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# Interannual Variability of Equatorial Pacific CO<sub>2</sub> Fluxes Estimated from Temperature and Salinity Data

H. Loukos<sup>1,2</sup>, F. Vivier<sup>3</sup>, P. P. Murphy<sup>4</sup>, D. E. Harrison<sup>5</sup> and C. Le Quéré<sup>6</sup>

**Abstract.** Based on atmospheric data and models, the tropical CO<sub>2</sub> source anomaly reaches up to 2 GtC yr<sup>-1</sup>, but the respective contributions of the terrestrial biosphere and the oceans to this flux are difficult to quantify. Here we present a new method for estimating CO<sub>2</sub> fluxes from oceanic observations based on the surprisingly good predictive skill of temperature and salinity for surface dissolved inorganic carbon. Using historical temperature and salinity data, we reconstruct the basin scale CO<sub>2</sub> flux to the atmosphere in the equatorial Pacific from 1982 to 1993. We find that interannual anomalies do not exceed 0.4 ± 0.2 GtC yr<sup>-1</sup> which suggests that up to 80% of the tropical CO<sub>2</sub> source anomaly is due to the terrestrial biosphere.

## Introduction

The largest natural source of CO<sub>2</sub> to the atmosphere is the tropical ocean [Takahashi *et al.*, 1997]. Upwelling of subsurface waters rich in dissolved inorganic carbon (DIC) maintains a higher partial pressure of CO<sub>2</sub> (pCO<sub>2</sub>) in the surface ocean than in the overlying atmosphere, which results in a CO<sub>2</sub> outgassing from the ocean. As the largest tropical ocean and the main scene of upwelling variability forced by El Niño–Southern Oscillation events (ENSO), the equatorial Pacific is a major contributor to the global interannual fluctuations of oceanic CO<sub>2</sub> evasion rates [Lee *et al.*, 1998; Le Quéré *et al.*, 1999].

The most direct approach to determine the mean regional value of pCO<sub>2</sub> and its interannual variability is from *in situ* measurements. In the equatorial Pacific, pCO<sub>2</sub> has been measured regularly on five longitudinal sections since 1992 [Feely *et al.*, 1999]. Despite increasingly better data coverage, basin scale estimates are still affected by the need of significant interpolation. An alternative approach is to infer pCO<sub>2</sub> from other water properties (mainly temperature) for which more data is available. Lee *et al.* [1998] use a SST-pCO<sub>2</sub> relationship for seasonal to interannual estimates in the equatorial Pacific. However, because of the coarse resolution of their method (4° × 5°), the

spatial structure of pCO<sub>2</sub> within its high-gradient region is weakly captured, particularly at the edge of the warm waters of the western Pacific (the Warm Pool). Boutin *et al.* [1999] propose to modulate the CO<sub>2</sub> flux by the zonal displacement of the Warm Pool, but this method is only valid between the equator and 5° South.

Here we propose to estimate pCO<sub>2</sub> from both SST and sea surface salinity (SSS) observations. In the equatorial Pacific, surface variability of nutrients and DIC is primarily controlled by the physical structure of the upper water column because biological activity closely follows the variability of dynamical processes that supply surface waters with the essential nutrients [Archer *et al.*, 1996; Garside and Garside, 1995]. Because SST and SSS variability reflect those physical changes, they can be used to estimate the surface concentrations of DIC, from which pCO<sub>2</sub> can be calculated. We show that a single SST-SSS-DIC relationship can be used in the equatorial Pacific for both ENSO and non-ENSO conditions. This allows us to estimate CO<sub>2</sub> evasion rates at the spatial and temporal coverage of the SST and SSS observations.

## Method

We calculate a polynomial approximation of DIC by linear least square fitting to surface data (< 20m depth) from six surveys between 10°N–10°S and 170°W–96°W covering both ENSO and non-ENSO conditions of the early 1990's (WOCE sections P17C, P17S/P16S, P16C of 1991, EQPAC cruises of 1992, and WOCE section P18 of 1994). We limit our study to DIC measurements made using similar methods and calibrated against standard reference materials<sup>1</sup>. A fourth order polynomial is selected, since increasing order does not significantly decrease quadratic error.

$$\begin{aligned}
 dic = & 0.4665 - 0.7004t + 0.6352s - 0.1841t^2 \\
 & + 0.1021ts - 0.4304s^2 - 0.0349t^3 + 0.0757t^2s \\
 & + 0.0899ts^2 - 0.1831s^3 + 0.0181t^4 - 0.0341t^3s \\
 & + 0.0884t^2s^2 + 0.0211ts^3 - 0.0220s^4
 \end{aligned} \quad (1)$$

where  $dic = (\text{DIC}-2000)/50$ ,  $t = (\text{SST}-25)/3$ , and  $s = (\text{SSS}-35)/0.5$ .  $\sigma = \pm 9 \mu\text{mol kg}^{-1}$ , with 435 samples. This fit is remarkable given the range of SST and SSS and the seasonal-to-interannual variability encountered during those cruises (17–30°C and 33.7–35.9). The SSS dependency accounts mainly for the south-north gradient of DIC, while upwelled waters with high DIC content have SST-SSS characteristics much more scattered with increasing temperature. Data

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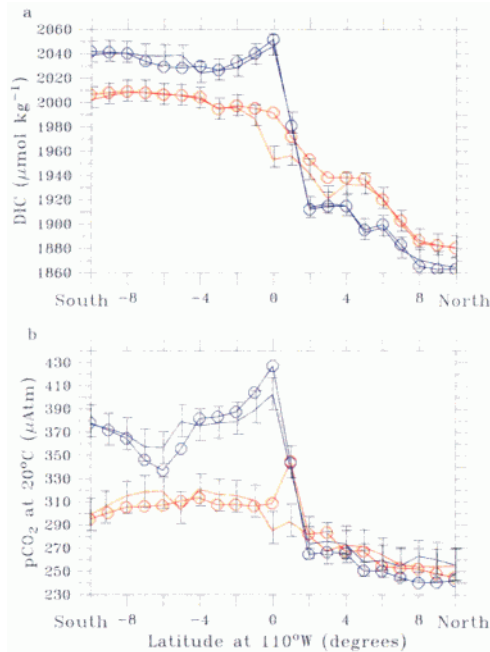
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<sup>1</sup>Reported standard deviations for DIC measurements from each cruise are ± 2 μmol kg<sup>-1</sup>. Certified reference material for DIC developed by A. Dickson (UNESCO Report 60, 1990) were used to control the quality of the measurements. In consequence, the between-cruise bias was estimated to be < 4 μmol kg<sup>-1</sup>



**Figure 1.** DIC and pCO<sub>2</sub> at 20° C at 110° W. Data (circles) and estimates (line) for ENSO (Spring of 1992) in red and non-ENSO conditions (Fall of 1992) in blue.

from the same cruises show that pCO<sub>2</sub> and (SST, SSS) couples have no one-to-one relationship. Thus, pCO<sub>2</sub> is calculated from the estimated DIC and alkalinity (TA) using a thermodynamic model for the carbonate system in sea-water [Murphy, 1996] where TA is also expressed as a function of SST and SSS according to Millero *et al.*, [1998]. Figure 1 illustrates the satisfactory agreement between measured and estimated surface DIC and pCO<sub>2</sub>.

To evaluate the potential of this method, we estimate the bulk pCO<sub>2</sub> difference between the ocean and the atmosphere ( $\Delta p\text{CO}_2$ ) on a  $1.5^\circ \times 1.5^\circ$  grid ( $140^\circ\text{E} - 95^\circ\text{W}$  and  $5^\circ\text{N} - 10^\circ\text{S}$ , Figure 2a) from surface DIC and TA inferred from basin scale monthly SST [Reynolds and Smith, 1994] and SSS [Delcroix, 1998]. However, like DIC, the estimated pCO<sub>2</sub> does not include changes induced by anthropogenic CO<sub>2</sub> emissions ( $\approx 5 \mu\text{mol kg}^{-1}\text{yr}^{-1}$  or  $1.25 \mu\text{atm yr}^{-1}$ ). Therefore, to estimate  $\Delta p\text{CO}_2$  we assume a similar anthropogenic CO<sub>2</sub> increase in the ocean and in the atmosphere, which agrees with the observed trend [Feely *et al.*, 1999].

The CO<sub>2</sub> flux to the atmosphere ( $F_C$ ) is estimated from  $\Delta p\text{CO}_2$  as  $F_C = GEC \times \Delta p\text{CO}_2$  where  $GEC$  is the gas exchange coefficient which depends primarily on the wind speed [Wanninkhof, 1992] (Figure 2b). We use monthly values from the wind speed reanalysis (1979–1993) by the European Centre for Medium Range Weather Forecasts, a consistent product that assimilates wind speed data measured by the TAO array. The beginning of our study period is set by the SST field and the end by the wind product.

The flux estimates (Figure 2c) are subject to errors from (1) DIC ( $\sigma = \pm 9 \mu\text{mol kg}^{-1}$ ) and TA ( $\sigma = \pm 5 \mu\text{mol kg}^{-1}$ ); (2) uncertainties in the SST ( $\pm 0.5^\circ\text{C}$ ) and SSS ( $\pm 0.1$ ) fields; (3) accuracy of the pCO<sub>2</sub> calculation ( $\pm 5 \mu\text{atm}$ ); (4) uncertainties in the wind dependency of  $GEC$  and wind forcing (25%); and (5) the ocean skin temperature effect ( $\pm 0.3^\circ\text{C}$ ). Only (1) and (3) are specific to this study. The DIC fit will be

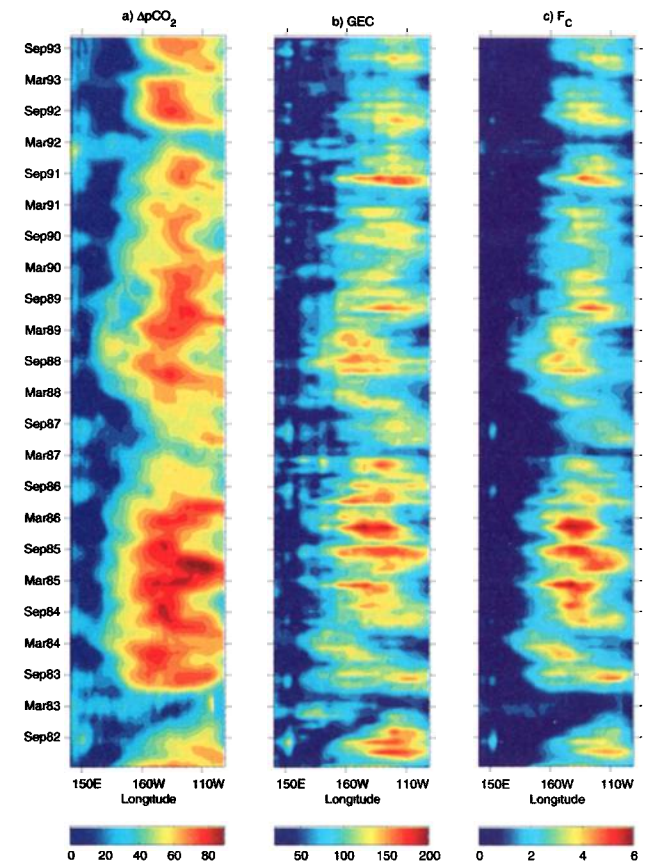
improved as more data become available, especially for La Niña events, in order to include large SST-DIC anomalies. Considering all the above errors and after comparison with field data, we estimate a 50% uncertainty on the CO<sub>2</sub> fluxes.

## Results and discussion

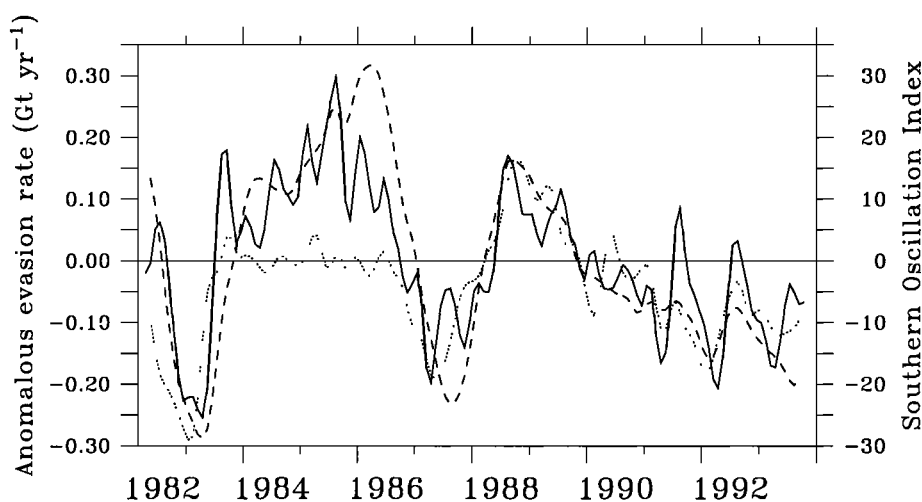
High  $\Delta p\text{CO}_2$  (high DIC) values occur in the cool-salty central and eastern equatorial Pacific, and low  $\Delta p\text{CO}_2$  (low DIC) cover the warm-fresh western Pacific (Figure 2a). During ENSO events, the eastward advection of western waters is reflected in the appearance of low  $\Delta p\text{CO}_2$  along the equator. The most dramatic change for the 1982–93 period occurs during the 1982–1983 El Niño. Moreover, periods of low  $\Delta p\text{CO}_2$  coincide with low  $GEC$  since reduced upwelling is associated with a relaxation of the Trade Winds. Those two parameters act together to modulate  $F_C$ .

The evasion rate ( $R_C$ , the basin integral of  $F_C$ , Figure 3) is in agreement with field estimates (Table 1). Indeed, our estimate of  $F_C$  falls within the uncertainties of the observations. The only substantial discrepancy is for the 1982–83 El Niño: we estimate a minimum evasion rate of  $0.3 \text{ Gt yr}^{-1}$  while *in situ* data suggest no-flux conditions. Because the field estimate did not include observations south of the equator, where recent data indicate relatively higher pCO<sub>2</sub>, we find no cause for concern over the difference in flux estimate.

Negative anomalies of  $R_C$  follow the Southern Oscillation Index (SOI) with several months lag (3–4 months like SST). They occur during 1982–83, 1986–87 and almost continu-



**Figure 2.** (a)  $\Delta p\text{CO}_2$ : air-sea CO<sub>2</sub> gradient ( $\mu\text{atm}$ ), (b)  $GEC$ : gas exchange coefficient (moles  $\text{m}^{-2} \text{day}^{-1} \text{atm}^{-1}$ ), and (c)  $F_C$ : CO<sub>2</sub> evasion rate (moles  $\text{m}^{-2} \text{day}^{-1}$ ) averaged between  $5^\circ\text{N} - 10^\circ\text{S}$ .



**Figure 3.** Anomaly of the CO<sub>2</sub> evasion rate between 140°E–90°W and 5°N–10°S (thick line). The estimate from *Le Quéré et al.* (dashed line) and the Southern Oscillation Index (dotted line) are plotted for comparison. Errors are  $\pm 50\%$  (see text). All data is smoothed with 3-month Hanning filter.

ously after 1989. Positive anomalies do not always correlate with the SOI. Even though  $R_C$  was high during the 1988–89 La Niña (highest SOI of the period), its highest values occurred between the 1982–83 and 1986–87 ENSO events, when the SOI did not record anomalous conditions. During the cold event of 1988–89, estimated DIC reached a maximum, but the potential increase in  $\Delta p\text{CO}_2$  was reduced by a SSS-driven TA maximum centered in 1988–89 (not shown).

The variability of  $R_C$  inferred from this study and from the model simulation of *Le Quéré et al.* are in excellent agreement (Figure 3). These authors use the same wind product to evaluate the *GEC* for the model predicted  $p\text{CO}_2$ . Therefore differences are mostly related to  $p\text{CO}_2$  estimates. Our results also support those of *Lee et al.*, 1998 and *Boutin et al.*, 1999 that suggest a relatively small interannual variability of  $R_C$  in the equatorial Pacific. Such a satisfactory agreement between four inherently different approaches

(*in situ* data, SST–SSS–DIC relationship, SST– $p\text{CO}_2$  relationship, model study) strongly suggests that ENSO-related anomalies of CO<sub>2</sub> evasion rates do not exceed  $\pm 0.4 \text{ Gt yr}^{-1}$  in the equatorial Pacific.

Another way to evaluate the oceanic contribution to interannual CO<sub>2</sub> variations is from inverse methods using the <sup>13</sup>C ratio of atmospheric CO<sub>2</sub>. These methods allow to distinguish between oceanic and terrestrial contributions to atmospheric CO<sub>2</sub> variability (e.g. [*Francey et al.*, 1995; *Keeling et al.*, 1995]). On a global scale, <sup>13</sup>C inversions attribute a much greater role to the ocean than is inferred from oceanic studies [*Lee et al.*, 1998; *Le Quéré et al.*, 1999]. This remains true for the tropics. A recent study by [*Rayner et al.*, 1999] estimates the ENSO-related tropical CO<sub>2</sub> source anomaly as 1–2 Gt yr<sup>-1</sup> from which 1 Gt yr<sup>-1</sup> is attributed to the ocean. Our flux estimate is  $0.5 \pm 0.3 \text{ GtC yr}^{-1}$  with anomalies of  $0.4 \pm 0.2 \text{ GtC yr}^{-1}$  (we increase  $R_C$  by 20% to roughly estimate the contribution of the tropical Atlantic and Indian oceans [*Takahashi et al.*, 1997]).

Do oceanic/atmospheric based estimates under/over-estimate oceanic uptake variability? The answer is certainly not straightforward, but we would argue in favour of a greater role played by the terrestrial biosphere, compared to that of the ocean as to ENSO-related changes of tropical sources of CO<sub>2</sub>. Although we recognize that ocean-based methods may underestimate the variability on ENSO timescales due for instance to insufficient spatio-temporal coverage in the data, or underestimation of interannual variability by geochemical models, recent improvements of atmospheric-based methods either by reducing uncertainties in the <sup>13</sup>C method [*Fung et al.*, 1997] or by inversions that do not use <sup>13</sup>C data but only the spatial distribution of CO<sub>2</sub> [*Bousquet et al.*, 1999], show better agreement with ocean-based estimates such as the one presented here.

The use of SST and SSS data to reconstruct spatial and temporal variability of ocean  $p\text{CO}_2$  is shown to be reasonable in the equatorial Pacific. The developing ability to remotely sense SSS [*Lagerloef*, 1999] and of  $p\text{CO}_2$  algorithms similar to that presented here could provide useful tools for global mapping of time dependent  $p\text{CO}_2$  in the equatorial Pacific and possibly elsewhere.

**Table 1.** Comparison of  $\Delta p\text{CO}_2$  and CO<sub>2</sub> evasion rates

Region	Year	Field estimate ( $\mu\text{atm/GtC yr}^{-1}$ )	This study
<i>Non-ENSO conditions</i>			
10°S–10°N, 80°W–135°E	1984	60/0.6 (1)	50/0.7
10°S–10°N, 80°W–120°E	1984	60/0.8 (2)	51/0.8
10°S–10°N, 80°W–135°E	1989	50/0.4 (3)	44/0.7
5.5°S–5.5°N, 80.5°W–134.5°E	1989	79/1.0 (4)	53/0.5
<i>ENSO conditions</i>			
10°S–10°N, 80°W–135°E	1983	2/0.02 (1)	24/0.3
10°S–10°N, 80°W–135°E	1987	5/0.09 (3)	30/0.4
5.5°S–5.5°N, 80.5°W–135.5°E	1987	31/0.4 (4)	31/0.2
10°S–10°N, 80°W–135°E	1992	27/0.3 (5)	32/0.4

$\Delta p\text{CO}_2$  estimates are directly comparable while part of the discrepancies in CO<sub>2</sub> fluxes can be attributed to differences in formulations and wind forcing. Sources: 1, [*Feely et al.*, 1987]; 2, [*Volk et al.*, 1989]; 3, [*Wong et al.*, 1993]; 4, [*Inoue and Sigmura*, 1992]; 5, [*Feely et al.*, 1995].

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