

Satellite sea surface temperature: a powerful tool for interpreting in situ $p\text{CO}_2$ measurements in the equatorial Pacific Ocean

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ABSTRACT

In order to determine the seasonal and interannual variability of the CO_2 released to the atmosphere from the equatorial Pacific, we have developed $p\text{CO}_2$ -temperature relationships based upon shipboard oceanic CO_2 partial pressure measurements, $p\text{CO}_2$, and satellite sea surface temperature, SST, measurements. We interpret the spatial variability in $p\text{CO}_2$ with the help of the SST imagery. In the eastern equatorial Pacific, at 5°S , $p\text{CO}_2$ variations of up to $100\ \mu\text{atm}$ are caused by undulations in the southern boundary of the equatorial upwelled waters. These undulations appear to be periodic with a phase and a wavelength comparable to tropical instability waves, TIW, observed at the northern boundary of the equatorial upwelling. Once the $p\text{CO}_2$ signature of the TIW is removed from the Alize II cruise measurements in January 1991, the equatorial $p\text{CO}_2$ data exhibit a diel cycle of about $10\ \mu\text{atm}$ with maximum values occurring at night. In the western equatorial Pacific, the variability in $p\text{CO}_2$ is primarily governed by the displacement of the boundary between warm pool waters, where air–sea CO_2 fluxes are weak, and equatorial upwelled waters which release high CO_2 fluxes to the atmosphere. We detect this boundary using satellite SST maps. East of the warm pool, ΔP is related to SST and SST anomalies. The 1985–1997 CO_2 flux is computed in a 5° wide latitudinal band as a combination of ΔP and CO_2 exchange coefficient, K , deduced from satellite wind speeds, U . It exhibits up to a factor 2 seasonal variation caused by K -seasonal variation and a large interannual variability, a factor 5 variation between 1987 and 1988. The interannual variability is primarily driven by displacements of the warm pool that makes the surface area of the outgassing region variable. The contribution of ΔP to the flux variability is about half the contribution of K . The mean CO_2 flux computed using either the Liss and Merlivat (1986) or the Wanninkhof (1992) K – U parametrization amounts to $0.11\ \text{GtC yr}^{-1}$ or to $0.18\ \text{GtC yr}^{-1}$, respectively. The error in the integrated flux, without taking into account the uncertainty on the K – U parametrization, is less than 31%.

1. Introduction

The equatorial Pacific is of primary interest for air–sea CO_2 exchange studies since it represents

the largest outgassing region in the global ocean and is subject to large interannual variations linked to El Niño and La Niña events (Feely et al., 1995; Wanninkhof et al., 1996; Boutin and Etcheto, 1997). Recently, several JGOFS cruises were conducted in this region. However, the ocean CO_2 partial pressure, $p\text{CO}_2$, measured during these

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cruises exhibits a strong spatial and temporal variability. Therefore, the extrapolation of such measurements is not straightforward and it is a challenge to determine from these highly variable in-situ $p\text{CO}_2$ measurements the air-sea CO_2 flux over the entire equatorial Pacific region. Because satellites offer a quasi permanent and synoptic view of the ocean, they provide a unique opportunity to interpret the observed $p\text{CO}_2$ variability and hence decide how to interpolate in-situ measurements.

In this paper, we use satellite imagery of sea surface temperature to help in the interpretation of spatial variability in $p\text{CO}_2$ measured during various cruises in the equatorial Pacific. Data are presented in Section 2. In Section 3 we focus on the $p\text{CO}_2$ variability in the eastern Pacific Ocean related to the presence of instability waves. In Section 4, we discuss the diel cycle of $p\text{CO}_2$ observed during Alize II cruise and finally in Section 5, we propose that variability in the central and western equatorial Pacific is related to the displacement of the warm pool boundary.

2. Data

We use underway $p\text{CO}_2$ measurements collected during the last 12 years. The date and location of most of these measurements have been described by Boutin and Etcheto (1997). In particular, measurements were made by research vessels during the Alize II cruise (Lefèvre and Dandonneau, 1992), several NOAA cruises (Murphy et al., 1994), the EqPac/OACES cruises (Feely et al., 1995; Wanninkhof et al., 1996), Japanese cruises (Inoue and Sugimura, 1992) and by merchant ships during the Lillooet cruises (Wong et al., 1993) and ECOA cruises (Dandonneau, 1995). In this current study, we also include measurements made in the eastern equatorial Pacific during the IronEx 2 cruise (Cooper et al., 1996) in May–June 1995, during ECOA cruises 8 to 20 which cross the equator near 100°W every 3 months from 1993 to 1997, during NOAA/OACES cruises (Wanninkhof et al., 1996; Ho et al., 1997; Masters et al., 1997) in April–May 1993, April–June 1994, August–September 1994 and November 1995, respectively, and made by a CARIOCA buoy in January 1997 (Etcheto et al., 1999). We also add measurements made in the central and western

equatorial Pacific during Olipac and Flupac cruises (LeBorgne et al., 1995; Metzl and Bres, 1996) in October–November 1994 and during Japanese cruises (Inoue et al., 1996) in November 1992, January–February 1994 and November–December 1994.

The $p\text{CO}_2$ measurements we use are all continuous surface measurements made at a temperature close to in situ temperature, thereby avoiding large uncertainties due to temperature correction (Lee et al., 1997). They have been performed by several research teams, most of them have participated in intercomparison experiments which demonstrate a reasonable agreement, 1 to 2 μatm , between $p\text{CO}_2$ measured by the different groups. We compute the air-sea CO_2 partial pressure gradient, ΔP , by taking the difference between $p\text{CO}_2$ and atmospheric CO_2 partial pressure, $p\text{CO}_2 \text{ atm}$. $p\text{CO}_2 \text{ atm}$ is either measured during the same cruise as $p\text{CO}_2$ or deduced from CO_2 mixing ratio in dry air (P. Ciais, personal communication) measured in NOAA/CMDL flasks at Christmas island (central Pacific, 2°S) and smoothed over the time domain using the CCGVU procedure (Thoning et al., 1989). In the latter case, $p\text{CO}_2 \text{ atm}$ is deduced from the CO_2 mixing ratio assuming a standard atmospheric pressure of 1013 mbar and taking the saturation vapor pressure over water corresponding to the local SST from Kennish (1989).

To interpret the spatial variability in these in situ measurements, we use high and low resolution sea surface temperature (SST) derived from the Advanced Very-High Resolution Radiometer, AVHRR, data. The low (1° and 1 week) resolution, NMC SST analyzed fields have been generated by the optimal interpolation of Reynolds and Smith (1994) from 1985 to 1997. These weekly fields have been smoothed in time using a 1/4, 1/2, 1/4 weighted average of three weekly SST maps, as recommended by Reynolds. The high (daily and 9 km) resolution SST images are the NOAA/NASA Oceans Pathfinder AVHRR product #091 (best SST product) obtained from the NASA Physical Oceanography Distributed Archive Center at the Jet Propulsion Laboratory, California Institute of Technology. The images shown in the paper are daily composites obtained by averaging day and night images. We use them to interpret $p\text{CO}_2$ measurements in April 1992, April 1994 and May 1995.

We also use in situ SST measured on the TOGA-TAO moorings located at 110°W, 5°S and 110°W, 2°S. Details concerning the instrument characteristics can be found in Hayes et al. (1991) and a description of the TOGA-TAO array is given in McPhaden (1993).

To compute air–sea CO₂ flux, we combine ΔP , with CO₂ exchange coefficient, K , deduced from satellite wind speed and from Reynolds SST using the Liss and Merlivat (1986) and the Wanninkhof (1992) relationships as described in Boutin and Etcheto (1997). We use satellite wind speed measurements at 10 m height above the sea level derived from the Geosat altimeter from April 1985 to January 1988, the Special Sensor Microwave Imager (SSM/I) microwave radiometer from February 1988 to July 1991, the European Radar Sensor (ERS) 1/Active Microwave Instrument (AMI) scatterometer from August 1991 to April 1996, the ERS2/AMI from May 1996 to September 1996 and the NASA Scatterometer (NSCAT) from October 1996 to June 1997. Geosat radar cross sections were inverted using the Witter and Chelton (1991) algorithm, the SSM/I wind speeds were derived by Wentz (1992) and the ERS/AMI wind speeds were produced by Ifremer/CERSAT at Brest (Quilfen and Bentamy, 1994). The processing of Geosat and SSM/I data has been described by Boutin and Etcheto (1997), the ERS/AMI processing has been described by Boutin and Etcheto (1995). Concerning NSCAT, we use the NSCAT 1 Science data product Level 2 wind speed at 50 km resolution (NASA_Scatterometer_Report, 1997). We only keep the best wind speed retrieval indicated as a wind vector cell quality flag equal to 0.

The various sets of satellite wind speeds agree rather well and we have checked that the variability in K discussed in Section 5 is reproduced by the various data sets during their overlapping periods. Boutin and Etcheto (1996) collocated Geosat and SSM/I F08 wind speeds, SSM/I F08 and SSM/I F10 wind speeds and ERS1 and SSM/I F10 wind speeds in the range 3–15 m s⁻¹, which is the wind speed range of interest for our study of K . The mean differences are always less than 0.3 m s⁻¹, the r.m.s. of the differences being less than 1.7 m s⁻¹, depending on the instruments. The scatterometer measurements are of much better quality than those performed by other instruments. We have collocated ERS1 and ERS2 wind

speeds in March 1996 and ERS2 and NSCAT wind speeds at the beginning of October 1997, in the same wind speed range. The mean differences are equal to 0.08 and -0.04 m s⁻¹ and the r.m.s. of the differences are equal to 0.85 and 0.91 m s⁻¹, respectively (Bucher, 1997).

3. Variability related to instability waves

In the eastern equatorial Pacific, tropical instability waves, TIW, are clearly visible on satellite SST images (Legeckis, 1977; Yoder et al., 1994) as westward propagating cusp shaped disturbances along the front between the cold equatorial upwelled waters and the warmer waters north of the equator. These waves have been extensively studied using in situ measurements as reviewed by Qiao and Weisberg (1995). They are created by dynamical instabilities in the zonal equatorial currents and they occur when the westward flow of the South Equatorial Current (SEC) is well developed. The site where they are generated remains unclear. Possible locations could be the shear zone between the westward SEC and the eastward North Equatorial Counter-Current (NECC), between the SEC and the northern flank of the eastward Equatorial Undercurrent or within the SEC itself (Qiao and Weisberg, 1995). Flament et al. (1996) even suggest that sometimes several of these processes could act to generate distinct equatorial and sub-equatorial instability waves.

The signature of these waves in terms of $p\text{CO}_2$ has been observed during the Alize II and EqPac Spring 92 equatorial cruises (Lefèvre and Dandonneau, 1992; Feely et al., 1994). South of the front, $p\text{CO}_2$ is high and the SST is relatively low, indicating recently upwelled water, whereas north of the front $p\text{CO}_2$ is close to the equilibrium with the atmosphere and the water is warmer.

Such waves were also propagating during ECOA 4 cruise which crossed the equator at 100°W on 20 June 1992, during the same season as EqPac 1992 Spring cruise. The cross equatorial track of ECOA 4 superimposed on the daily AVHRR composite of 20 June 1992, is presented in Fig. 1, top. The in situ $p\text{CO}_2$ and SST measurements are shown in Fig. 2 versus longitude. The ship crossed the crest of a TIW at 96°W, 1.5°N. This corresponds to an abrupt increase in $p\text{CO}_2$ of 50 μatm and a SST decrease of more than 1°C.

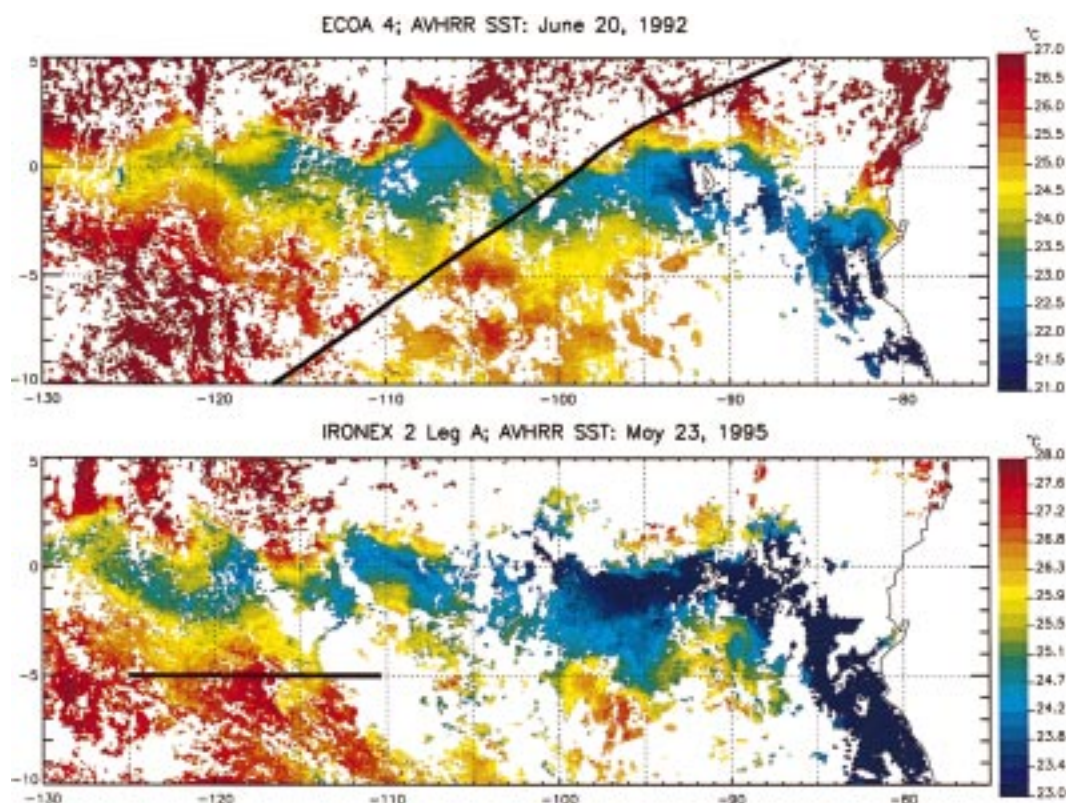


Fig. 1. SST images during ECOA 4 cruise (top) and IronEx 2 cruise (bottom). The track of the cruise is superimposed as a black line.

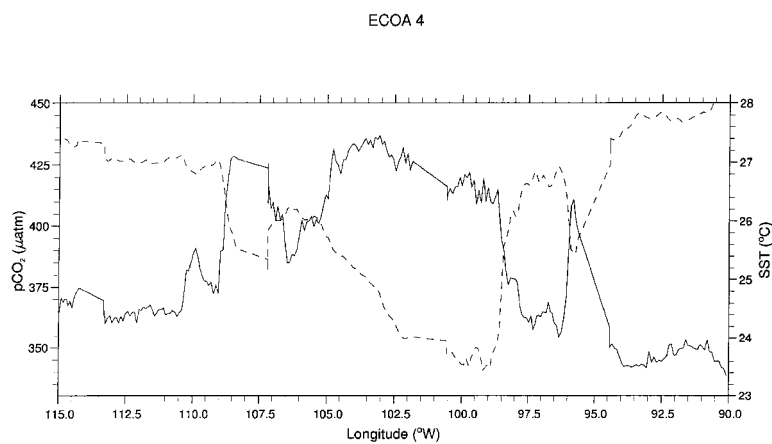


Fig. 2. $p\text{CO}_2$ (solid line) and SST (dashed line) measured during ECOA 4 between 19 and 22 June 1992, as a function of longitude. Increases of $p\text{CO}_2$ correlated with decreases of SST at 96°W and 107°W are due to the crossing of TIW.

This was followed by a $p\text{CO}_2$ decrease and a SST increase of almost the same magnitude when the ship exited the recently upwelled water. It is noticeable that the SST measured on the crest (25.5°C) is 1.5°C warmer than the SST measured in the equatorial upwelling south of the equator because waters of the crest were in the transition zone between the equatorial upwelled waters and the subtropical waters. When the ship entered the equatorial upwelling at 98.8°W , such a transition zone is observed between 98°W and 98.8°W and is characterized by strong SST and $p\text{CO}_2$ gradients. There, the SST of 25.5°C is associated with a $p\text{CO}_2$ value of $385\ \mu\text{atm}$, a lower value than the $405\ \mu\text{atm}$ observed at 95.9°W . This could be due to larger $p\text{CO}_2$ in the upwelled water at 95°W than at 99°W because of the presence of the local upwelling close to the Galapagos islands. The signature in SST of this local upwelling is clear in Fig. 1, top, around 92°W .

South of the equator, the variability in $p\text{CO}_2$ has been much less studied. On ECOA 4, the ship crosses a warm, low $p\text{CO}_2$, region between 105°W and 107°W (Fig. 2) which looks like part of a meander of the upwelled waters (Fig. 1, top). Such meanders in the southern boundary of the equatorial upwelled water have also been observed along zonal transects made at 5°S during the IronEx 2, cruise (Leg A, 21–25 May 1995) and the EqPac Spring 94 cruise (23–26 April 1994). The $p\text{CO}_2$ and SST measured outside the iron patch during IronEx 2 Leg A, are shown in Fig. 3, top. The $p\text{CO}_2$ data exhibit a huge variability (more than $100\ \mu\text{atm}$) which appears to be anticorrelated with the SST variability, in particular between 110°W and 117°W . On the corresponding SST composite (Fig. 1, bottom), the relatively cold water around 115°W and around 123°W appears to origin from the equatorial upwelling. The signature of the meander crossing is weaker in the SST signal than in that of the $p\text{CO}_2$ because SST changes by heat exchange with the atmosphere are much more rapid than changes in $p\text{CO}_2$ due to air–sea CO_2 exchange and/or biological activity (Etcheto et al., 1999). The same kind of variability was observed during the EqPac Spring 94 zonal transect at 5°S between 95°W and 110°W (Fig. 3, bottom). AVHRR composites during the cruise are not shown because they were too cloudy, especially along 5°S . However, the clear pixels between 1°S and 4°S , show a meander in the

southern boundary of the cold upwelled water which penetrates further south around 100°W and 108°W , where high $p\text{CO}_2$ values are observed (Fig. 3, bottom). The composites also show the presence of TIW north of the equator.

The SST measured on the TAO mooring at 110°W , 5°S in 1992 and 110°W , 2°S in 1994 are examined in order to complement the spatial information given by the satellite imagery (Fig. 1). These locations are chosen so as to sample the maximum SST variability related to the meanders in the southern boundary of the cold upwelled waters which were further south during ECOA 4 (Fig. 1, top) than during IronEx 2 (Fig. 1, bottom). The SST time series in 1992 (Fig. 4, top), shows oscillations superimposed on the SST seasonal cycle, between mid-June and November 1992 with a period of about 1 month; the same kind of oscillations appear between February 1995 and June 1995 (Fig. 4, bottom). The amplitude of the oscillations is about 2°C , i.e., the same order of magnitude as the spatial variability visible in the SST imagery. Although no measurement of surface currents is available at these latitudes, the SST time series confirm that meanders at the southern boundary of the upwelled water could be due to the presence of TIW south of the equator. The wave characteristics (period about 1 month and wavelength about 800 km as determined from the satellite imagery) are similar to those at the northern boundary and we always detect meanders in the southern boundary of the upwelled waters when TIW are present north of the equator. The signature of the waves observed south of the upwelling on SST imagery (Fig. 1) is more diffuse than north of the upwelling. This could be because the SEC is maximum north of the equator (Qiao and Weisberg, 1995).

The amplitude of the $p\text{CO}_2$ change across the southern front between equatorial upwelled water and subtropical water is variable from one year to another: about $40\ \mu\text{atm}$ in 1992, $75\ \mu\text{atm}$ in 1994 and $100\ \mu\text{atm}$ in 1995. This can be related (1) to weaker upwelling in the east Pacific in May–June 92 than in April–May 94 and April–May 1995 (Etcheto et al., 1999), and (2) to the different cruise tracks with respect to the shape of the meander.

In the 3 cases we have studied, meanders have been observed in spring, which is not the usual period during which the SEC accelerates. Southern

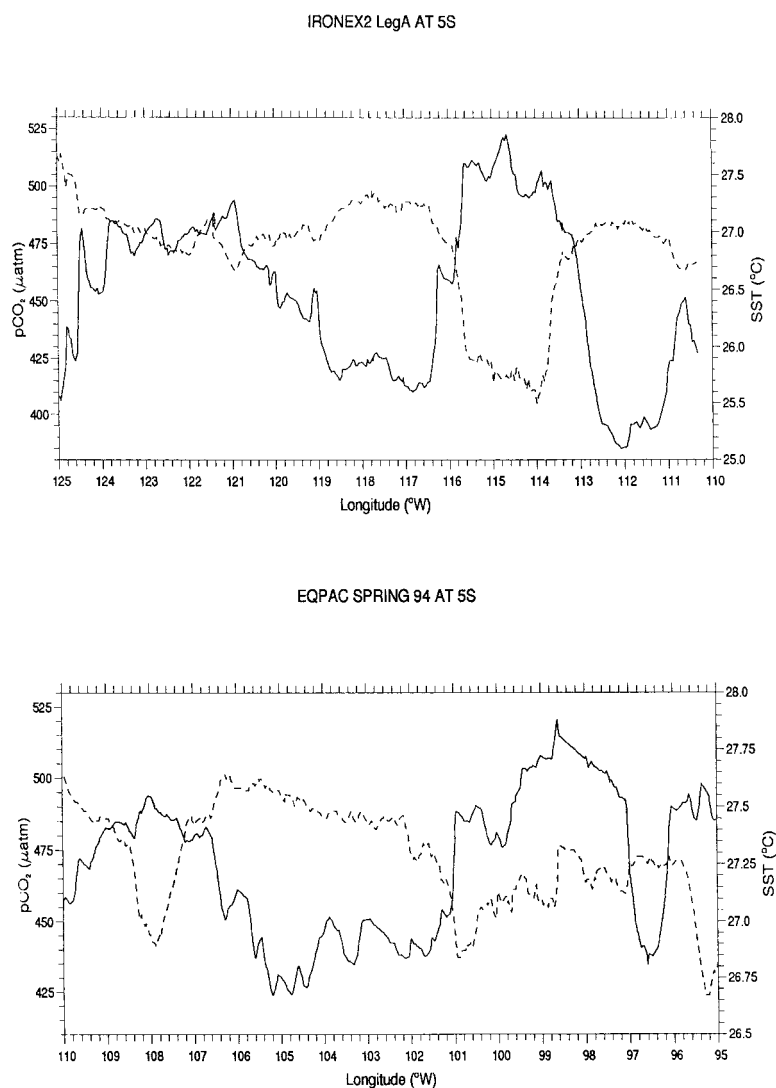


Fig. 3. $p\text{CO}_2$ (solid line) and SST (dashed line) measured along 5°S during IronEx 2 (21–24 May 95) (top) and during EqPac Spring 94 (23–25 April 94) (bottom) cruises.

meanders are also visible at other seasons on the AVHRR SST images but no $p\text{CO}_2$ measurements were available at this time.

4. Diel variability observed during Alize cruise

The Alize II cruise (January to March 1991) crosses the equatorial Pacific from Panama to

Nouméa via Tahiti (Lefèvre and Dandonneau, 1992). Continuous measurements of $p\text{CO}_2$ were performed using an infrared CO_2 analyzer and sea water pumped at 2 m depth, every 20 min. SST was measured only during stations, about every 3 h. Lefèvre et al. (1994) studied the surface distribution of $p\text{CO}_2$ in relation to chemical and physical properties measured during the cruise and they estimated the contribution of physical, chemical and biological processes to the observed $p\text{CO}_2$

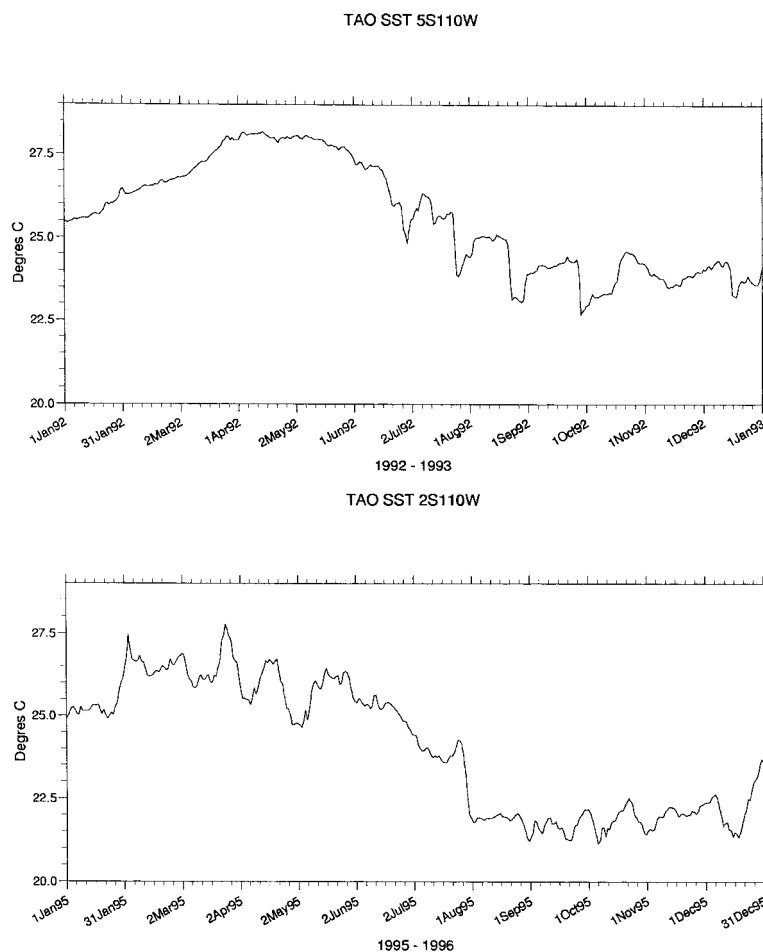


Fig. 4. SST measured on TAO mooring at 5°S, 110°W in 1992 (top) and at 2°S, 110°W in 1995 (bottom).

variability. Here we focus on the high frequency temporal variability in $p\text{CO}_2$ measured in the upwelled waters east of 140°W. Fig. 5 (thin dots) shows the $p\text{CO}_2$ measured during the Alize cruise from 95°W (9 January 1991) to 138°W (27 January 1991) between 1°N and 1°S, as a function of local time. The large decreases of $p\text{CO}_2$ observed on 13, 17 and 24 January were due to the presence of instability waves (Lefèvre and Dandonneau, 1992). The corresponding periods with low $p\text{CO}_2$ values, indicated with vertical dashed lines on Fig. 5, were removed for the following study.

Surface waters are warmed by solar heating during the day, and since CO_2 solubility and dissociation constants in seawater depend on the temperature, $p\text{CO}_2$ at the ocean surface increases

by about $4.23\%^\circ\text{C}^{-1}$ (Copin-Montegut, 1988, 1989; Takahashi et al., 1993). Correlations in the diel variations of $p\text{CO}_2$ and sea surface temperature have been observed in high frequency measurements made at 1.6 m depth by CARIOCA buoys (McNeil and Merlivat, 1996). Such variations have also been observed in the equatorial Pacific at 140°W during EqPac92 by Goyet and Peltzer (1997) and at 150°W during Olipac by Metzl and Bres (1996).

On another hand, biological activity and diurnal convective mixing of the surface water may be responsible for $p\text{CO}_2$ diel variations anticorrelated with SST. Such $p\text{CO}_2$ variations have been observed in the equatorial Atlantic upwelling in continuous CARIOCA measurements (Bakker,

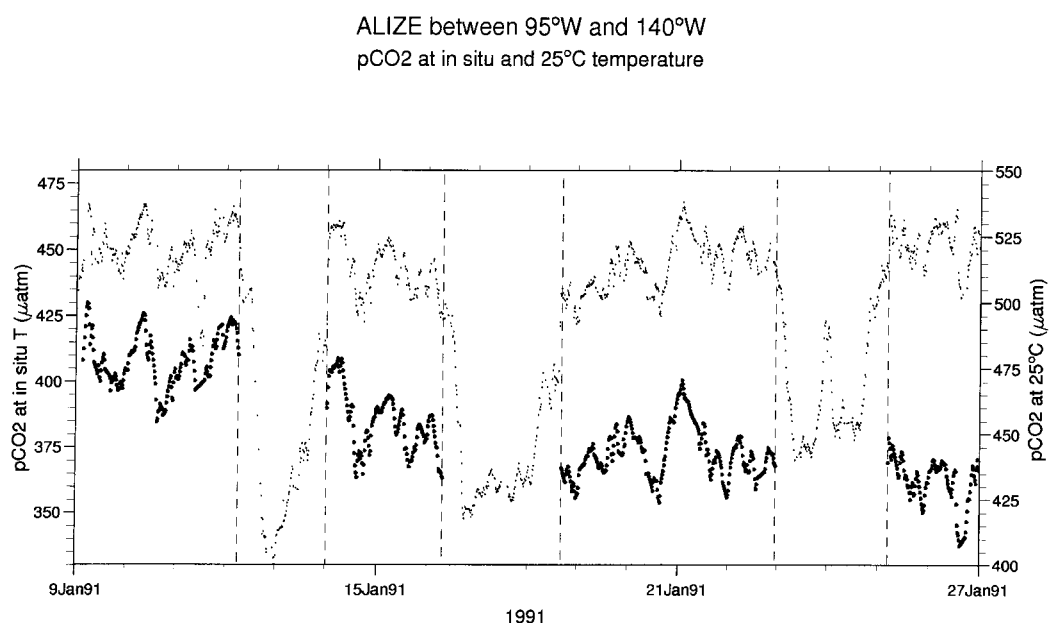


Fig. 5. $p\text{CO}_2$ (light points) and $p\text{CO}_2$ at 25°C (dark points) measured during the Alize II cruise from 95°W to 138°W between 1°N and 1°S as a function of local time. Vertical bars indicate periods with low $p\text{CO}_2$ values due to the presence of instability waves. In the case of $p\text{CO}_2$ at 25°C, these low values have been removed.

personal communication). During the North Atlantic Bloom Experiment (NABE), Robertson et al. (1993) have also observed such diel $p\text{CO}_2$ variability anticorrelated with oxygen variations. Taylor and Stephens (1993) have shown, using a simple biogeochemical model, that the variations observed during NABE, up to 10 μatm peak-to-peak variations, are too large to be due to photosynthetic activity alone but that they could be due to the diurnal convection in the surface mixed layer.

During Alize II, SST is minimum in the early morning and maximum in the late afternoon, but its diel cycle is weak, less than a 0.5°C peak-to-peak variation, which would correspond to a less than 8 μatm in phase variation of $p\text{CO}_2$. We do not observe such an effect in the data. Most of the time, $p\text{CO}_2$ is maximum in the early morning, around 6 h local time, and minimum in the afternoon between 15 h and 18 h local time, i.e., in opposite phase with SST. In order to compute the amplitude spectrum of $p\text{CO}_2$, we evenly resample the time series shown in Fig. 5 with a linear interpolation after having discarded data gaps, due to the presence of instability waves or to

extraequatorial sections, longer than 1 day. We obtain 930 interpolated points sampled at 20 mn intervals. We compute their amplitude spectrum using a Fourier analysis. The $p\text{CO}_2$ diel cycle is clearly seen on the amplitude spectrum of $p\text{CO}_2$ (Fig. 6) which shows a sharp peak at a 24 h period. If the $p\text{CO}_2$ signal were purely sinusoidal, a 2.5 $\mu\text{atm Hz}^{-0.5}$ spectrum amplitude would correspond to a peak-to-peak variation of $p\text{CO}_2$ of 10 μatm . This is the order of magnitude of the diel variations observed in Fig. 5.

In order to remove the temperature effect, we also convert the $p\text{CO}_2$ measured at in situ temperature to $p\text{CO}_2$ at 25°C using the Copin-Montegut (1988) formulae. The resulting $p\text{CO}_2$ at 25°C in the equatorial upwelled waters are shown in Fig. 5 (thick dots). The decrease in $p\text{CO}_2$ at 25°C between 9 and 27 January is due to the large scale zonal increase of SST (not shown) from about 23°C at 95°W to 26.5°C at 138°W. The spectrum of $p\text{CO}_2$ at 25°C is shown in Fig. 6 (dashed line). The 24 h peak is higher than the one for $p\text{CO}_2$ at in situ temperature since the SST diel cycle is anticorrelated with $p\text{CO}_2$ diel variations.

The diel variations discussed here have to be

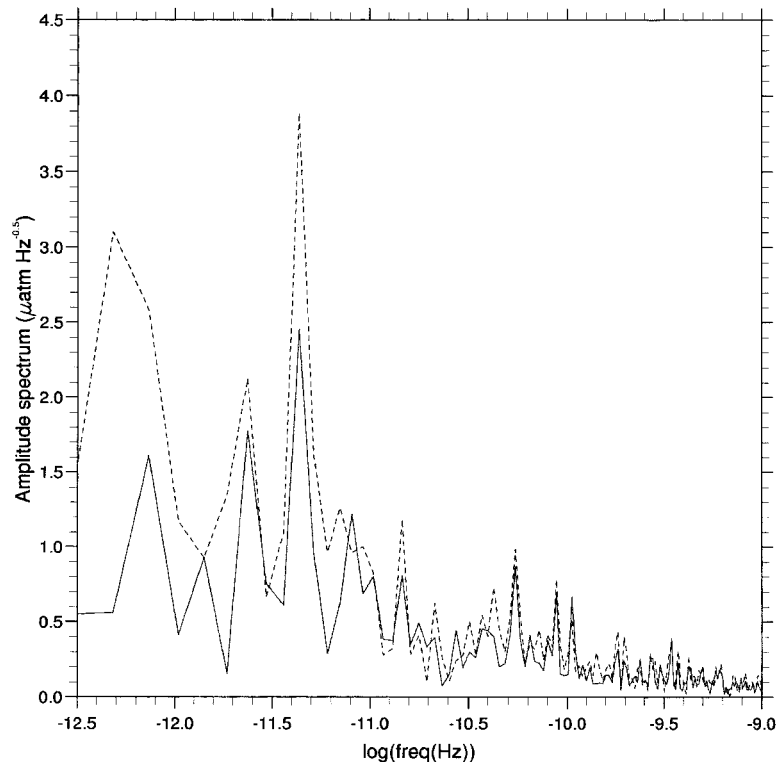
SPECTRA of ALIZE $p\text{CO}_2$ 

Fig. 6. Amplitude spectrum of Alize $p\text{CO}_2$ at in situ temperature (solid line) and at 25°C (dashed line). The natural logarithm of the frequency equal to -11.37 corresponds to a 24 h period.

considered with care since the $p\text{CO}_2$ measurements were made while the ship was underway and include spatial variations. More detailed observations including physical, chemical and biological profiles measured on a fixed point during several days are clearly needed to clarify the amplitude and the origin of such a diel cycle.

In the absence of systematic relationship between $p\text{CO}_2$ and SST diel variations in the equatorial Pacific, when looking for large scale estimate of $p\text{CO}_2$, we will average $p\text{CO}_2$ measurements at fixed location over 24 h.

5. Variability related to zonal displacement of the warm pool

In the equatorial eastern Pacific Ocean, the equatorial upwelling brings cold waters to the

ocean surface. These cold waters are advected westward and poleward; during their drift they are warmed so that the coldest waters are observed in the eastern Pacific Ocean and a minimum of surface temperature occurs at the equator in any meridional section, leading to the so-called "cold tongue" (Bryden and Brady, 1985). The warm pool waters are separated from the recently upwelled waters by a well defined salinity front because the recently upwelled waters are saltier than the warm pool waters. This front is zonally displaced in relation with El Niño and La Niña events, to the east when El Niño conditions reinforce, and to the west when La Niña conditions settle (Picaud et al., 1996). This has been supported by observational studies (Picaud et al., 1996; Ando and McPhaden, 1997) and model studies (Vialard and Delecluse, 1997). This front is expected to have a strong signature in $p\text{CO}_2$ because the

recently upwelled waters are highly oversaturated whereas waters in the warm pool are usually close to the equilibrium with respect to the atmosphere (Feely et al., 1995). In the western and central equatorial Pacific, Inoue et al. (1996) pointed out the existence of an east–west discontinuity of ΔP decreasing from a few tens of μatm on the east side to near equilibrium on the west side correlated with a salinity front. Moreover they showed that the zonal displacement of this gradient is correlated with the Southern Oscillation Index.

Because of this zonal displacement, the surface of the region under the influence of the equatorial upwelling is reduced during an El Niño event and enlarged during a La Niña event. In order to quantify such a surface effect on the CO_2 flux integrated over the equatorial Pacific region, we monitor the location of the front using NMC SST maps. The location of the front is more or less well correlated with the 29°C isotherm (Picaud et al., 1996). However, SST alone is not enough to detect this front. Indeed, westerly wind bursts in the western Pacific can be associated with cooling of SST: Ralph et al. (1997) have observed during the Tropical Oceans Global Atmosphere (TOGA) Coupled Ocean–Atmosphere Response Experiment (COARE) that eastward flow of drifters created by westerly wind burst was associated with cooling of up to 1°C of the water traveling eastward. On the other side, when the equatorial upwelling is weak, upwelled waters can be warmed up to 29°C during their westward drift (Edlin et al., 1999). We therefore elaborate a method to automatically detect the western boundary of the cold tongue, in regions where the SST averaged between 1°N and 1°S is warmer than 28.5°C . We consider that a tongue of upwelled water is present if the SST increases by at least 0.8°C from the equator to either 5°N or 5°S . In order not to consider local slowing down of the equatorial upwelling activity, we identify the front as the westernmost position where the tongue of upwelled water does not exist over several consecutive degrees of longitude: 3° of longitude if the SST at the equator is warmer than 29°C , and 6° of longitude if it is between 28.5°C and 29°C . When this procedure continues to detect upwelling but the SST is warmer than 29.2°C , we consider that the front is reached. This method has the advantage of using SST gradients, satellite SST

being more reliable in relative than in absolute value.

In order to keep ΔP measurements representative of the equatorial upwelling region, we consider ΔP measured between 0°N and 3°S , and we reject (1) low ΔP values caused by the presence of instability wave during Alize and EqPac Spring 92 cruises and (2) ΔP measurements south of 1.5°S during EPOCS 88 (May 88) and south of 2°S during mri88_2 (February 88). In the latter case, strong meridional gradients between the equator and 3°S were observed during these cruises that we attribute to a southern boundary of the upwelled waters at 1.5°S and 2°S , respectively. We obtain 7637 points spread among 59 cruises. Taking data between the equator and 1°S instead of 3°S reduces the number of points by 15% but does not change the results of the following analysis, nor improve the quality of the fits. For the quantitative analysis, west of 110°W , we average ΔP measurements either in a radius of 1° in longitude or during 24 h at fixed stations in order to temper the weight of cruises with dense observations and to smooth diel $p\text{CO}_2$ variations. The number of averaged ΔP is 332. East of 110°W , we use averages made by Etcheto et al. (1999) computed as the mean, for each cruise, of ΔP in the region where ΔP is between its maximum value, P_{max} , and $P_{\text{max}} - 10\%$.

The detected front is superimposed on the non-averaged CO_2 air–sea partial pressure gradient (Fig. 7, top). Our detected boundary generally correlates well with the ΔP zonal gradient. Some discrepancies appear especially when the boundary displacement is rapid (see for instance, Flupac in October 1994). This can be attributed to the low temporal resolution (1 week) of the SST map. However, the difference between our detected boundary and the location of the ΔP zonal gradient is usually less than 5° of longitude as demonstrated in Fig. 8, top. This plot shows the 2° , 24 h averages of ΔP and the associated r.m.s. as a function of the difference between the longitude of the measurement, lon , and the longitude of the detected boundary, lon_{wp} . Large dots are averaged ΔP values east of 110°W . West of the boundary ($\text{lon} - \text{lon}_{\text{wp}} < 0$), in the warm pool region, ΔP are between -25 and $25 \mu\text{atm}$ except for 2 points which reach $50 \mu\text{atm}$ (see Fig. 8, top) and were taken during the El Niño 1986–1987. East of the boundary and west of 110°W , ΔP is larger than

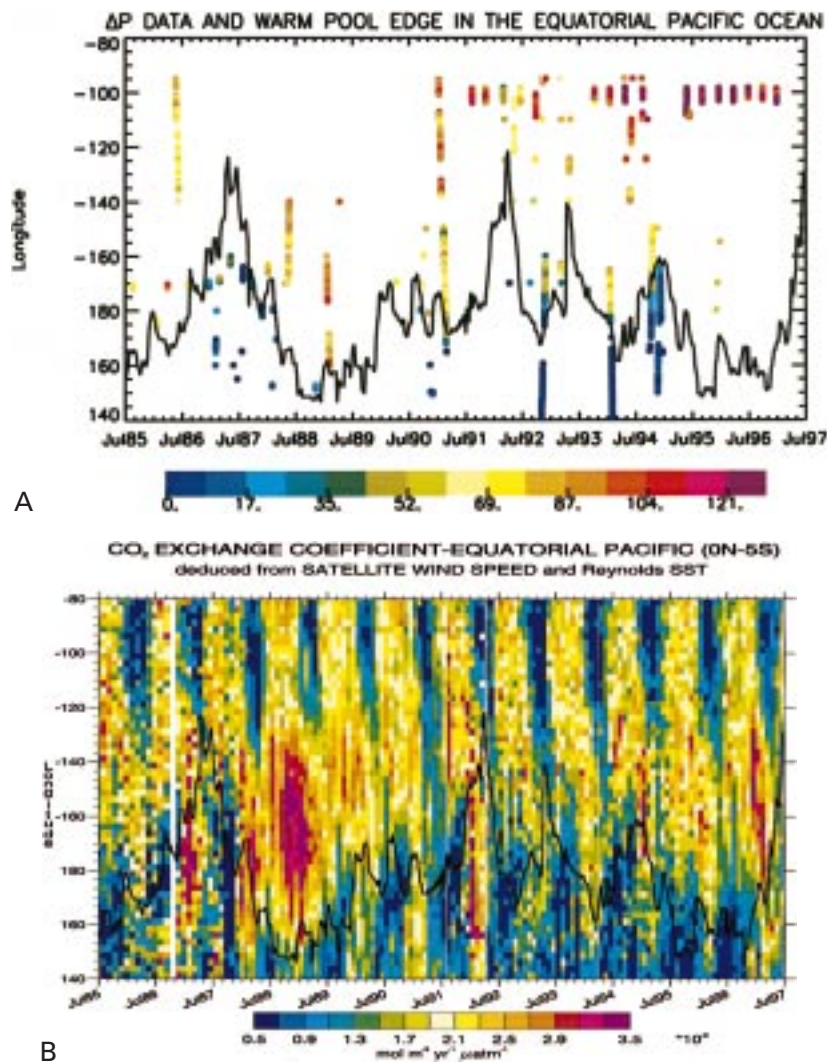


Fig. 7. Zonal distribution in the equatorial Pacific of ΔP usually between 0°N and 3°S (see text) (top) and of K computed with the Liss and Merlivat relationship between 0°N and 5°S (bottom). The dark line indicates the location of the boundary between the equatorial upwelled waters and the warm pool waters, detected using SST maps.

30 μatm and increases towards the east. This behavior is due to the east–west decrease of the quantity of carbon reaching the ocean surface by upwelling and to the decrease of ΔP towards the west, caused by biological activity and by air–sea exchange. It is noteworthy that this west–east increase is approximately linear as a function of (lon–lonwp) whereas it does not correlate with the local longitude (not shown). Actually, at a given longitude in the equatorial Pacific, the ΔP

is even higher when the upwelling activity is stronger and the warm pool boundary further west. This east–west decrease of ΔP is well correlated with the east–west increase of SST due to warming of the upwelled cold waters during their westward advection (Fig. 8, bottom). The large ΔP range for SST between 28.5°C and 29.8°C confirms that a criterion based on the absolute value of SST is not sufficient to identify the warm pool boundary. East of 110°W, no relationship

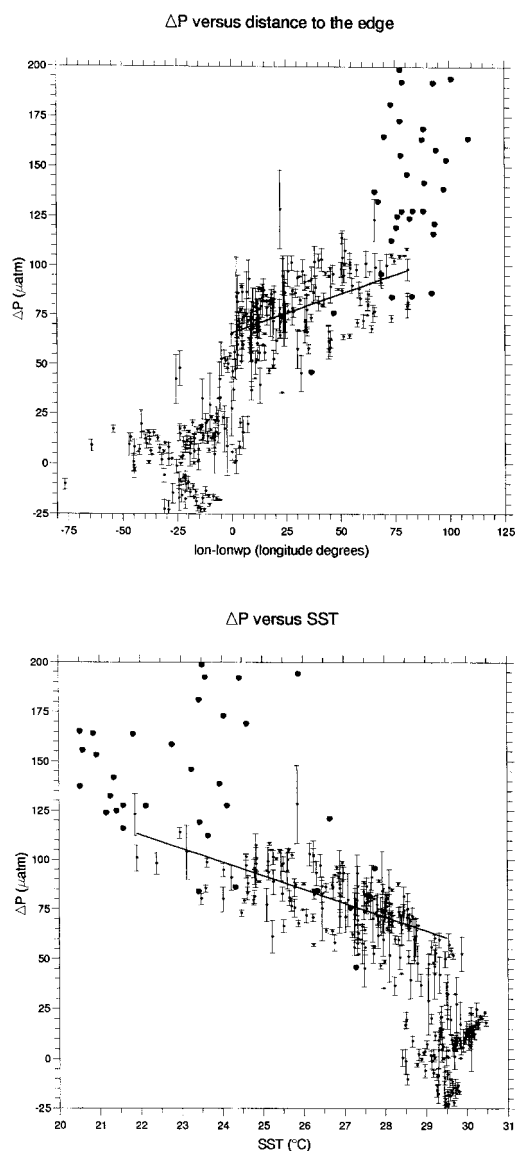


Fig. 8. ΔP as a function of the longitudinal distance of the measurement to the detected boundary (top) and of the measured SST (bottom). Small dots and error bars are ΔP averages over 2° of longitude and 24 h west of 110°W and their associated r.m.s.; large dots are ΔP averages east of 110°W . The superimposed line is the least square fit to the data east of the detected boundary and west of 110°W .

exists with the distance to the warm pool nor with the SST. Two reasons may explain this absence of relationship: (1) the region east of 110°W is under the influence of the Peru upwelling and (2) the amplitude of the SST seasonal cycle in the eastern Pacific (up to 7°C) is larger than the interannual SST variations whereas the ΔP is dominated by the interannual variability. Cosca and Feely (in preparation) found a small seasonal variation for data collected along 95°W and 110°W .

We will now estimate the respective influence of K , ΔP , and the surface of the equatorial upwelling region on the CO_2 flux integrated over the equatorial Pacific.

West of the boundary ($\text{r lon} - \text{r lonwp} < 0$), in the warm pool region, ΔP values are small (Fig. 7, top; Fig. 8, top). The warm pool region is a location with very low wind speeds so that K is usually weak west of the boundary as can be seen in Fig. 7 bottom, with the exception of the onset of El Niño events due to westerly wind bursts. In the following discussion, we will neglect the flux west of the detected boundary.

East of the boundary and west of 110°W , we will approximate ΔP according to 2 different methods based on least square fits between ΔP and $\text{lon} - \text{lonwp}$ and between ΔP and SST. Here below, we call these fits the “lon-lonwp fit” and the “SST fit”, respectively. Both fits are calculated using averaged 24 h, 2° ΔP in the equatorial upwelling region that we define as being greater than $25 \mu\text{atm}$ and as being east of lonwp ($\text{lon} - \text{lonwp} > 0$). The former test discards low measurements 5° from the boundary corresponding to an inaccurate detection of the warm pool boundary. The fits, computed with 197 points, are shown in Fig. 8, and regression equations are:

$$\Delta P = 0.395 \times (\text{lon} - \text{lonwp}) + 65.9, \quad (1)$$

$$\Delta P = -6.96 \times \text{SST} + 266, \quad (2)$$

where ΔP is in μatm , $(\text{lon} - \text{lonwp})$ is in degrees of longitude east, and SST is in $^\circ\text{C}$. The correlation coefficients are 0.53 and 0.65, and the standard deviation of the points with respect to the fits are $14 \mu\text{atm}$ and $12.6 \mu\text{atm}$ for (1) and (2), respectively. The quality of (2) appears to be slightly better than (1). However, the SST exhibits a seasonal variation in the eastern and central Pacific. On a 10-year SST climatology, the SST between 140°W and 110°W is maximum in April and minimum in

September; its seasonal variation is about 3°C at 140°W and 7°C at 110°W . Therefore (2) induces a seasonal variation of ΔP of $20\text{ }\mu\text{atm}$ at 140°W which increases towards the east to $50\text{ }\mu\text{atm}$ at 110°W . Such seasonal variation appears to be large compared with in situ measurements. We will therefore consider both fits. The ΔP value of $125\text{ }\mu\text{atm}$ seen at $\text{lon}-\text{lonwp}=25^{\circ}$ of longitude (which is far from the fit in Fig. 8, top) comes from a local increase of ΔP measured during a La Niña event at 180°W correlated with a local decrease of SST. To check that no systematic bias related to El Niño and La Niña events is included in the fits, we compare the residuals between the fits and the measurements to the SST anomaly computed with respect to a 15-year climatology of NMC SST in the Niño 3 box (150°W – 90°W ; 5°N – 5°S). The corresponding plot for the “SST fit” is drawn in Fig. 9. No systematic bias appears, the dispersion of the points around the fits are mostly due to spatial variability.

Between 110°W and 90°W , we use the method developed by Etcheto et al. (1999), which relates ΔP to the SST anomaly in the Peru upwelling with a 5-week delay:

$$p\text{CO}_2 = 463 - 22.1 \times \text{east SST anomaly},$$

where $p\text{CO}_2$ is in μatm , and east SST anomaly

(in $^{\circ}\text{C}$) is the SST anomaly measured in the box (83°W – 85°W , 2°S – 4°S) 5 weeks before $p\text{CO}_2$ measurement.

We limit our study west of 90°W because lack of data east of 90°W in a very variable area makes impossible to develop a reliable diagnostic method for this region.

The estimated ΔP averaged between the warm pool boundary and 90°W are shown in Fig. 10, top, from January 1985 to July 1997. The 12-year average obtained with (1) and (2) is equal to $90\text{ }\mu\text{atm}$ and $88\text{ }\mu\text{atm}$, respectively. Extrapolated ΔP obtained with (1) and (2) are very similar; the main differences occur at the beginning of the year because (2) introduces a seasonal variation, ΔP being weaker at the beginning of the year. On average, this variation is small because the largest ΔP come from the eastern Pacific where the extrapolation does not include seasonal variation.

When examining the meridional profiles of ΔP as was done by Boutin and Etcheto (1997), we notice that the high equatorial ΔP values usually extend over 5° of latitude. The northern and southern border of this high ΔP region is variable, depending on the presence of instability waves in the eastern Pacific. Even in the west, the northern border can be shifted by several degrees of latitude from one cruise to another. However, the width of the equatorial upwelled water as well as the location of its boundary with subtropical water is not easily detectable on low resolution SST maps because (1) TIW moves of several degrees of longitude in 1 week and (2) the meridional SST gradients are frequently weak, especially south of the equator (see Section 3) and are smoothed by the SST interpolation. On high resolution SST maps, clouds are a major obstacle to systematic detection. Therefore, in the following analysis, we will consider that our set of ΔP , usually between 0°N and 3°S , is representative of a band of 5° of latitude. To compute the flux, we combine the extrapolated ΔP with K averaged between 0°N and 5°S , although the 5° latitudinal band can be shifted north of 0°N , because the K averaged between 2.5°N and 2.5°S is not significantly different from the 0°N – 5°S average.

The resulting fluxes are shown in Fig. 10. They are almost the same whatever extrapolated ΔP is used. Over 12 years, the fluxes computed using the Liss and Merlivat (1986) relationship (Fig. 10, top and middle), ϕ , and the Wanninkhof (1992)

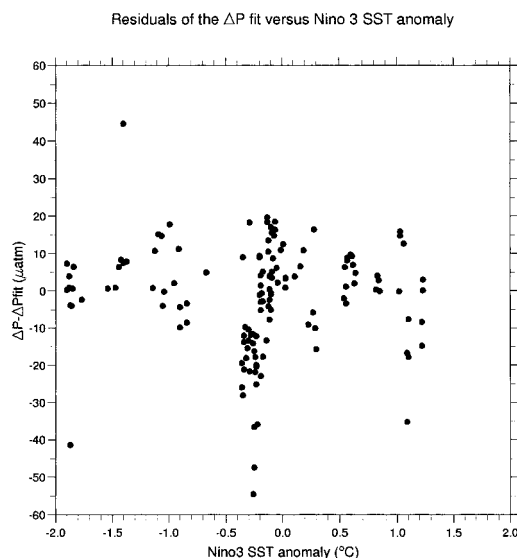
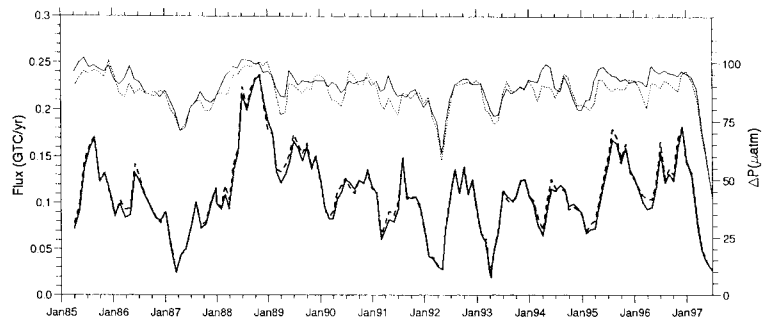
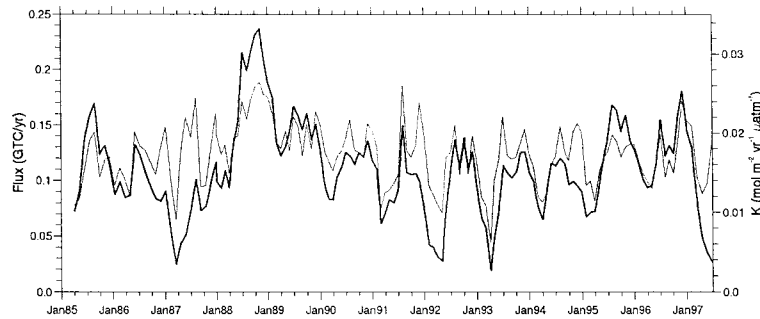


Fig. 9. Residuals of ΔP with respect to the “SST fit” versus the SST anomaly in the Niño 3 region.

AIR-SEA CO₂ FLUX over the EQUATORIAL PACIFIC OCEAN
Extrapolated ΔP average; Liss and Merlivat relationship



K average; Liss and Merlivat relationship



K average; Wanninkhof relationship

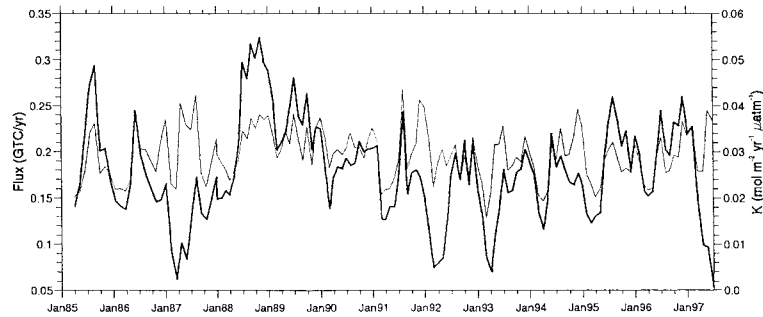


Fig. 10. Top: time series of the mean ΔP averaged between the boundary and 90°W and extrapolated using a function of SST (thin line) and a function of $\text{rlon}-\text{rlonwp}$ (thin dashed line) and of the CO₂ flux integrated over 5° of latitude and calculated using the Liss and Merlivat parametrization for K and a function of SST (thick line) or a function of $\text{lon}-\text{lonwp}$ (thick dashed line) to estimate ΔP . Middle: time series of the mean K computed using the Liss and Merlivat parametrization and averaged between the boundary and 90°W (thin dashed line) and of the CO₂ flux integrated over 5° of latitude and calculated using a function of SST (thick line). Bottom: same as middle but using the Wanninkhof parametrization to compute K instead of the Liss and Merlivat one. Note that middle and bottom figures have different scales.

relationship (Fig. 10, bottom), ϕ_w , average to 0.11 GtC yr^{-1} and to 0.18 GtC yr^{-1} , respectively. Because of the strong non-linearity of the Liss and Merlivat relationship at 3.6 m s^{-1} , ϕ is slightly more variable in relative units than ϕ_w : their coefficients of variation, defined as the ratio of their r.m.s. over their average, are 38% and 30%, respectively. Whatever K – U relationship is used, the flux variations are large compared to the variations of ΔP . For instance between the 1987 El Niño and the 1988 La Niña event, the outgassing flux varies by a factor 5 whereas ΔP varies by only 25%. Moreover, the flux exhibits a large seasonal variation contrary to ΔP which, according to our extrapolation, has small or no seasonal variation. The seasonal flux variation is due to the seasonal variation in K as can be seen in Fig. 10, middle and bottom. Finally, the large variations of the flux between (1) normal year, (2) the 1987, 1992, 1997 El Niño years and (3) the 1988 La Niña year are primarily driven by the variation in the surface area over which the flux is integrated, as can be seen by comparing Fig. 7 with Fig. 10.

If the differences in the integration area are taken into account, our integrated fluxes can be compared with previous estimates. Feely et al. (1995) found a flux per unit area of $0.77 \text{ mol m}^{-2} \text{ yr}^{-1}$ in March–May 1992 in the 110°W – 170°W , 0°N – 10°S region and of $2 \text{ mol m}^{-2} \text{ yr}^{-1}$ in September–December 1992 in the 95°W – 170°W , 0°N – 10°S region. These fluxes were computed using the Wanninkhof (1992) parametrization and ECMWF wind speeds. Boutin et al. (1996) showed that in the 10°N – 10°S equatorial band in 1992 the ECMWF wind speeds were in good agreement with the ERS1 wind speeds so that the fluxes we compute using the Wanninkhof parametrization should be close to the ones computed by Feely et al. (1995). In March–May 1992, we found that the warm pool boundary is located at 146°W and that east of it, the outgassing flux computed using the Wanninkhof relationship amounts to $0.054 \text{ GtC yr}^{-1}$. We normalize this flux to the same area as Feely et al. (1995), considering that the flux is zero south of 5°S , since Fig. 9 of (Feely et al., 1995) shows that the region of large outgassing flux extends to about 5°S . The flux per unit area equal to $0.61 \text{ mol m}^{-2} \text{ yr}^{-1}$, which is 21% lower than Feely et al. (1995) estimate.

In September–December 1992, we found that the warm pool boundary is located at 173°E and that the flux integrated between 0° – 5°S and 95°W – 170°W is $0.146 \text{ GtC yr}^{-1}$. According to Fig. 9 of Feely et al. (1995), the outgassing flux extends to 10°S , east of 120°W . We therefore assume that the flux between 5°S and 10°S is equal to the 0 – 5°S value east of 120°W and is zero west of 120°W . The flux per unit area amounts to $1.76 \text{ mol m}^{-2} \text{ yr}^{-1}$ which is 12% lower than the Feely et al. estimate.

We will now examine quantitatively the contribution of the parameters, K , ΔP and the area of integration, S , to the variability in the flux, ϕ computed with the Liss and Merlivat (1986) relationship.

To simplify, we neglect the spatial correlation of K , ΔP and S . We justify it a posteriori by computing the flux as the product of the spatially averaged K , ΔP and S . The resulting flux is only 3% lower than the flux computed by integrating the $K\Delta P$ product over S , the r.m.s. of the difference being 6% of the flux. Therefore, in the following we only study the temporal variation of the spatially integrated quantities. Taking the derivative of the ϕ logarithm, the relative variation of ϕ , $d\phi$, can be written as:

$$|d\phi/\phi| = |dK/K + d\Delta P/\Delta P + dS/S|, \quad (3)$$

where $| \cdot |$ denotes the absolute value and dK/K , $d\Delta P/\Delta P$ and dS/S are the relative variations of K , ΔP and S , respectively. Eq. (3) can be bounded by:

$$|d\phi/\phi| \leq |dK/K| + |d\Delta P/\Delta P| + |dS/S|. \quad (4)$$

We estimate the magnitude of each of these terms using the coefficients of variation in K , ΔP and S defined as the ratio of their r.m.s. over their average. The coefficient of variation in K is 21.6%, that for ΔP is 9.5% and that for S is 20.6%, over the 12 years study. The contributions of K and S to the flux variability are of the same order of magnitude whereas the contribution of ΔP is smaller by a factor 2.3. At first order, K governs the seasonal variation of the flux while S drives its interannual variations.

6. Discussion and conclusion

Satellite SST imagery enables us to show that the large $p\text{CO}_2$ variability in the eastern equatorial

Pacific is due to the presence of meanders in the front between recently upwelled water reaching the ocean surface and tropical surface water. Although this phenomenon was known at the northern boundary of the upwelling region and has been related to the presence of TIW, it has never been shown to affect $p\text{CO}_2$ south of the equatorial upwelling zone. According to the TOGA/TAO SST time series and to the SST imagery, southern meanders appear to be periodic with a period and a wavelength close to the northern TIW.

Once low $p\text{CO}_2$ measurements corresponding to intrusion of tropical water at the equator are removed from the Alize II data set, a strong $p\text{CO}_2$ diel cycle (10 μatm peak-to-peak variation) is found to be anticorrelated with the SST variations. Such a diel cycle has been observed in the equatorial Atlantic upwelling by CARIOCA autonomous drifter (Bakker, personal communication) and during the NABE experiment (Robertson et al., 1993). Since $p\text{CO}_2$ diel variations are also frequently observed to be correlated with SST, for large scale study, it is preferable to use 24 h averages of fixed $p\text{CO}_2$ measurements. This would not bias CO_2 flux determination since the diel variability of the wind speed in the equatorial Pacific is very weak.

In the western part of the equatorial Pacific, the variability in $p\text{CO}_2$ is primarily governed by the displacement of the warm pool. We estimate that this phenomenon has a very large influence on the CO_2 flux integrated over the outgassing region of the equatorial Pacific because it makes the surface of integration variable by up to a factor of two between El Niño and La Niña years. Over 12 years, the flux integrated between 0°N and 5°S averages to 0.11 GtC yr^{-1} and the mean surface of integration is $5.7 \times 10^6 \text{ km}^2$.

In this study, the integration area is limited to a 5° latitudinal band. In the eastern Pacific, the high $p\text{CO}_2$ region extends over a broader band than 5° of latitude (Dandonneau, 1995; Feely et al., 1995). During the ECOA cruises which cross 10°S at about 115°W , a large decrease of $p\text{CO}_2$ corresponding to the transition between the Peruvian and equatorial upwelling waters, and subtropical waters is frequently observed near 10°S but it varies by several degrees of latitude from one cruise to another. High variability in the location of this transition has also been observed on chloro-

phyll measurements by Dandonneau and Edlin (1987) who show that the transition is associated with the convergence of the westward South Equatorial Current and the subtropical eastward flow. A thorough study of this transition as well as of the meridional variation of $p\text{CO}_2$ between 5°S and the transition is clearly needed to reach a reliable estimate of the integrated flux over the entire region under the influence of the upwelling.

In 1992, our flux estimates agree within 20% with previous estimates determined using the same data but employing a simpler extrapolation scheme and a different wind data set. The error associated with the flux can be more generally estimated as:

$$d\phi/\phi = d\phi_W/\phi + d\phi_E/\phi,$$

where $d\phi_W$ and $d\phi_E$ are the errors on the integrated fluxes west and east of 110°W , respectively, computed using eq. (4) and replacing dK , $d\Delta P$ and dS by the error on each parameter. According to the residuals to the fits, we estimate the error on ΔP to be $12.6 \mu\text{atm}$ that is a 16% relative error west of 110°W , and Etcheto et al. (1999) estimate the relative error on ΔP to be 24.5% east of 110°W . We also estimate that the boundary is detected within 5° of longitude that is a 7% relative error on the flux west of 110°W . Boutin and Etcheto (1997) have shown that in this region the accuracy of Geosat and SSM/I K is $0.14 \times 10^{-2} \text{ mol m}^{-2} \text{ yr}^{-1} \mu\text{atm}^{-1}$, that is a 8% relative error; this error must be considered as an upper bound since the scatterometer measurements quality is better. Given that the fluxes east and west of 110°W are 0.03 GtC yr^{-1} and 0.08 GtC yr^{-1} , respectively, we estimate the error on the integrated flux between the warm pool and 90°W to be 31%. Of course this does not include the discrepancy by a factor 2 between the Liss and Merlivat and the Wanninkhof relationships.

Although limited to a restricted area, our study indicates that it is the location of the boundary of the warm pool and the K variability which primarily govern the variability of the integrated flux.

New CARIOCA drifters continuously measuring $p\text{CO}_2$ in near surface will soon complement ship measurements. This should help to answer questions raised in this paper about the meridional extension of the signature of the upwelling on $p\text{CO}_2$ and the associated meridional variation of $p\text{CO}_2$, about diel $p\text{CO}_2$ variability and about a

possible seasonal cycle of $p\text{CO}_2$. They will also allow a check on our interpolation method by providing independent measurements. This was not possible in this study given the limited number of cruises.

Improvements of the $p\text{CO}_2$ approximation method are also expected from the recently launched ocean color sensors (Seawifs, OCTS/POLDER) which will complement the SST information.

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