The frontal area at the eastern edge of the western equatorial Pacific warm pool in April 2001

Gérard Eldin and Thierry Delcroix
Laboratoire d’Etudes en Géophysique et Océanographie Spatiales, CNES, CNRS, IRD, Université Paul Sabatier, Toulouse, France

Martine Rodier
Institut de Recherche pour le Développement, Nouméa, New Caledonia

Received 8 August 2003; revised 5 March 2004; accepted 9 April 2004; published 3 July 2004.

[1] The western part of the equatorial Pacific Ocean is characterized, on average, by a relatively warm, fresh, and oligotrophic near-surface water mass, the warm pool. On its eastern side the warm pool joins cooler, saltier, and richer waters from the equatorial upwelling. Previous works have shown that the boundary between these two water masses is highly variable in zonal location and structure, depending on large-scale climatic conditions and/or local atmospheric forcing. This paper reports results from an oceanographic cruise carried out in April 2001, dedicated to the study of that boundary. By combining physical and biogeochemical data, the frontal area could unambiguously be found close to 157.5°E, in the far-western equatorial Pacific, in accordance with La Niña climatic conditions. Some characteristics of the frontal zone were similar to the scarce previous observations: a sharp \( pCO_2 \) step of about 40 ppm and a variation of NO3 concentration from about 1 \( \mu M \) to complete depletion. Other features were not previously reported, particularly the unexpected lack of a noticeable surface salinity front. Additional data from the Tropical Atmosphere-Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON) moorings array provide a tentative explanation for that lack of salinity gradient: It appears that at the time of our observations the eastern part of the warm pool presented anomalously high surface salinity, about 0.5 above normal. The local evaporation-precipitation budget did not explain this high salinity. Rather, it was associated with an anomalous northward extension of the South Pacific high-salinity tongue in the upper thermocline.


1. Introduction

[2] The western part of the equatorial Pacific Ocean is characterized, on average, by relatively warm and fresh near-surface water masses, forming the warm pool (also called fresh pool). Surface waters in that area are also characterized by a relatively low partial pressure of CO2, \( (pCO_2) \) [Feely et al., 2002; Le Borgne et al., 2002]; nutrient depletion and low chlorophyll concentration in surface and low mesozooplankton biomass are also typical of the area [Mackey et al., 1995; Le Borgne and Rodier, 1997]. The warm pool is a huge heat reservoir, at the origin the strongest atmospheric convection area of the world ocean [Webster and Lukas, 1992]. At the surface its eastern edge is roughly delimited by the 28°C isotherm and/or the 35 isohaline, and it is subject to zonal displacements of several thousand kilometers, eastward or westward, during mature phases of El Niño or La Niña, respectively. During a strong El Niño the warm pool can reach the South American coast, while during La Niña it may be confined to the west of 150°E [Fu et al., 1986; Picaut et al., 1996]. That spatial variability and the resulting changes in the zonal extension of the warm waters are deeply involved in the ocean-atmosphere coupling processes of El Niño-Southern Oscillation (ENSO) events. In addition, by modifying the area of low \( pCO_2 \), that variability will affect the global ocean-atmosphere carbon exchange budget [Feely et al., 2002].

[3] Except during strong El Niño events, when equatorial upwelling is completely absent, the eastern edge of the warm pool joins the western edge of the equatorial upwelling area of the central/eastern Pacific. This region often presents sharp frontal structures in surface salinity [Kuroda and McPhaden, 1993; Eldin et al., 1997; Hénin et al., 1998].
and $p$CO$_2$ [Inoue et al., 1996], as well as in nutrients and primary production [Le Borgne and Rodier, 1997; Stoens et al., 1999]. These frontal structures mostly result from the zonal convergence of western Pacific and central Pacific waters maintained by the average flows of the equatorial current system [Picaut et al., 2001]. That zonal convergence plays an important part in sustaining the so-called salinity barrier layer [Lukas and Lindstrom, 1991] that dynamically isolates near-surface waters from thermocline waters and thus favors zonal displacements of the warm pool [Vialard and Delecluse, 1998].

[4] A better knowledge of the structure and dynamics of that frontal zone will clearly help in better understanding processes associated with the warm pool zonal migration and extension, as well as in potentially improving ENSO prediction lead time [Clarke and Van Gorder, 2001]. It is worth noting that remote observations of the frontal zone are difficult, as surface fields derived from satellites (e.g., sea surface temperature, seawater color, and altimetry) do not usually present sharp variations at the front. In addition, horizontal or temporal resolution of these remotely sensed data is often too coarse, and/or intense cloud coverage associated with warm waters precludes reliable observations. In situ observations are therefore essential.

[5] To date, in situ observations of the frontal zone are relatively few. Lemasson and Piton [1968] provided some information on the warm pool edge between 160$^°$E and 170$^°$E during the ALIZE cruise in 1965 but with limited spatial resolution. A density front in February 1991 was present at 165$^°$E–170$^°$E [Eldin et al., 1992], and Kuroda and McPhaden [1993] described the frontal zone at 170$^°$E–180$^°$E in January–February 1990, with still low zonal resolution. In January–February 1994 the Japanese KY9401 cruise crossed the front, but, unfortunately, data are lacking around its location [see Feely et al., 2002]. A more precise description of the eastern edge of the warm pool was obtained during the Flux dans l’Ouest du Pacifique Équatorial (FLUPAC) cruise in October 1994, with a complex zonal structure, showing a physical front with a salinity change of 0.4; variability often occurring on less than 5 km; sea surface temperature, seawater color, and altimetry) do not usually present sharp variations at the front. In addition, horizontal or temporal resolution of these remotely sensed data is often too coarse, and/or intense cloud coverage associated with warm waters precludes reliable observations. In situ observations are therefore essential.

[6] The FRONTALIS cruise of R/V Alis of Institut de Recherche pour le Développement departed Nouméa, New Caledonia, on 29 March 2001. The cruise (Figure 1) began with a meridional section along 165$^°$E, including partial maintenance of some moorings of the Tropical Atmosphere-Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON) array at 8$^°$S, 5$^°$S, 2$^°$S, and 0$^°$N. It was followed by a zonal section westward along the equator, with the purpose of determining the longitude of the front through monitoring of surface data acquisition. Sampling of the front through two “butterfly” navigation patterns was scheduled, but the far-western position of the frontal area did not allow us to complete those patterns, given the limited ship autonomy. Thus a unique meridional section could be achieved along 157.5$^°$E, followed by a southward meridional section along 165$^°$E toward Nouméa. The cruise ended on 25 April 2001. This study mostly uses data acquired along part of that cruise track, between 5$^°$S and 5$^°$N.

[7] To allow detection of frontal characteristics at the surface, several data sets were acquired underway and displayed on board in real time all along the cruise track. The FRONTALIS cruise was carried out in March–April 2001 in order to provide a relatively high resolution description of the frontal area at the eastern edge of the warm pool, including physics and biogeochemical properties. This paper presents and discusses observations from that cruise. Section 2 describes data collection and analyses. Section 3 presents a description of the quasi-synoptic physical and biogeochemical structures encountered in the vicinity of the frontal zone. Section 4 provides an interpretation of these observations, placing them in the large-scale frame of climatic conditions and equatorial Pacific circulation.

2. Data and Methods

[7] The FRONTALIS cruise of R/V Alis of Institut de Recherche pour le Développement departed Nouméa, New Caledonia, on 29 March 2001. The cruise (Figure 1) began with a meridional section along 165$^°$E, including partial maintenance of some moorings of the Tropical Atmosphere-Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON) array at 8$^°$S, 5$^°$S, 2$^°$S, and 0$^°$N. It was followed by a zonal section westward along the equator, with the purpose of determining the longitude of the front through monitoring of surface data acquisition. Sampling of the front through two “butterfly” navigation patterns was scheduled, but the far-western position of the frontal area did not allow us to complete those patterns, given the limited ship autonomy. Thus a unique meridional section could be achieved along 157.5$^°$E, followed by a southward meridional section along 165$^°$E toward Nouméa. The cruise ended on 25 April 2001. This study mostly uses data acquired along part of that cruise track, between 5$^°$S and 5$^°$N.
Sea surface temperature (SST) and salinity (SSS) were continuously recorded by a Sea-Bird model SBE-16 thermosalinograph, providing 1 min averaged records corresponding to a spatial resolution of about 0.3 km. At stations surrounding the front, surface drifters were launched to provide Lagrangian estimates of SST as well as surface currents to show drifter convergence toward the front position. CO₂ mole fractions were measured with a Li-COR 6262 infrared analyzer linked to a shipboard equilibrator following the method described by Dandonneau [1995]. Mole fractions were converted to partial pressure of CO₂ in seawater (pCO₂) using ambient atmospheric pressure and SST.

[10] Hydrographic stations were carried out every degree along most sections, but sampling was increased to every half degree along 157.5°E (2.5°S–2.5°N) and 165°E, 5°S–0°N on the northward leg and 5°N–5°S on the southward leg (except for nutrients). Conductivity-temperature-depth (CTD) casts to 500 m were done with a Sea-Bird 911+ CTD model and a 12 L Niskin bottles Carousel® sampler. Temperature and conductivity sensors were calibrated at the factory after the cruise, as well as thermosalinograph sensors. Acoustic Doppler current profiler (ADCP) profiles every 5 min were obtained within 0–250 m from an RDI BB-150 vessel-mounted model. Raw data were processed using the CODAS3 software [Bahr et al., 1990]. Additional processing after the cruise included corrections of gyro drifts using GPS attitude data from a SERCEL NR-130 receiver. Nutrients (nitrate, nitrite, and phosphate) and chlorophyll a samples were collected to 200 m from the Niskin bottles. The nutrient analyses were conducted immediately on board with a Bran+Luebbe auto-analyzer 3 (A3), following procedures described by Oudot and Montel [1988] and Oudot and Baurand [1994]. Detection limits for NO₃, NO₂, and PO₄ were 0.02 μM, 0.02 μM, and 0.05 μM, respectively. Chlorophyll a (Chl a) samples were pressure-filtered through Whatman GF/F filters and stored at −4°C. Analyses were carried out postcruise in Nouméa by fluorometry on methanol extracts as described by Le Bouteiller et al. [1992], using a Turner Design TD-700 fluorometer. Ash-free dry weights were computed from mesozooplankton samples (>200 μm) taken by a triple WP2 net on 0–200 m at some stations [Le Borgne and Rodier, 1997].

### 3. Cruise Observations

#### 3.1. Equatorial Section

[10] The equatorial section was carried out from 165°E to 155°E on 5–9 April 2001 to obtain a small-scale description of surface and subsurface structures associated with the transition from equatorial upwelling waters to warm pool waters. ADCP data (Figure 2a) show the westward flow of the South Equatorial Current (SEC) above 150 m, with maximum westward flow of more than 50 cm s⁻¹ at 164°E. Surprisingly, a large area of eastward current reaching 40 cm s⁻¹ is embedded in the SEC, centered at 100 m depth, from 158°E to 163°E. At 155°E–157°E the current is weak or reversed in the surface but is above 30 cm s⁻¹ at 150 m depth. The Equatorial Undercurrent (EUC) is well defined below 150 m, with a velocity core centered at 200–250 m of average value 70 cm s⁻¹ and local maxima above 80 cm s⁻¹. The meridional velocity section shows that areas of strongest SEC around 164°E and 155°E–157°E are rather northwestern. In the EUC, meridional components are weak and variable.

[11] Temperature and salinity sections from 0–300 m are presented in Figure 2b. This limited depth range is appropriate because we are here mainly interested in structures and variability in and above the thermocline. The thermocline (centered on the 20°C isotherm) is found around 200 m, in the upper part of the EUC. From 165°E westward, average temperature of the upper layer increases, associated with a progressive deepening of the 28.5°C isotherm (except at 162°E–163°E, where a 0.2°–0.3°C cooler area is found). West of 158°E this increase accelerates, and SST exceeds 30°C at 155°E–156°E, with the 29°C isotherm reaching as deep as 150 m. Surface layer temperatures are
close to climatology (Reynolds SST data from National Oceanographic and Atmospheric Administration/Cooperative Institute for Research in Environmental Sciences (NOAA/CIRES)), while the thermocline is about 15–20 m deeper than average [Kessler and McCreary, 1993]. Salinity above the thermocline shows a maximum above 35.6 at 164°E–165°E through the whole upper layer and progressively decreases westward, faster in surface than at depth. This leads to vertical salinity stratification, as well as a slight zonal SSS gradient at 155°E–160°E, from 35.1 to more than 35.4. A vertical salinity gradient at the base of the quasi-homogeneous temperature layer hints at the presence of a salinity barrier layer to the west. Two vertical profiles of temperature, salinity, and density are presented in Figure 3 to illustrate the differences in vertical stratification: In the east (160°E) the weak vertical density gradient is almost constant, from the surface to the pycnocline, as often found in upwelling waters. In the west (156°E) a 50 m well-mixed surface layer is found, above a step-like increase in density associated with both cooler and saltier waters. The vertical profiles at 156°E are not exactly similar to the canonical barrier layer structure where a salinity gradient would appear within an isothermal layer. However, it shows a dynamical decoupling between near-surface water and the thermocline, a characteristic of the warm pool. In the EUC, salinity variability is caused by fluctuations of the strong meridional salinity gradient at that depth [e.g., Eldin et al., 1997].

[12] Nutrient sections are shown in Figure 4. They present zonal gradients of properties comparable to the hydrology sections. NO₃ concentrations above the thermocline vary from 2.5 μM in the east to values of less than 0.1 μM, close to the detection limit, indicating NO₃ depletion at the westernmost end of the section on a 40 m thick surface layer. In the nutrient-rich area, zonal gradient of nutrient concentration is not monotonic: A local maximum of 3–4 μM is found at 162°E–163°E around 100 m depth. NO₂ presents a subsurface maximum along the whole section east of 158°E, with an interruption at 164°E. This maximum of more than 1.4 μM is at 100 m depth at 165°E, then shrinks and slightly deepens to 140 m at 158°E. West of 158°E the subsurface maximum disappears, and surface NO₂ is below the limit of detection. This distribution is evidence of strong remineralization, and thus relatively high primary production, east of 158°E, with the exception of the 162°E–164°E area, where

Figure 3. Vertical profiles of CTD-derived temperature (solid line), salinity (short-dashed line), and density excess (long-dashed line) at the equator, 160°E (7 April 2001) and 156°E (8 April 2001).

Figure 4. Nutrients and chlorophyll a concentrations from stations along the equatorial section, 5–9 April 2001. Sampling is shown by crosses. The 0.1 μM NO₃ and NO₂ isolines are dashed.
minima in surface NO$_3$ as well as in the NO$_2$ maximum are present. The zonal distribution of phosphate is quite similar to nitrate, but concentrations always remain at significant levels (>0.1 $\mu$M) even in the west. These different features are common outside of the equatorial upwelling tongue [Mackey et al., 1995; Murray et al., 1995; Raimbault et al., 1999]: Where nutrient supply is scarce, nitrogen becomes limiting more rapidly than phosphorus and is thus more quickly depleted. At 162°E–164°E, PO$_4$ surface concentrations also show a relative minimum.

[13] Chl $a$ concentrations (Figure 4) present a subsurface maximum of more than 0.3 $\mu$g L$^{-1}$ between 40 and 80 m depth along most of the section. The layer of high Chl $a$ (>0.2 $\mu$g L$^{-1}$) is deepest at 164°E–165°E; then it reaches the surface at 162°E–163°E and progressively deepens westward.

[14] In summary, interpretations of physical and biogeochemical data from the equatorial section converge: Two different surface water masses were sampled, with a boundary somewhere between 157°E–158°E. East of that boundary, relatively cool and salty waters within 0–150 m, with surface NO$_3$, remineralization shown by the NO$_2$ subsurface maximum, and with a Chl $a$ maximum layer closer to the surface are all indicative of a water mass originating from equatorial upwelling and presenting features of a high-nutrient, low-chlorophyll (HNLC) regime [Le Borgne et al., 2002]. Inside that zone, small-scale variability is evident, with richer and more productive waters at 161°E–163°E and slightly warmer and NO$_3$-depleted waters at 164°E. West of that boundary, warmer and slightly fresher waters within 0–50 m, the existence of a barrier layer, no N-nutrients in surface, and a slightly deeper Chl $a$ maximum are characteristics of warm pool waters.

3.2. Surface Data and Integrated Biomass, 165°E–155°E

[15] The four surface drifters launched on the equator (two at 165°E, one at 157.5°E, and one at 155°E) had a lifespan of more than 3 months, after being deployed between 6 and 18 April 2001. Their trajectories as well as recorded SSTs are presented in Figure 5. They all show similar drift patterns: first, a westward drift for 4–8 weeks, then an eastward drift after a zonal current reversal somewhere between 150°E and 160°E longitude. This behavior is consistent with the idea that this area presents a zonal convergence of currents, as surmised from trajectories of hypothetical drifters displaced by modeled currents [Picaut et al., 2001]. In agreement with these model results, our observations corroborate that the convergence zone is a main characteristic of the boundary between upwelling and warm pool waters. In conformity with CTD data of Figure 2, SST increases from east to west on the equator. From 150°E to 160°E, small-scale space-time variability is evidenced, with amplitudes of a few tenths of a degree Celsius.

[16] Continuously recorded data are now used in order to obtain a high-resolution description of horizontal property gradients in the surface layer, hence complementing the 1° resolution hydrography and nutrient sampling. Figure 6 allows a comparison of these data sets along the equator and, additionally, presents results of zooplankton biomass estimates. In Figure 6 (top), SST increases by 0.7°C from 29.2°C to 29.9°C in 50 km, around 157.5°E, with the sharpest gradient at 157.70°E (about 0.3°C in 3 km). Unlike SST, SSS does not present a rapid variation but rather a continuous decrease west of the same longitude. The pCO$_2$ shows a sharp decrease of about 40 ppm in the same area (strongest gradient of 4 ppm in 1 km at 157.36°E) from oversaturated values toward values closer to equilibrium with the atmosphere. The 0–40 m averaged NO$_3$ presents a small zonal gradient in the upwelling area, then a rapid decrease of order 1 $\mu$M at 157.5°E. The 0–40 m integrated Chl $a$ fluctuates in the upwelling area, then decreases below 6 mg m$^{-2}$ west of the frontal area. Mesozooplankton biomass (only five net hauls available) shows lower values west of 157.5°E (about twice as low). Some of these parameters present the same frontal characteristics reported from previous measurements at the boundary between warm pool and upwelling waters [Rodier et al., 2000; Kobayashi and Takahashi, 2002], with an estimated front at 157.5°E longitude: amplitude and sharpness of the pCO$_2$ drop, variation of integrated NO$_3$, Chl $a$, and zooplankton biomass. The two main differences with other observations are
the lack of a salinity front and the relatively strong and sharp SST increase, which is of importance to climate, given the strong sensitivity of atmospheric convection to SST warmer than 28°C–29°C.

[17] The above analysis of observations along the equator is completed by a plot of surface data and Chl a vertical distribution along the 157.5°E meridional section (Figure 7). Meridional frontal features in SST and pCO₂ are found centered exactly at 0.00°N, with slightly smaller amplitudes than on the zonal section but similar maximum small-scale gradients (0.2°C in 1 km and 10 ppm in 3 km, respectively). SSS does not vary sharply at 157.5°E–0°N but, nonetheless, presents slightly higher values south of the equator. North of about 2°N, low SSS and low pCO₂ values are linked with the North Equatorial Counter Current (NECC) flow. Strong small-scale variability at these latitudes in the three surface parameters is associated with heavy localized rainfall. Averaged NO₃ in the surface layer increases to about 2 µM at 0.5°N–1°N and then decreases to less than 0.2 µM north of 2°N, in agreement with SSS and pCO₂ decreases. Integrated Chl a does not vary much from 1.5°S to 2.5°N, from 9 to 13 mg m⁻². However, its vertical distribution is more revealing: South of the equator a well-defined deep Chl a maximum is present at 30–60 m, as well as minima in the surface, characteristics of an oligotrophic regime; at 0° N–2°N the vertical Chl a distribution is more homogeneous, closer to an HNLC regime, and north of 2°N, hints of a deep maximum are found again.

[18] Contour plots of surface data obtained along the whole cruise track within 3°N–3°S provide more information on the complex spatial structure of the frontal zone (Figure 8). The SST pattern shows a southeast/ northwest orientation, crossing the equator slightly east of 157.5°E. The large-scale pattern of surface isotherms is confirmed by weekly Reynolds SST data from NOAA-CIRES (not shown). SSS distribution around 157.5°E is complicated, with a maximum in the southeast of more than 35.3 and weak zonal and meridional surface gradients, almost orthogonal to the SST gradient west of 157.5°E. The strong SSS gradient at 1°N–2°N confirms the presence of NECC waters hinted at in Figure 7. The pCO₂ maximum gradient is located slightly west of the SST maximum gradient. These complex spatial distribu-

![Figure 6](image-url)  
**Figure 6.** Data along the equatorial section, 5–9 April 2001. (top to bottom) 1 min resolution data from thermosalinograph (SST and SSS) and CO₂ partial pressure recorder (pCO₂), 0–40 m averaged NO₃ (asterisks) and 0–40 m integrated Chl a (triangles) from bottle sampling, and 0–200 m ash-free dry weight (AFDW) of mesozooplankton biomass at five stations (stars). Dashed line shows estimated zonal location of the front.

![Figure 7](image-url)  
**Figure 7.** Meridional section along 157.5°E, 10–12 April 2001. (top to bottom) Surface physical data, as in Figure 6; integrated chlorophyll a (triangles) and averaged NO₃ (asterisks); and vertical distribution of chlorophyll a, contour interval 0.05 mg m⁻³. The dashed line shows the meridional location of the frontal zone, south of the upwelling tongue. The short-and-long-dashed line is the approximate northern limit of the tongue.
tions, as well as probable short-term variability, explain the small differences in frontal amplitudes and locations observed in Figures 7 and 8 on the zonal and meridional sections (60 hours time between the two samplings at 157.5°E

[19] At 161°E–163°E, slightly cooler SST, higher SSS, and $pCO_2$ (Figures 6 and 8) confirm the presence of a locally more intense upwelling area. East of 163°E a higher SST, lower SSS, and lower $pCO_2$ area extends from the equator to at least 3°S along 165°E. These different water properties are associated with northward meridional flow on the equator (Figure 2), and thus probably advected from the south.

[20] Surface data as well as vertical distribution of physical and biogeochemical parameters converge toward a coherent description of the frontal area: It appears that upwelling waters formed a wedge that obliquely penetrated the oligotrophic waters of the warm pool. This wedge was bounded to the north by the NECC eastward flow. In the surface, while SST and $pCO_2$ spatial gradients are quite similar, with a northeast/southwest orientation, SSS variability around 157.5°E is weaker than what had been previously observed, and its spatial distribution is not apparently linked to SST distribution.

4. Discussion and Conclusion

[21] The most striking feature of our April 2001 observations is the lack of a sharp SSS front at the eastern edge of the warm pool, contrasting with the few previous observations (see section 1) which, in fact, had strongly motivated our cruise programming. Our data show SSS equatorial values (at 155°E–157.5°E) in the warm pool (Figure 6) of 35.1–35.3, well above expected values of less than 35. However, parameters like $pCO_2$, nutrients, and biology did show frontal variations comparable in amplitude to those previous observations. Another peculiar feature is the SEC reversal on the equator at 158°E–163°E. Data from the TAO/TRITON moorings array are used below to complement our observations, although the 10° longitude spacing of the moorings does not allow us to resolve the fine spatial scales.

[22] Along 156°E, TAO/TRITON provides 0–250 m salinity time series at 2°S, 0°N, and 2°N, in addition to the temperature series (salinity data courtesy of K. Ando and M. Kuroda, Japan Marine Science and Technology Center, Japan). The equatorial mooring time series at 156°E (Figure 9) shows alternative periods of relatively low and high salinities on the 0–100 m layer. In particular, a period of high salinity with SSS above 35.0 spans mid-January to the end of April 2001. This matches a general increase of salinity above the midthermocline (200 m), with maxima higher than 35.5 centered at 100 m in February 2001, and at about 120 m during and after the FRONTALIS cruise. Similar patches had appeared in May–June 2000.
before that, the time series presents too many gaps to provide useful information. No local or vertical physical processes can simply explain the appearance of such patches of maximum salinity at depth, so we believe they originate from horizontal advection. In fact, other time series of salinity at 156°E, at 100–120 m depth, 2°S and 2°N (not shown), indicate that the equatorial salinity maxima at depth were actually spanning the equator at the correct time to explain the features of Figure 9. The presence of a high-salinity layer at these depths in the western Pacific is a manifestation of the Southern Hemisphere high-salinity tongue (HST) extending from the southern Pacific and roughly centered on the 24.5 isopycnal [Delcroix et al., 1987]. This HST is evidenced in data from the second section at 165°E (Figure 10), where the 35.5 isohaline crosses the equator and reaches 2°N.

Although the HST is a fairly well known feature of the western Pacific, this is, to the best of our knowledge, the northernmost extension of the 35.5 isohaline ever observed along this regularly well sampled meridian (e.g., see Delcroix and Eldin [1995] for the 17 Survey of the Tropical Pacific (SURTROPAC) cruises carried out during 1984–1992). However, nearby cruise observations at 170°E in 1967 had shown a similar northward extension of the HST [Hisard et al., 1969]. Kessler [1999] has shown that the maximum salinity of the HST is increasing on a decadal timescale. Together with that increase in salinity, the tongue widens and extends northward, with temporary maximum extensions during La Niña years (1989, 1996, and 1999). Therefore it is tempting to hypothesize that the equatorial salinity increase around 100 m depth we observed at the end of the unusually long lasting 1999–2001 La Niña event at 156°E, 157.5°E, and 165°E was associated with these interannual and/or decadal changes in the HST.

This anomalous high salinity at depth is probably responsible for the observed anomalously high SSS of warm pool waters. A salinity increase above the main thermocline reinforces the vertical salinity gradient, which favors barrier layer formation, a feature which increases the decoupling between surface layers and the thermocline. In addition, because of the prevailing La Niña climatic conditions in April 2001 at the time of the cruise, moderate trade winds (0–5 m s\(^{-1}\)) had been continuously present over the warm pool for 3 months before the cruise, east of 155°E. QuikSCAT high-resolution satellite winds show that some occurrences of westerly events appear only to the west of that longitude (Figure 11). The dynamical response to the easterlies implies that equatorial upwelling was thus active in warm pool waters most of the time, but because of the strong barrier layer and the unusually deep thermocline during the ongoing La Niña it only acted upon the surface...
layer above 100 m (approximately above the 29°C isotherm and salinities of 35.3–35.5), as deduced from the time series of Figure 9. High-salinity water was then brought to the surface, but the upwelling could not reach deep enough to increase surface values of pCO₂ or nutrients, as pCO₂ increases only in thermocline waters, and the 1 μM NO₃ isoline was deep, at about 100 m. It is worth noting that SSS changes cannot be easily related to the local evaporation-precipitation budget: There is no apparent correlation between precipitation (Figure 11) and surface salinity (Figure 9) at 156°E–0°N; for instance, although February 2001 is the rainiest month of the period, SSS stays high (35.1–35.2) and only marginally decreases at the beginning of March.

Another interesting feature observed along the equatorial section deserves a tentative explanation. The 160°E–163°E region presents relatively stronger upwelling characteristics (Figures 2 and 4), but zonal current here was mostly eastward, with a subsurface maximum around 100 m, which appears contradictory. A time series from the closest TAO/TRITON mooring with current data, at 0°N–165°E, shows that occurrences of similar subsurface (about 100 m) eastward current patches were not at all uncommon in 2000–2001, with duration of 10 days to 1 month (Figure 12), in a time of prevailing easterly winds, that is, in November 2000, January 2001, and the beginning and end of March 2001; these patches are clearly different from those caused by westerly wind bursts (for instance, in mid-May and July 2001), which extend from the surface to the EUC, with higher velocities close to the surface. Current reversals within the isothermal layer in the area have been studied by Cronin et al. [2000] and have been linked to temporary reversals of the local zonal pressure gradient associated with changes in surface winds. In our case this process would imply an increase of easterly surface forcing during current reversals, which does not appear to be the case from the wind data of Figure 12.

Other processes, like transits of tropical instability waves (TIW) and/or remote effects, could be hypothesized. Although La Niña conditions are favorable to a westward extension of TIW fields [Eldin and Rodier, 2003], they do not seem to be involved here: Assuming that the current patch observed at the end of March 2001 at 165°E (Figure 12) is translated westward to lead to our current observations in early April around 161°E (Figure 2a) implies a westward TIW phase speed of 0.2–0.3 m s⁻¹, which is too slow [Qiao and Weisberg, 1995]. A better hint may be found in data from the 165°E section (Figure 10): The specific thermohaline structure at the base of the mixed layer (about 100 m), with slightly cooler and saltier water at 1°N–2°N than at the equator, leads to a dynamic height anomaly profile favorable to eastward flow at about 100 m depth north of the equator (Figure 10b), although exact computation of geostrophic velocities from a single section so close to the equator would be unrealistic. In fact, direct current measurements at 165°E (Figure 10a) show the presence of eastward flow as close as 30 km north of the equator at 60 m depth. Small meridional fluctuations of similar flows could lead them right to the equator and explain our observations of Figure 2a. Notice that these flows are associated with high-salinity waters of the HST, south of 3°N, and thus are not connected with the low-salinity eastward flow of the North Equatorial Counter Current north of 3°N (Figure 10). Similar patches of eastward flows associated with 35.2–35.4 salinity had already been mentioned by early investigators at 170°E [Hisard et al., 1969].

Finally, a combination of physical and biogeochemical measurements allowed us to refine the characteristics of the frontal zone between upwelling and warm pool waters. We have shown that some previously thought canonical features could not be observed, like a strong zonal SSS front, because of anomalously high SSS in the warm pool. Further studies of the boundary area are thus needed, including similar multidisciplinary observations, to clarify the mechanisms at work in this salinity increase of warm pool waters. A modeling effort is necessary to better understand potential consequences of such salinity variations. In particular, if decadal-scale phenomena affect salinity at the base of the surface mixed layer, as discussed above, their effects on barrier layer physics and possibly on ENSO-associated zonal displacements of the warm pool deserve to be thoroughly investigated.

Acknowledgments. Thanks to the captain and crew of R/V Alis, who worked tirelessly for the success of the FRONTALIS cruise. François Baurand, Francis Gallois, and David Varillon provided excellent technical and data collection support. Surface drifters were kindly provided by AOML (Miami, USA), and pCO₂ data could be obtained thanks to the infrared analyzer made available by Jean François Ternon. Christophe Maes was very helpful in trying to estimate the front position before the cruise. Fruitful discussions with Robert Le Borgne are acknowledged. This work was supported by the French Programme National d’Etude de la Dynamique du Climat (PNEDC) and l’Institut de Recherche pour le Développement (IRD).

References
