the French Polynesian sector is characterized by temperatures around 22–24°C in the south–east sector closest to the centre of the gyre, and by temperatures above 27°C in the north–west sector (Rougerie *et al.*, 1985). Salinity correlates directly with regional evaporation: salinity values above 36 psu were found east of Tahiti and in the Tuamotu's and values below 35.5 west of 155°W (Delcroix & Henin, 1989).

Figure 5 shows cross-sections of the water column measured during the HYDROPOL 8 cruise (October, 1989) along longitude 140°W, from the Marquesas (5°S) to the Australes (28°S) through the central part of the Tuamotu's. The vertical thermal field (Fig. 5(a)) was

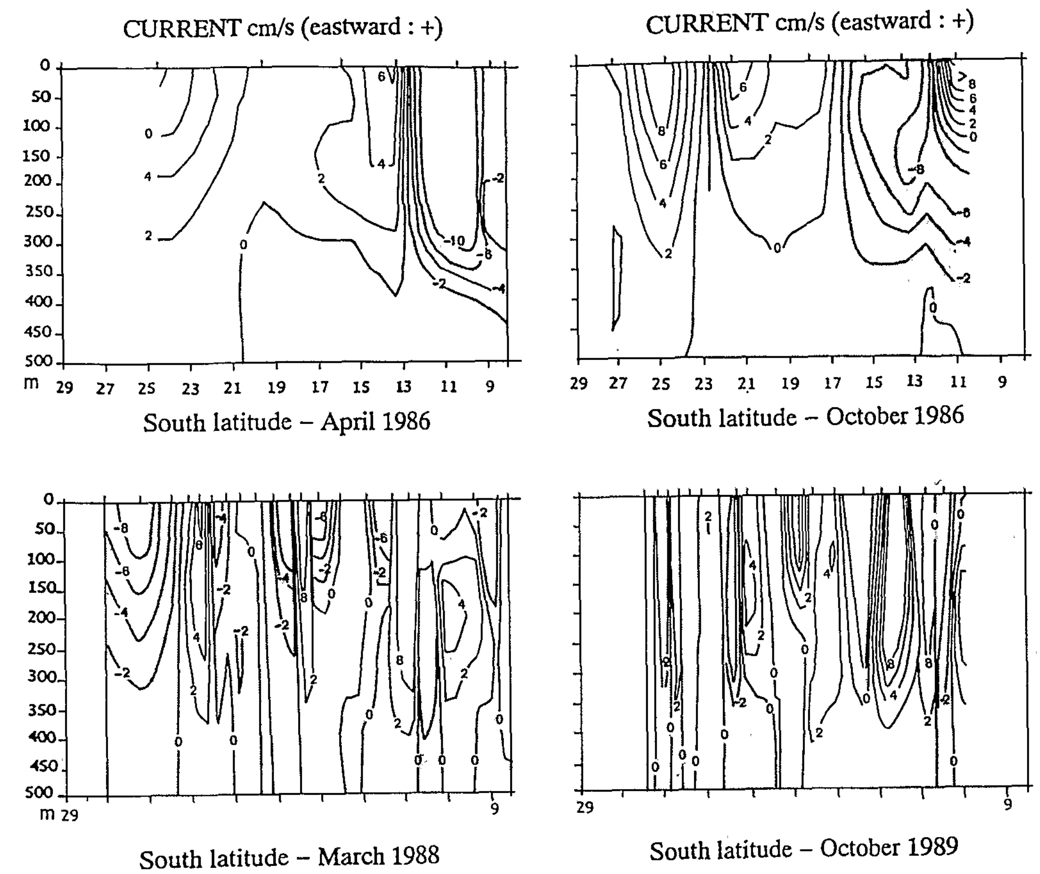

Fig. 3 Zonal components of currents, computed with the dynamical method; 1986–1989 HYDROPOL cruises, in the 140–145°W band (grey surface: westward). Low speed values (maximal westward component is 10 cm s^{-1}) and high interannual variability reflect existence of short and meso-scale eddies.

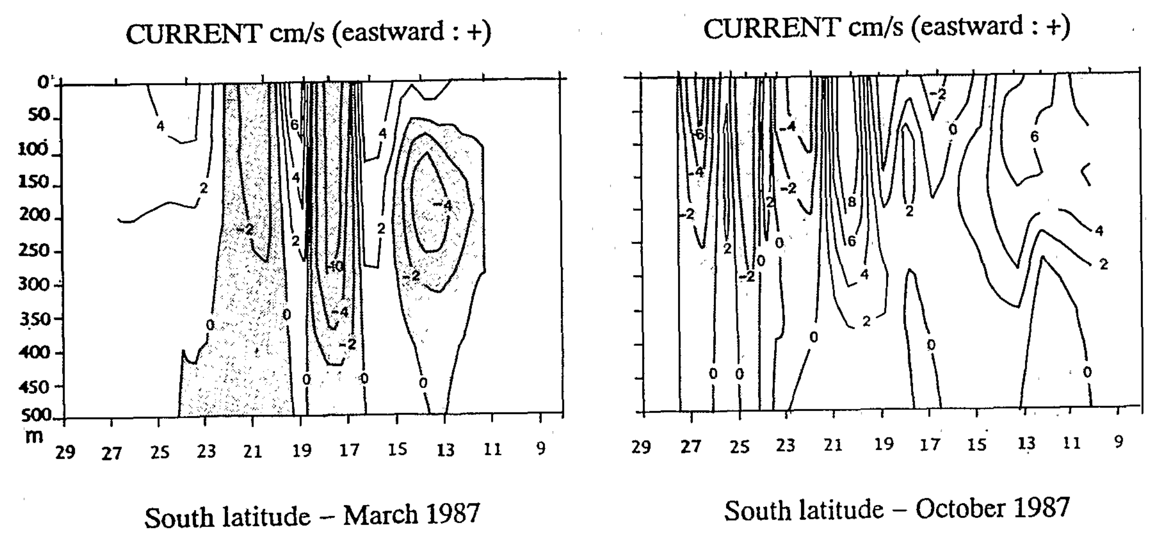

Fig. 4 Zonal currents in ENSO year 1987. In March the South Equatorial Current (grey surface) is cut in three parts; in October (end of the anomaly) eastward flow is dominant in the 9–22°S band and the South Equatorial Current, weak and shallow, is only running between 15 and 19°S.

characterized by the high thermal content of the 0–100 m mixed layer, through the southern tropic. The thermocline was deep (150–400 m) and widespread. Salinity (Fig. 5(c)) was greater than 36 psu in the 12–23°S zone and less than 35.6 north of the Marquesas and south of the Tropic of Capricorn. Density profiles (Fig. 5(b)) indicated strong stratification and the presence of a deep pycnocline between 200 and 500 m in depth. In the central zone (15–22°S), the surface mixed layer 0–150 m was isohaline at around 36.2 ± 0.2, which indicates the surface formation of a salty water mass and its sinking until it reaches hydrostatic equilibrium ($\sigma t = 25\pm0.5$) around 150–200 m. Southern Tropical Water, formed at the surface in the arid region of the Tuamotu (between 140 and 125°W),

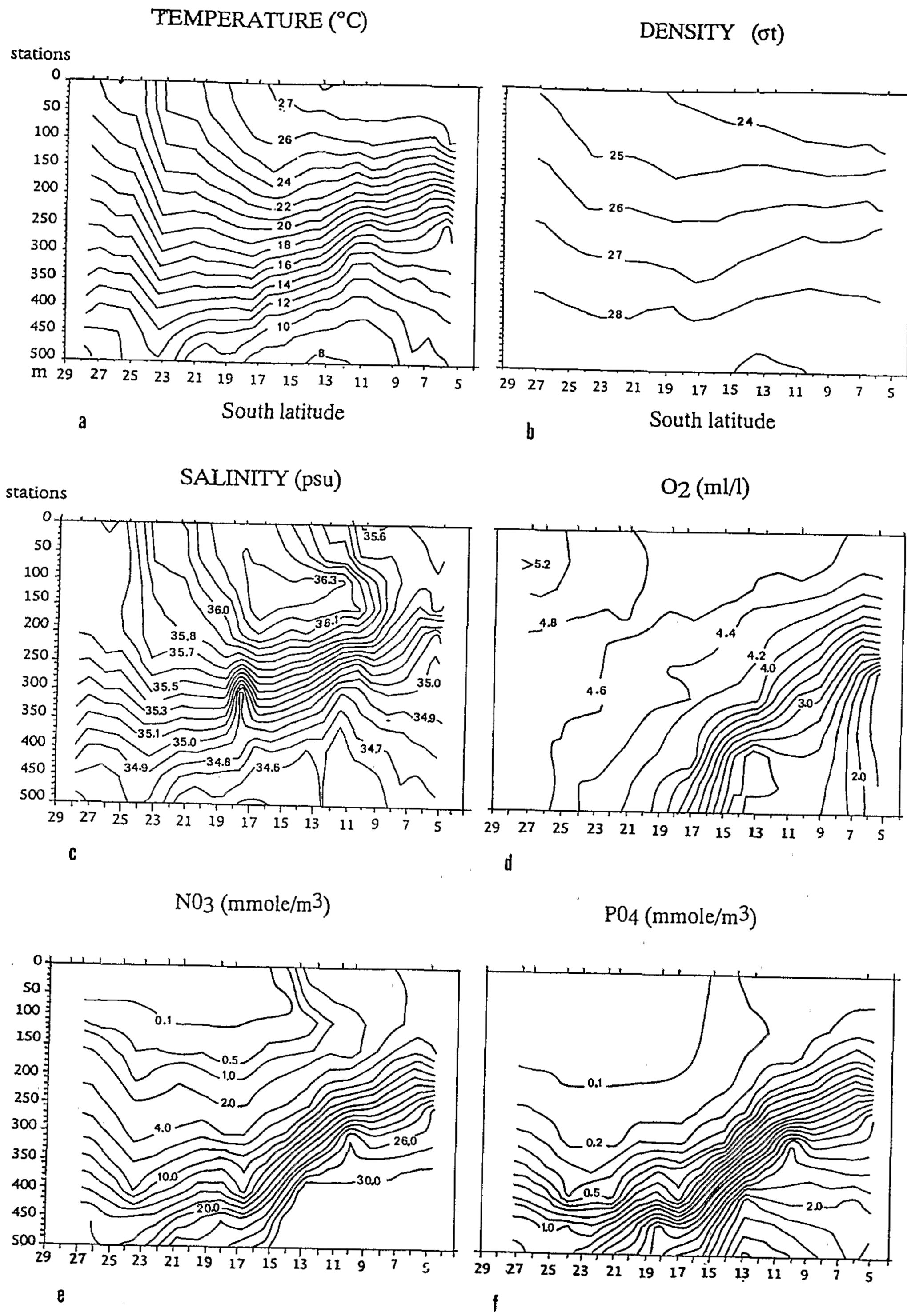

Fig. 5 Vertical transections along meridian band 140–145°W crossing through Marquesas, Tuamotu and Australes archipelagoes. N.O. *Marara* cruise HYDROPOL 3. October 1986. (a) Temperature (°C); (b) density (σt); (c) salinity (psu); (d) dissolved oxygen (ml l^{-1}); (e) phosphate inorganic dissolved (μM); (f) nitrate inorganic dissolved (μM). The thermohaline stratification marks the separation between the warm, salty surface mixed layer (0–200 m) and the cold, less saline AIW (>500 m). Dissolved oxygen is at saturation level south of 15°S. Surface mixed layer is depleted in phosphate and nitrate, especially in the tropical gyre belt (15–30°S).

is the saltiest of the entire Pacific and its subsurface circulation towards the west-northwest brings it into the equatorial New Guinean zone, where it is partially included in the ascending process through the equatorial upwelling (Wauthy, 1986).

Below 500 m depth, low salinity of 34.5 psu characterizes the Antarctic Intermediate Water (AIW). This AIW (temperature <8°C, 1–1.5 km thick) originates near the polar ice cap, sinks in the sub-antarctic convergences that occur between 50 and 60°S. This dense cold water moves northwards (direct branch) and eastwards (Levitus, 1985).

The vertical distribution of parameters controlling primary production shows that to the south of the 15°S parallel, the isoplethes of nitrate and phosphate descend deeply (Fig. 5(e), (f)), revealing the presence of a 0–150 m oceanic layer very poor in nutrients ($PO_4<0.1$, $NO_3<0.2$ μM). The other hydrological parameters, oxygen (Fig. 5(d)) and pH, are correlated inversely with the low nutritional status of the euphotic layer; dissolved oxygen content was above saturation and pH = 8.3. These values only changed below 250 m where strong vertical gradients in nutrients were directed between the mixed layer and the deeper AIW. The surface mixed layer, relatively isothermal and isohaline, extends to 150–200 m throughout the Tuamotu–Australes zone including the whole euphotic layer. The oligotrophy of this layer is a direct consequence of the permanent thermal stratification which isolates it efficiently from the AIW rich in nutrients. The separation between these two layers reaches its maximum at the latitude of the Tropic of Capricorn. The formation of the Southern Tropical Water by evaporation near the centre of the anticyclonic high and its descent (downwelling) to the pycnocline represents a dynamic process that reinforces the oligotrophy of the surface layer as planktonic organisms and particulate matter are advected down below the euphotic layer. The low particulate levels of this water are indicated by chlorophyll-*a* levels below 0.05 mg m^{-3} south of 15°S; the waters of the tropical zone which bathe the Tuamotu, Society and Australes archipelagoes are among the most oligotrophic and clear waters of the Pacific Ocean. Data gathered in the 1987 ENSO year indicate that this oligotrophic state is maintained and that nutrient-depleted warm waters tend to invade the equatorial surface zone. Even during these anomalous events, the pycnocline/nutricline remained deep and there was no significant tilting of the isoplethes, as observed downward in oriental (Peru) (Hansen, 1990) or upward in occidental boundaries at the same time (Wyrtki, 1977).

From the Open Ocean to Barrier and Atoll Reefs

In the oceanic domain, the atolls of the five Polynesian archipelagoes constitute a weak and porous network of resistance and present little obstacle to the mainstream current; the elevated islands are only a few thousand km^2 (1000 km^2 for Tahiti, the largest in French Polynesia); the largest atolls of the Tuamotu (Rangiroa, Fakarava) reach 1500 km^2 at sea-level (all emergent surfaces), but the majority do not exceed 300 km^2 with some limited to less than 10 km^2. All these islands and atolls have similar geomorphological characteristics: the external slope consists of a living coral barrier reef with extremely rich biota between 0 and 70 m, extending into a detrital slope of biocemented limestone called the impermeable apron. This limestone, with an average slope of 45°, reaches the basaltic sub-basement between a few hundred metres to a kilometre deep (based on profiles of reefs in the Tuamotu). The atolls have a very steep drop-off that is not accompanied by variations of ocean colour or transparency, explaining the fear of ancient navigators who named the Tuamotu's 'the dangerous achipelago'.

The water column around the barrier reefs and atolls

Numerous data have been collected around the barrier reefs of the Society Islands and the atolls of the Tuamotu archipelago during the past decade, especially with the R.V. *Tainui*, *Coriolis* and *Marara*. These data are summarized as follows:

1. Around Tahiti barrier reef. The hydrology and dynamic of oceanic water around the barrier reef of the north sector of Tahiti (149°30′W, 17°30′S) have been intensively studied in 1982–1984 (Kessler & Monbet, 1984, 1985), by use of permanent moorings fitted with current meters and thermistor chains and by hydrological casts and samplings. The thermohaline properties of this off-reef oceanic sector (0–3 km from the island) have not revealed any particular differences when compared with the deep ocean pattern. However, the continuous recording of temperature clearly shows thermal oscillations of 1–2.5°C around the mean value. These oscillations are observed at all depths monitored, between the surficial mixed layer and 1 km (Fig. 6). The record of current velocities, taken at the same time and depth, shows that the dominant NE–SW component has a periodic reversal centred around 6 h; this signal seems to be closely related to the tide signal which around Tahiti is semi-diurnal, with 0.3 m of tide range during spring tide and 0 during neaps, due to the proximity of an amphidromic point (Bongers & Wyrtki, 1987). The thermal oscillations are higher in the 200–500 m range, that is, at the depths where the thermocline spread. The apparent linkage between the cycles of current reversal and tidal oscillations make it possible to recognize more than a simple diurnal heating effect on the interpretations involving the generation of thermal oscillations. It is probable that internal waves are generated along the steep flanks of the island and, accordingly, might produce a periodic deformation of the thermohaline pattern (Tomczak, 1988). Of related concern is the possible effect of these internal waves on the stratification of the tropical ocean; this matter is important because any vertical advection or upsurge of deep (nutrient-rich) water can have significant positive effects on the productivity of the euphotic layer. When plotted on the temperature–depth curve, a thermal oscillation of 2°C corresponds in the water involved, to an oscillation of 30–50 m in vertical amplitude; exceptionally large thermal extremes of 4–5°C could reflect oscillations of 100 m, which mean for example that a body of water at 300 m depth would oscillate between 250 and 350 m. Regardless of their real amplitudes, these oscillations–internal waves do not mean that there is a resulting net displacement upward or downward of the water involved.

It is possible to test this by monitoring the nutrient distribution, both vertically and spatially, as done by Kessler & Monbet (1984). Figure 7 shows that there is no modification of the oligotrophy along an ocean reef flank section as proved by the distribution of phosphate, silicate and nitrate and no thermal cooling, as proved by the associated thermal field. The great depth of the thermocline-nutricline, below 250–300 m, is a permanent obstacle against any deep water upsurge. Our conclusion is that

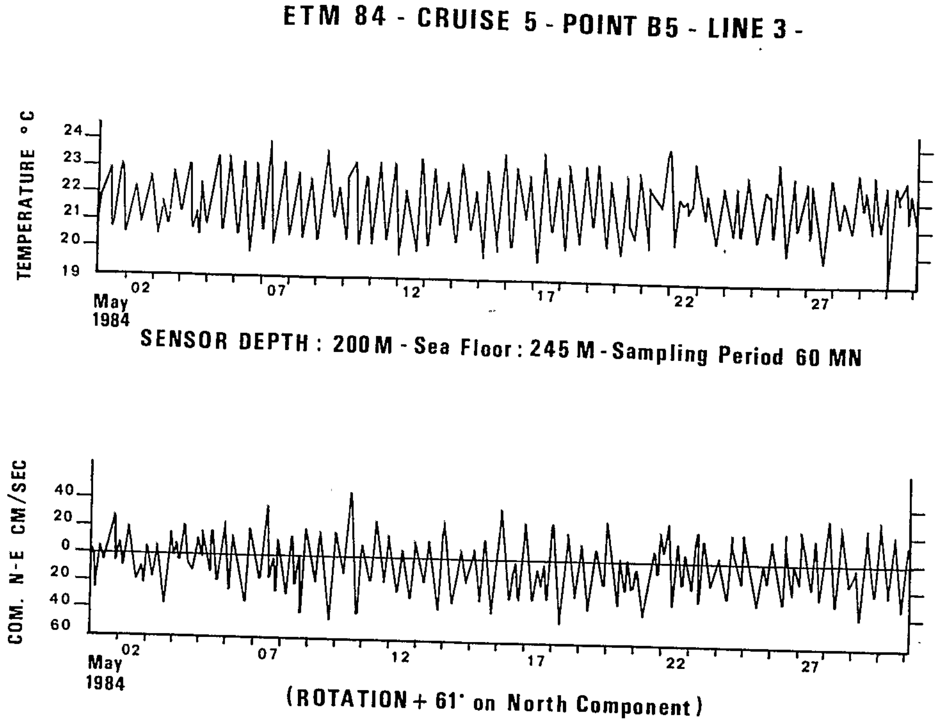

Fig. 6 Thermal oscillations and northeast/southwest current component along the barrier reef of Tahiti in May 1984. Mean temperature is 21.8°C with minima at 20.5°C and maxima at 23.0°C (95% of the signal). Mean current is 14 cm s^{-1} (northeast component) with minima at 5 cm s^{-1} and maxima 35 cm s^{-1} (95% of the signal) (from Kessler & Monbet, 1985, p. 29). Thermal and current oscillations are correlated and controlled by the tide regime whose variability on all time-scales is very low, owing to the proximity of an amphidromic point (Bongers & Wyrtki, 1987).

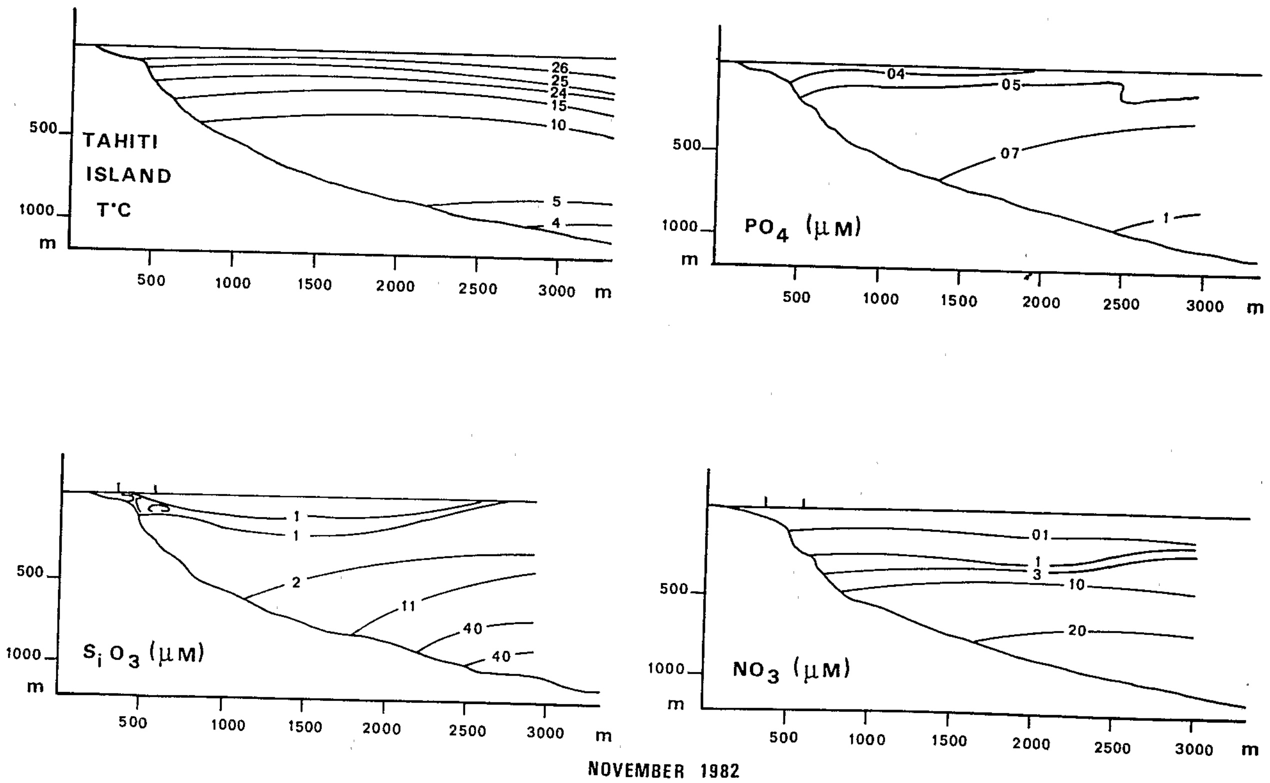

Fig. 7 Distribution of T°C, phosphate, silicate and nitrate along a cross section Tahiti barrier reef—open ocean in November 1982. The euphotic mixed layer (0–150 m) which bathes the algo-coral reef (0–60 m) has the same permanent oligotrophy as offshore oceanic waters (from Kessler & Monbet, 1984, p. 469).

if thermal oscillations–internal waves are a reality along the Tahitian barrier reef, there is no evidence that they can modify the permanent stratification–oligotrophy of the ocean. Our decade-long survey on the same barrier reef oceanic sector has never revealed any form of perturbation of the euphotic layer: it has always, including during typhoons (1983 and 1991) kept its main features, there being oligotrophy, high clarity and high temperature (from 26°C in winter to 29°C in summer). The only episodes when the vicinity of the reef showed some evidence of turbidity were closely related to extreme rainfalls; floods from swollen rivers injected large quantities of terrigenic and organic loads in the lagoon, which in turn expelled its turbid waters into the ocean through the passes. Lagoons of Tahiti are also more and more affected by pollution and eutrophication owing to anthropogenic activities with the well-known consequences (Naim, 1993) that coral colonies and patch reefs tend to be choked and killed by filamentous and macro-algae.

2. Around Tuamotu atolls. The water column above the reef slope has been studied from 5–10 m to depths of 1 km which occur within 1–3 km of the atoll barrier reef. These studies have shown no perturbation of the oceanic field by the mass of the atolls, even in the shallowest water column above the reef slope. Figure 8 presents a transect from ocean to lagoonal waters via the pass based on information collected in November 1985 near three atolls with deep passes. Figure 8 shows that the oceanic waters conserve their characteristic thermohaline properties up to the slopes of the atoll. At low tide, a small effect of the discharge of lagoon waters, which are slightly less dense (22.9 compared with 23.1 in the ocean), was observed but only close to the pass. Profiles of dissolved nutrients confirms that the oligotrophic status of the mixed layer is maintained right to the reef slope; the 0–200 m layer is devoid of phosphates and nitrates with the nutricline starting at 250 m, well below the saline layer (>36.0 psu) of the Southern Tropical Water and also below the euphotic layer. Lagoon water is slightly less depleted in nutrients than the oceanic euphotic layer and is characterized by a larger variability in these values (0.2±0.1 µM in dissolved phosphates and nitrates).

The water column around the atolls also retains high levels of dissolved oxygen (4.4 ml^{-1} on average in the core of the euphotic layer), equivalent to saturation. Lagoon water is also very well-oxygenated with values fluctuating between 100 and 120% of saturation. Higher levels of chlorophyll were observed in lagoons than in the ocean, but integration by area yielded equivalent numbers (15 mg m^{-2} in the upper 200 m of the ocean vs 12–18 mg m^{-2} in the upper 40 m of the lagoon). The attenuation of light at 660 nm (Zaneveld light-meter coupled with a Bissett–Berman probe) is significantly greater in the lagoons (At=560±10) than in the ocean (At=440), which indicates a greater lagoonal particulate load.

The thermohaline stratification observed around reefs is maintained all year, even in the subtropical zone (23–28°S) where winter temperatures are 6°C lower. Figure 9 shows close agreement between the vertical thermal profile and the cesium 137 profile near Mururoa Atoll. Both were relatively constant in the 0–200 m layer which was separated from the AIW (beyond 500 m) by a steep density gradient. The persistence of the pycnocline is therefore indicated, which suggests that no processes of vertical advection occur near or around atolls that are adequate to disturb the oceanic stratification. The vertical profiles of nutrients down the slopes of the Mururoa atoll confirm the absence of any diapycnal process across the euphotic–oligotrophic zone. Collectively, our data show no 'island mass effect' on the isoplethes close to these atolls, or near the other high islands of French Polynesia with the notable exception of the Marquesas which are surrounded by a drowned barrier reef (Rougerie *et al.*, 1992).

Expressed in terms of Reynolds number (ratio of inertial to viscous forces which is a function of current velocity), Andrews & Pickard (1990) showed that the formation of wakes or vortex eddies around reefs requires very swift flows, generally only found in passes or small gaps between adjacent reefs. Similar conclusions were reached by Wolanski *et al.* (1984) in their study of island wakes in shallow coastal waters of the Great Barrier Reef. For oceanic atolls, like those forming the Tuamotu Archipelago, the oceanic current field seems insufficient to produce high Reynolds numbers and noticeable stirring. Heywood *et al.* (1990) anoted that when the incident flows are weak (<20 cm s^{-1}) and fluctuating, eddies are not observed, and there is no upsurge or doming of the isopycnals. To observe the latter in the lee of islands or atolls, requires a strong and steady flow (more than 50 cm s^{-1}). Correlative biological enrichment is only possible if the nutricline is pushed upward by such a doming. In the Polynesian tropical central zone where the nutricline is very deep, such doming has never been observed and the separation between the euphotic-oligotrophic layer and the AIW appears to be a permanent feature. This pattern is maintained right up to the slopes of the atolls where the clear, blue water is legendary; coral–algal symbioses thus receive enough light energy for healthy coral growth to 60–80 m depth. As experimentally established (by Ledwell *et al.*, 1993), diapycnal mixing can be neglected in the dynamics model of the thermocline, which implies that heat, salt and tracers can penetrate the thermocline mostly by transport along, rather than across, density surfaces. The close similarity we found in Polynesian waters in the spatial and vertical distribution of the thermocline and nutricline barriers seems a good demonstration of that.

Enrichment processes: equatorial upwelling and reef endo-upwelling

The Tuamotu Archipelago is extended toward the northwest to the equator by the atolls of the Line Islands. The oceanic environment of the latter atolls is different from those of the Tuamotu's because there is evidence of the doming of the isopletes towards the surface nearer to the equator (Fig. 5). This phenomena is a direct result of vertical movements that affect the water mass, there being divergence and upwelling in the equatorial zone, convergence and downwelling in the 12–30°S band. In the Tuamotu's, the nutricline is always deeper than the

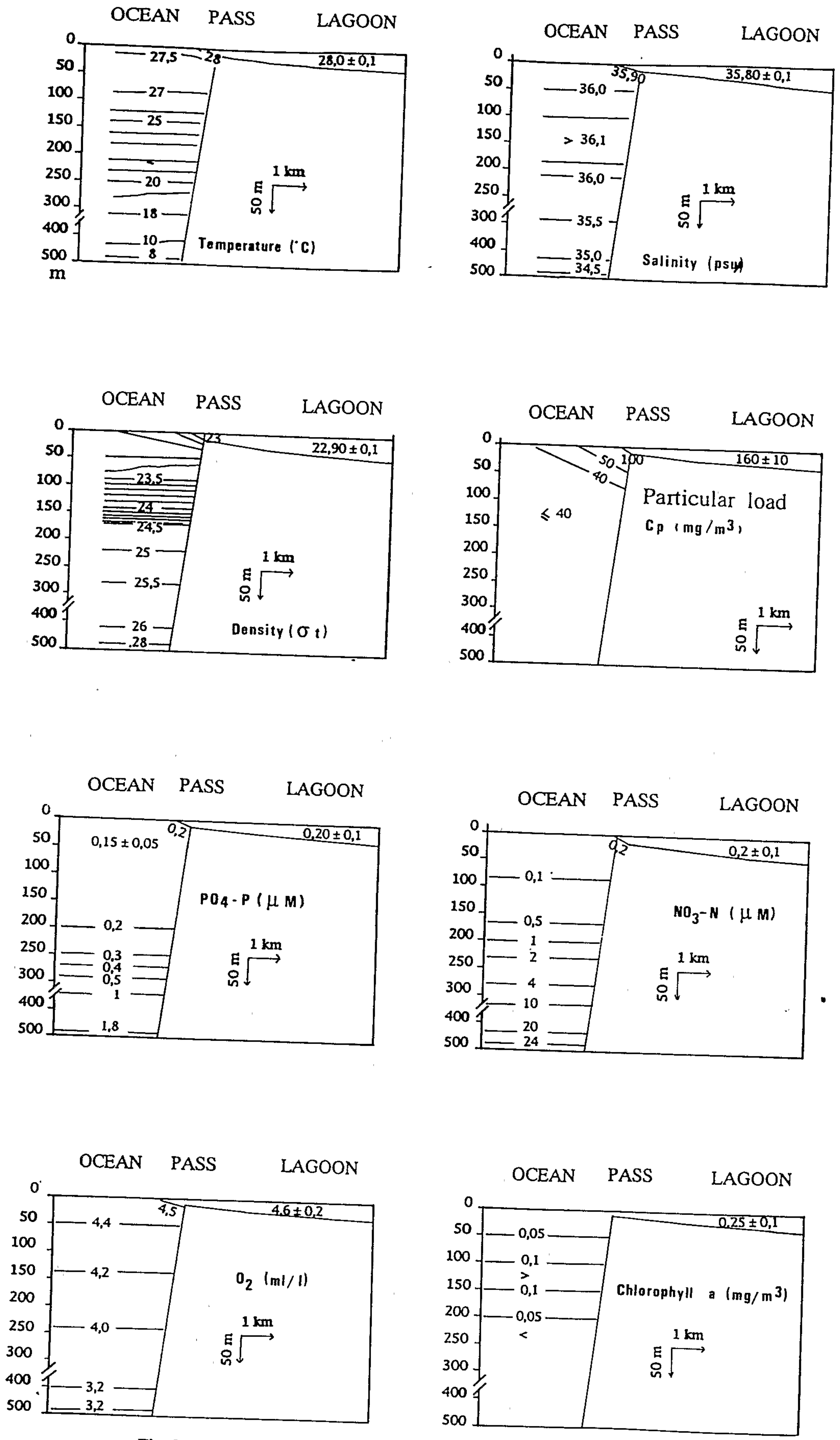

Fig. 8 Cross-sections: ocean–pass–lagoon. Average of data from the Tuamotu zone 15–16°S and 150–146°W which includes atolls of Tikehau, Rangiroa and Toau; R.V. *Coriolis* cruise TATU, November 1985. Oceanic thermohaline stratification and nutrients oligotrophy are maintained in atolls vicinity and above reef flanks. No 'atoll mass effect' can be detected, owing to the permanence of density stratification and weak oceanic advection (low Reynolds number).

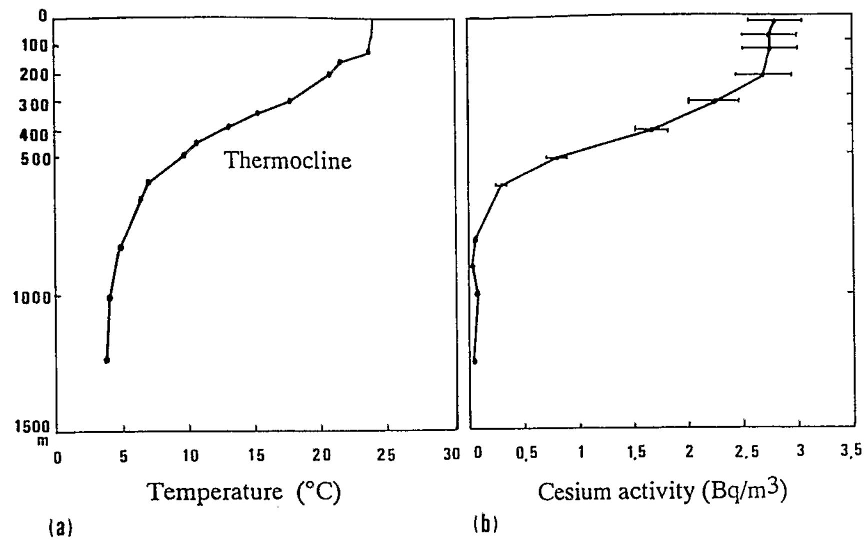

Fig. 9 Oceanic properties offshore Mururoa atoll (southeast Tuamotu, 22°S–139°W). (a) Vertical profile of temperature; (b) vertical profile of cesium activity (Bourlat *et al.*, 1991). Both profiles show the same cline separating a warm surface layer (0–100 m) from a deep AIW (beyond 500 m depth), confirming the absence of any vertical advection process, either up or down.

euphotic zone. However, the situation changes as one approaches the equator; the Malden (4°S) and Christmas, (2°N) Atolls situated in the 155–160°W longitudinal band are influenced by oceanic waters enriched by the equatorial upwelling (Bender & McPhaden, 1990). Along the equator, the availability of nitrates in the 0–20 m layer declines from 10 µM near the Galapagos islands to 6–8 µM at 140°W (North of the Marquesas) and to 1 µM to the West of 170°E (Le Bouteiller & Blanchot, 1991). In this equatorial band, chlorophyll-*a* values average around 0.25 mg m^{-3} with peaks up to 0.4 mg m^{-3}, and are thus 5–10 times higher than in the southern tropical zone. Paradoxically, observations on atolls situated in this upwelling zone (Kiribati, Christmas) have not indicated that their coral communities are very productive or healthy. On the contrary, the increased nutrients trigger blooms of phytoplankton that reduce the transparency of the water with adverse effects on the efficiency of the coral–zooxanthellae symbiosis. The sensitivity of this symbiosis to light penetration may explain why the atolls and islands of the clear and oligotrophic tropical zone are the most productive and have well-developed barrier reefs. The question of nutrient input, needed to sustain the high productivity/calcification level of these barrier reefs, is then raised. This paradox, first noted by Darwin, may be answered by the geothermal endo-upwelling model. In this hypothesis (Rougerie & Wauthy, 1993), deep oceanic waters (AIW) penetrate into the base of the porous coralline structures to rise through thermal convection to the top of the reef, bringing the new nutrients necessary to sustain growth in an otherwise nutrient limited system. Ocean waves play a mechanical role in this process by rinsing and cleaning the upper reef, thus preventing the plugging of the interstitial network to the flow of endo-upwelled nutrients that feed the living algo-coral veneer (Rougerie & Wauthy, 1992).

Although one might imagine, in highly dynamic zones with strong eddies, diapycnal or upwardly moving processes bringing subsurface nutrient rich water into the euphotic zone (Wolanski *et al.*, 1986), we have never observed them, either directly or indirectly. It is obvious that if enormous internal waves were able to lift the thermocline–nutricline by more than 100 or 200 m, they could locally contribute to enrichment of the base of the euphotic layer, with a boost to its primary production. However, instead of being helped by that enrichment, the autotrophic coral community will suffer from it, owing to rapid invasion of choking macro-algae, over predation by fish and borers and mainly by the limitation of sunlight penetration which controls zooxanthellae growth. This nutrient paradox has been convincingly demonstrated by Hallock & Schlager (1986) and Hallock (1988); we believe and observe that the consequences to reefs described by these authors, apply to the Polynesian ocean. Conversely, where and when upwellings or local enrichment–eutrophication do occur in tropical areas, such as along the coast of Peru, California, West African Coast, etcetera, there are no coral reefs; these upwelled nutrient-rich water are green, turbid and very productive, and anchovies have replaced corals.

Highlights

The Central Tropical South Pacific constitutes a vast oceanic zone in which the parameters controlling the hydroclimatic pattern are both interactive and episodically erratic, as during ENSO events. The large anticyclonic gyre associated with the Easter Island–Kermadec high pressure cell is a zone of high evaporation where the saline Tropical South Water is formed. The sinking of this water creates a downwelling that enhances seston loss from the euphotic layer and prevents any upwards migration of nutrients. Both factors contribute to the oligotrophic status of the surface mixed layer of the tropical gyre. The Equatorial Current, is rapid (60 cm s^{-1}) and undergoes a divergence along the equatorial line that triggers a vertical advection. This upwelling brings new nutrients into the euphotic layer and constitutes the main enrichment process of the intertropical zone. The South Equatorial Current is weaker (5–10 cm s^{-1}), albeit with a southern component that grows stronger with increasing latitude. South of the Tropic of Capricorn eastward drift dominates and constitutes the southern branch of the gyre. Countercurrents follow zones of wind convergence and bring warm waters accumulated at western boundaries back towards the Central Pacific.

Counter currents are much stronger during ENSO events, both in speed and breadth; they constitute an aperiodic but significant shift in the circulation pattern of the central and eastern Pacific. The meaning of ENSO anomalies must be assessed both in terms of the redistribution of water masses and of their biological impacts, there being cessation of equatorial upwelling and reduced equatorial fertility, coral bleaching, erratic dispersion of larvae, etcetera. During the last decade ENSO events have occurred during almost half of the time (1982–1983; 1986–1987; 1991–1993) which raises the question of whether it is accurate to use the term 'anomaly' to qualify ENSO episodes.

The weak and irregular flows running in the south tropical zone, (10°S to 29°S) seem insufficient to produce high Reynolds numbers even along the flanks of Tahiti Island where observed currents are 14 cm s^{-1}. Consequently, no turbulent vertical mixing or isopycnal doming can be expected in the lee of Polynesian islands or atolls that would be sufficient to stir the nutricline and to have a significant effect on productivity within the mixed layer. Even in the case of exceptional flow accelerations produced by strong west winds or typhoons, the disturbances generated in the thermohaline stratification do not modify the properties of the thick oligotrophic mixed layer. This result is consistent with observations in the Indian South Equatorial Current where no 'atoll mass effect' was detectable when surface advection was less than 30 cm s^{-1} (Heywood *et al.*, 1990).

Close to the reef flanks, thermal oscillations and current reversals are driven by the tide regime, which displays a very weak variability on all time scales considered. The cyclic perturbations or internal waves produced have no impact on the vertical nutrient distribution: the sole nutrient reservoir consists of the AIW, below 500 m depth. Oceanic thermohaline stratification is maintained in the vicinity of barrier reefs and atolls which are steep, narrow obstacles which represent little resistance to oceanic currents. Internally, however, the coralline matrix is porous and permeable. Both features have important consequences and dominate 'in reef' micro-scale processes including biogeochemical diagenesis and biogeochemical chemical fluxes across the reef–ocean interface. In the endo-upwelling model, the abrupt change in topography from the reef crest to the seaward slope is viewed as the geomorphological response by the coral–algal ecosystem set up by a thermo-convective/low-energy hydrothermalism process, that brings up new nutrients (Rougerie & Wauthy, 1993). The permanent low levels of nutrients and chlorophyll in the surrounding ocean results in high transparency that permits optimal light penetration needed for the symbiotic micro-algae in the corals to thrive. The only slight exceptions to this pattern are linked to lagoons, reef gaps and passes where wind or tidal driven outflows can create dynamic jets with high particulate and organic loads.

The central south Pacific constitutes a purely maritime province where oceanic mass transport is driven at three different scales:

- large scale, represented by the vast south Pacific anticyclonic gyre;
- meso-scale, illustrated by the high variability of zonal components of currents and the presence of fluctuating eddies;
- small-scale, with vortices and turbulences around islands and atolls during very active wind conditions. These turbulences and stirrings seem to have negligible biological impact in the euphotic layer, however, owing to the great depth of the pycnocline.

The permanent decoupling between the oligotrophic euphotic layer and the nutrient-rich deep layer (AIW) is the most outstanding feature of the South Pacific tropical ocean. Barrier and atoll reefs here are constantly surrounded by clear, unproductive, warm water, that paradoxically, constitutes an ideal environment for their growth and long-term survival.

Most of the results discussed in this paper were obtained during cruises of the R.V. *Tainui* (1983), R.V. *Coriolis* (1982 and 1985) and R.V. *Marara* (1986–1989). These cruises reflect a co-operative programme conducted by IFREMER then by SMSR/CEA/DIRCEN and ORSTOM. The captains and crews of these research vessels are acknowledged for their support at sea during these cruises.

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