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New insights of $p\text{CO}_2$ variability in the tropical eastern Pacific Ocean using SMOS SSS

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BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Abstract

Complex oceanic circulation and air–sea interaction make the eastern tropical Pacific Ocean (ETPO) a highly variable source of CO₂ to the atmosphere. Although the scientific community have amassed 70 000 surface partial-pressure of carbon dioxide ($p\text{CO}_2$) datapoints within the ETPO region over the past 25 years, the spatial and temporal resolution of this dataset is insufficient to fully quantify the seasonal to inter-annual variability of the region, a region where $p\text{CO}_2$ has been observed to fluctuate by > 300 μatm .

Upwelling and rainfall events dominate the surface physical and chemical characteristics of the ETPO, with both yielding unique signatures in sea surface temperature and salinity. Thus, we explore the potential of using a statistical description of $p\text{CO}_2$ within sea-surface salinity-temperature space. These SSS/SST relationships are based on in-situ SOCAT data collected within the ETPO. This statistical description is then applied to high resolution (0.25°) SMOS sea surface salinity and OSTIA sea surface temperature in order to compute regional $p\text{CO}_2$. As a result, we are able to resolve $p\text{CO}_2$ at sufficiently high resolution to elucidate the influence various physical processes have on the $p\text{CO}_2$ of the surface ETPO.

Normalised (to 2014) oceanic $p\text{CO}_2$ between July 2010 and June 2014 within the entire ETPO was 41 μatm supersaturated with respect to 2014 atmospheric partial pressures. Values of $p\text{CO}_2$ within the ETPO were found to be broadly split between southeast and a northwest regions. The north west, central and South Equatorial Current regions were supersaturated, with wintertime wind jet driven upwelling found to be the first order control on $p\text{CO}_2$ values. This contrasts with the southeastern/Gulf of Panama region, where heavy rainfall combined with rapid stratification of the upper water-column act to dilute dissolved inorganic carbon, and yield $p\text{CO}_2$ values undersaturated with respect to atmospheric partial pressures of CO₂.

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



1 Introduction

Perturbations to the global carbon cycle caused by anthropogenically driven increases in atmospheric partial pressures of CO₂ ($p\text{CO}_2$) produce an acute requirement to understand inter-reservoir carbon fluxes (Le Quéré et al., 2014). However, assessing these fluxes, especially the flux between the atmosphere and surface ocean is challenging, as $p\text{CO}_2$ within the oceanic reservoir varies considerably both spatially and temporally. Fortunately, considerable effort has been made in recording oceanic $p\text{CO}_2$ over the past 50 years, with roughly 10 million individual measurements of global surface ocean $p\text{CO}_2$ taken, processed, flagged and assembled into two large datasets: the surface ocean CO₂ atlas (SOCAT) and Lamont–Doherty Earth Observatory (LDEO) carbon dioxide database (Bakker et al., 2014; Takahashi et al., 2014). Both databases make heavy use of “vessel of opportunity” derived $p\text{CO}_2$ data, resulting in a heterogeneous dataset with the majority of measurements collected within the tight confines of commercial shipping lanes.

The first stage in efforts to estimate large scale air sea fluxes requires the extrapolation of these discrete surface $p\text{CO}_2$ observations over large areas of surface ocean. Completing a basic extrapolation of this vessel of opportunity based data results in a spatially-patchy $p\text{CO}_2$ field, as seen in the one degree gridded product available from SOCAT (Bakker et al., 2014). Therefore, in order to achieve improved spatial coverage, a frequently used solution is to fit data-driven diagnostic models (e.g. Takahashi et al., 2009; Park et al., 2010; Rödenbeck et al., 2013). These models use observed correlations between physical properties/temporal variation of the ocean’s surface, and the $p\text{CO}_2$ observed under these conditions. In addition, statistical criteria based on the surface ocean observations (for example, satellite imagery) and/or neural networks have been used to identify biogeochemical provinces in order to improve the accuracy of the extrapolated field (Boutin et al., 1999; Rangama et al., 2005; Landschützer et al., 2014). Although the quality of these extrapolation methods have been refined over the past few years, in part due to the increasing number of in situ measurements, the in-

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



terannual variability of the global air–sea CO₂ flux obtained using different data-driven methods still substantially differ, and further work is required to unify our estimates and improve understanding of this air–sea flux.

The recent (2010–present) availability of sea surface salinity from the Soil Moisture Ocean Salinity (SMOS) mission has provided a new tool in statistically modelled ρCO_2 studies, as now the relationship between surface temperature, salinity, density and ρCO_2 can be utilised in high resolution statistical descriptions of ρCO_2 within TS space. It is this relationship that is explored within this paper in order to quantify inter and intra annual variability within the oceanographically complex eastern tropical Pacific Ocean (ETPO) region between 4 and 18° N, and east of 95° W (Fig. 1).

The ETPO region is influenced by Northern and Southern Hemisphere trade winds, the doldrums, strong seasonal wind jets, heavy rainfall, strong solar heating and the El Nino-Southern Oscillation (ENSO), (Kessler, 2006). Within the ETPO the ITCZ is neither zonally oriented, nor spatially fixed over the course of a year, thus wind fields are highly variable in both strength and direction (Kessler, 2006). Further wind variability is introduced by low altitude jets blowing through three low elevation gaps in the Central American Cordillera, with these jets observed predominantly between November and February (Kessler, 2002; Fig. 1 – schematically represented by orange arrows). These strong jet winds are generated from the pressure gradient force resulting from high-pressure synoptic midlatitude weather systems transiting North America towards the low pressure Equatorial Pacific during the winter months (Chelton et al., 2001). Jet wind velocities of up to 20–30 ms⁻¹ have been observed within the three gulfs, extending from the shoreline to at least 500 km into the ETPO (Chelton et al., 2001). The alignment of these jets is mainly meridional in the Gulf of Panama and Tehuantepec, and more zonal in the Gulf of Papagayo (Fig. 1). These jet winds result in strong wintertime upwelling in each of the three basins, with the northerly jet winds in the Gulfs of Tehuantepec and Panama promoting Ekman upwelling (Kessler, 2006). At 9° N, 90° W, the quasi permanent anticyclonic Costa Rica thermocline dome is energised by the westerly jet winds in the Gulf of Papagayo, again, resulting in upwelling (Fig. 1).

New insights of ρCO_2 variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Kessler, 2002). However, the pressure disequilibrium that spawns these jet winds builds and subsides quickly (on a sub-weekly timescale), resulting in a highly fluctuating jet wind field, and nonlinearities in the oceanic (and hence upwelling) response to these episodic events (Kessler, 2006). Recent studies have shown that satellite SSS are very well suited to capture the variability of these high SSS events (Grotsky et al., 2014; Reul et al., 2013).

During summertime, these jet winds are rare, and the northward deflection of the ITCZ over the ETPO result in very high levels of precipitation, particularly within the Gulf of Panama (Fig. 1; Alory et al., 2012). The effect of this rainfall is a strong freshening and stratification of the surface, especially in the Gulf of Panama region. For this reason, the Gulf of Panama is often referred to as the “Pacific freshpool”, with salinities of < 30 frequently observed (Alory et al., 2012).

2 Data and methods

2.1 Observations of in-situ $p\text{CO}_2$, DIC and atmospheric CO_2

We use SOCAT data as the basis of a statistical description of $p\text{CO}_2$ within the ETPO. All CO_2 data within the SOCAT database are recorded as fugacity of CO_2 ($f\text{CO}_2$); a measure of $p\text{CO}_2$ that is corrected for the non-ideal-gas behaviour carbon dioxide displays (Bakker et al., 2014). In this study we will ignore the difference between $p\text{CO}_2$ and the fugacity of CO_2 , owing to the negligible difference between the two measurements (< 2 μatm in this study). The logistical importance of the region has proven beneficial, producing a large database of $p\text{CO}_2$ /SSS/SST observations, with ~ 70 000 surface carbon, temperature and salinity data-points collected between 1991 and 2011 (within the region depicted in Fig. 1). We use data collected between 1991 and 2013 within the LDEO v2013 database as a semi-independent dataset to test our SOCAT based statistical description of $p\text{CO}_2$ within TS space (Takahashi et al., 2014). The

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



LDEO database extends two years longer than SOCAT, making it a useful evaluation product.

In addition to surface $p\text{CO}_2$, we use dissolved inorganic carbon (DIC) concentrations within the ETPO measured during World Ocean Circulation Experiment cruise P19 (Fig. 1, WOCE data available from <http://woceatlas.ucsd.edu/>), and weekly dry-air CO_2 mole fractions ($x\text{CO}_2$) measured at Mauna Loa (www.esrl.noaa.gov). We derive $p\text{CO}_2$ in the atmosphere according to:

$$p\text{CO}_{2\text{air}} = x\text{CO}_2 (P_{\text{atm}} - p\text{H}_2\text{O}) \quad (1)$$

Where P_{atm} is the atmospheric pressure taken from the ERA interim product (<http://www.ecmwf.int/>) and $p\text{H}_2\text{O}$, the saturated water pressure (Weiss, 1974).

2.2 Calculating air sea fluxes of CO_2

The air–sea flux of CO_2 ($\text{mmol m}^{-2} \text{d}^{-1}$) is derived using the difference between surface ocean and atmospheric $p\text{CO}_2$ values, the solubility of CO_2 in standard seawater (α_{CO_2} calculated using the values given by Weiss, 1974) and the gas transfer velocity k_{CO_2} (cm h^{-1}):

$$F = k_{\text{CO}_2} \alpha_{\text{CO}_2} (p\text{CO}_{2\text{sea}} - p\text{CO}_{2\text{air}}) \quad (2)$$

k is calculated using the 10 m wind speed based parameterisation described by Sweeney et al., 2007, where:

$$k = 0.27U^2(660/Sc)^{0.5} \quad (3)$$

where Sc is the Schmidt number for CO_2 (Wanninkhof, 1992). We use the 0.25° , daily resolution 10 m wind speed product from the Advanced Scatterometer (ASCAT, www.knmi.nl/scatterometer/), and air $p\text{CO}_2$ is calculated assuming water vapour saturation at the boundary layer, then using the description in Eq. (1).

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



2.3 Satellite observations of SST and SSS

Sea surface temperatures have been continuously measured from space via satellites since the AVHRR mission in 1981, and sea surface salinity has been measured since the launch of the European Space Agency's Soil Moisture and Ocean Salinity (SMOS) satellite mission in November 2009 (Kerr et al., 2010). In this paper, we use satellite T and S data to help in the interpretation of spatial variability of $p\text{CO}_2$ from the ETPO.

SMOS SSS maps at 0.25° resolution (running average over $100 \times 100 \text{ km}^2$) produced by the LOCEAN SMOS group (combining SMOS ascending and descending passes) were used within this paper (Boutin et al., 2013, data available at <http://catds.ifremer.fr/Products/Available-products-from-CEC-OS/Locean-v2013>). SMOS SSS data are not used prior to June 2010 due to variations in sensor configuration tested during the in-orbit sensor commissioning phase (Corbella et al., 2011). Data from SMOS within the ETPO has been used previously in a process study, with SSS found to perform well and featuring a low noise to signal ratio (Alory et al., 2012).

SSS was combined with sea surface temperatures from the operational sea surface temperature and sea ice analysis (OSTIA) system; an optimal interpolation of multiple microwave and infrared satellite-based data sources (Donlon et al., 2012). This SST daily product features a native resolution of 0.05° , which was subsequently re-sampled over the same spatial and temporal grid as the SMOS data. As auxiliary datasets for near surface T and S, data from the Argo float array were used.

2.4 The basis of using a SST/SSS statistical $p\text{CO}_2$ model in the ETPO

Using a $p\text{CO}_2$ model based on surface T and S properties requires the carbon properties of water masses to be quantified. We define three surface water-masses within the study region, based on the near-surface (0–100 m depth) T and S characteristics of the ETPO recorded by ~ 6000 Argo profiles between 17 and 4° N , and east of 95° W (Fig. 2). These water-masses are also independently featured within studies by Kessler (2002, 2006):

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$
variability in the
ETPO

C. W. Brown et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

1. An ETPO surface water, with temperatures $> 27^\circ\text{C}$ and salinities > 33.5 (Fig. 2, “B”).
2. Deep water that predominantly exists below the thermocline (defined by Kessler as the 20°C isotherm) and is only expressed at the surface of the ETPO during periods of upwelling (Fig. 2, “C”).
3. Rain influenced water, with salinities < 33.5 (Fig. 2, “A”).

Here, the water-mass at “C” (referred to as “deep water” herein) is predominantly found at depths greater than 80 m, whilst the near surface water is typically ETPO surface water and warmer than 27.5°C (“A” and “B”). Occasionally, the deep water-mass can be observed at shallower depths, advected upwards during times of strong upwelling (Fig. 2). Rainfall influenced waters display a wide range of salinities (from 29 to 34.5) within the upper 20 m (as observed in the salinity variation between “A” and “B”). At low salinities, densities as low as 1018 kg m^{-3} are observed; an eight kg m^{-3} difference between this surface water and deep water (featuring densities of 1026 kg m^{-3}). This large density gradient is an important feature of the ETPO, as periods when stratification and high thermoclines strengths are prevalent, the upwelling of deep water is inhibited (Fiedler and Talley, 2006).

To identify the carbon properties of these water-masses, DIC concentrations between the surface and 100 m depth within the ETPO measured during WOCE transect P19 are shown as red circles and associated concentrations plotted within TS space (Fig. 2). Deep water “C” was observed to have DIC concentrations $> 2200\text{ }\mu\text{mol kg}^{-1}$, contrasting values between “A” and “B” of $< 1950\text{ }\mu\text{mol kg}^{-1}$. End-member mixing occurs at intermediate temperatures between the two watermasses “B” and “C”, exhibiting in the gradient observed along the 34.2–34.75 isohalines in Fig. 2. Assuming that upwelling results in the expression of deep water at the surface, upwelling will act to increase the surface inventory of DIC, leading to increased $p\text{CO}_2$.

Rainfall results in surface layer dilution, thus reducing DIC between “A” and “B” (Turk et al., 2010). When the $p\text{CO}_2$ of both endmembers in this system (rainwater and sur-

face waters) are at equilibrium with the atmosphere, this dilution results in a lowering of $p\text{CO}_2$ within the sea surface, and ingassing of CO_2 . Turk et al.'s (2010) study in the western Pacific warm pool have indicated that decreases of 30–40 μatm in surface $p\text{CO}_2$ values can result from rainfall alone, with these effects strongest under highly stratified conditions. This suggests that rainfall could be an important influence on ETPO $p\text{CO}_2$, particularly within the Panama Basin region. Finally, although biological processes influence DIC (and hence $p\text{CO}_2$), observations of DIC within this region suggest that physical processes (such as upwelling or rainfall) are the first order control on DIC. For this reason, coupled to the lack of net community production data within the region, we have elected to concentrate our efforts on quantifying these physical processes.

2.5 Processing SOCAT data

Surface $p\text{CO}_2$ responds to changing levels of atmospheric $p\text{CO}_2$, which is increasing at approximately 20 μatm per decade, (Takahashi et al., 2009; Fig. 3). In order to calculate contemporary air sea fluxes, the multi-year SOCAT data must first be temporally corrected for the inter-annual trend of increasing atmospheric CO_2 . The mean annual increase in $p\text{CO}_2$ within the central equatorial Pacific (5°N and 5°S) between 1979 and 1990 (excluding El Niño events) was estimated at a rate of $1.1 \mu\text{atm} \pm 0.3$ per year (Takahashi et al., 2009). A second estimate, using data collected between 1990 and 2003 (within the region 10°N to 5°S) calculated an annual increase of $2.0 \mu\text{atm} \pm 0.2$ (Takahashi et al., 2009). Correcting 1991 data to contemporary July 2014 values would equate to a not-inconsiderable difference depending on which of these two estimates was chosen. Therefore, in order to choose the optimum correction for the smaller sub-region of the ETPO, all available SOCAT $p\text{CO}_2$ data 1991–2011 were binned by year the data was collected (Fig. 3). Fitting a linear regression to the average bin value of this data, resulted in an average annual increase of $1.95 \mu\text{atm yr}^{-1}$ ($\pm 0.38 \mu\text{atm}$). This linear regression was found to fall within the 5th and 95th percentile values for all bins, with variability in mean and median values caused by heterogeneous sampling of the

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



region and inter-annual variability (Fig. 3, percentiles plotted in blue). Our calculated rate of increase is both within the range of the annual atmospheric $p\text{CO}_2$ increase measured at Mauna Loa, Hawaii ($1.8 \mu\text{atm yr}^{-1} \pm 0.1 \mu\text{atm}$) suggesting that $p\text{CO}_2$ within the ETPO tracks the rate of increasing atmospheric $p\text{CO}_2$ (Fig. 3). Therefore, we apply a $1.95 \mu\text{atm}$ per year correction to the SOCAT data in order to normalise the $p\text{CO}_2$ data to 1 July 2014 values, with all results herein also normalised to this date.

Although the spatial distribution of $p\text{CO}_2$ SOCAT observations within the ETPO are irregular, the variability can be constrained once the TS properties of the surface ocean are accounted for (as-per DIC, Figs. 2 and 4). Data from the upper 100 m of Argo profiles completed in each of the three gulfs (Panama, Papagayo and Tehuantepec, as defined in Fig. 1) show strong TS similarities between each of the three gulf regions; low salinity (< 34) waters are observed in each gulf, with a thermocline separating surface-water and deep water at 80–100 m depth. The T–S properties of these Argo profiles are replicated at the surface, within SOCAT SST and SSS data (Fig. 4). However, the relative numbers of observations that fall into each water-mass definition is different. The warm ETPO surface water is dominant in all three regions for the majority of samples, but cooler ($< 20^\circ\text{C}$) deep water is occasionally expressed at the surface, with this signal strongest in the Gulf of Tehuantepec, followed by the Gulf of Papagayo, and seldom observed in the Gulf of Panama. This contrasts observations of low salinity water, which are very frequent within the Gulf of Panama, but rare in the other two gulfs (Fig. 4). Although both SOCAT and Argo data suffer from sparse sampling, we suggest that both datasets indicate that the relative dominance of each water-mass within each of the three gulfs is different. There are very few SOCAT observations within the South Equatorial Current (SEC) region (Fig. 1), thus we do not show a TS diagram from this region, however Argo profiles from this region are included in Fig. 2, highlighting that the SEC also shares the same watermasses as the three gulf regions.

The distribution of DIC concentrations within TS space is mirrored by $p\text{CO}_2$ values; deep water features higher $p\text{CO}_2$ values than ETPO surface water, and low salinity water feature the lowest $p\text{CO}_2$ values. The intermediate values of $p\text{CO}_2$ observed

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



between each of these watermasses suggest end-member mixing of $p\text{CO}_2$ between waters-masses. Finally, there are no watermasses that share TS characteristics, but feature very different DIC concentrations or $p\text{CO}_2$ values, meaning that DIC/ $p\text{CO}_2$ values are unique within TS space in this region (Figs. 2 and 4). Therefore, as both DIC and $p\text{CO}_2$ values behave pseudo-conservatively within TS space, it is possible to construct a statistical description of $p\text{CO}_2$ using solely SSS and SST.

2.6 Fitting a statistical description – the look-up-table

The similarities of T,S and $p\text{CO}_2$ properties of the water-masses observed both by Argo and within the SOCAT database across all three gulfs enable data from the entire ETPO to be considered as a single system (Fig. 5). By combining all SOCAT data from the ETPO onto a single T–S diagram, a look-up-table (LUT) can be produced in order to describe $p\text{CO}_2$ as a function of T and S (Fig. 5). The LUT was constructed by completing a linear interpolation of binned $0.1(\text{salinity}) \times 0.1^\circ\text{C}$ SOCAT data from the entire ETPO region. The number of observations per bin is depicted in the right-hand panel. Here, the highest number of observations fall in a narrow region of TS space, between salinities of 32–34.5 and 25–30°C (Fig. 5). We tested the quality of fit of the SOCAT based LUT, using LDEO v2013 data. Here $p\text{CO}_2$ values were computed using the LUT using T and S from the LDEO v2013 database, with this computed $p\text{CO}_2$ data compared to the measured (annual increase corrected) $p\text{CO}_2$ values. The LUT could be applied to 96.2% of all LDEO v2013 T–S measurements made between 1990 and 2013, with the remaining TS measurements falling outside the TS boundaries of the LUT.

The average root-mean squared error (RMSE) of the LUT was $16.8\ \mu\text{atm}$ between the LUT computed and measured $p\text{CO}_2$. This RMSE was asymmetrically distributed; highest at temperatures of 22–26°C/salinities of 34.4, and lowest at warmer temperatures/lower salinities and also at colder temperatures (Fig. 5). This suggests that $p\text{CO}_2$ variability is low in aged surface waters, and high in recently upwelled and warming water. As there are a number of measurements within the region of higher error (Fig. 5),

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



we hypothesize that this larger uncertainty is not due to lack of observations, but rather to mixing/heating processes that allow water with slightly different $p\text{CO}_2$ values to occupy the same TS space. For example, the $p\text{CO}_2$ value of water at salinities of 34.5 and 25 °C could either result from the stoichiometric mixing of two waters at salinities of 34.5, and temperatures of 18 and 32 °C, or solely from warmed 34.5/18 °C water. Upwelling, solar radiation, horizontal advection, biological productivity and diapycnal mixing processes all influence this system, and need to be accounted for in order to quantify completely the formation and $p\text{CO}_2$ observed within this water, but are outside the scope of this study using the data available in this region. However, this LUT technique works efficiently in determining the first-order variability of the system within the ETPO.

3 Results

3.1 Bi-monthly variability along ship tracks

Bimonthly $p\text{CO}_2$ variability can be seen in the (1991–2011 annual $p\text{CO}_2$ increase corrected) SOCAT observations plotted in Fig. 6. Here, the top six panels display SOCAT observations of $p\text{CO}_2$, and the lower six panels display $p\text{CO}_2$ calculated from SOCAT SST and SSS observations using the LUT. As the majority of SOCAT data is collected using ships of opportunity, resolution within shipping lanes in the ETPO is excellent; (for example multiple observations are made in regions south and west of the Panama Canal, at 9.1° N, 79.7° W), but very sparse outside of these shipping lanes. However, acknowledging these issues in data resolution, patterns in $p\text{CO}_2$ can still be detected. The lowest $p\text{CO}_2$ is typically observed within the Gulf of Panama and close to the coast. Higher $p\text{CO}_2$ is observed in the Gulfs of Tehuantepec and Papagayo. Inter seasonal variability is also observed, with the highest $p\text{CO}_2$ (> 440 μatm) occurring between November to February, with lower $p\text{CO}_2$ values (< 400 μatm) occurring across the entire region during the summer months (May–August). The data also show

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



a very high degree of variability at small spatial and temporal scales, for example during November- December within the Gulf of Papagayo. It is this variability that needs to be resolved using the LUT, when coupled to satellite SST and SSS observations at native resolution. Thus, it is important to confirm that the LUT is able to recreate this variability in $p\text{CO}_2$ using the initial (SOCAT) T and S conditions. Here, we find that the LUT performs well, with most of the measured $p\text{CO}_2$ variability also observed within the $p\text{CO}_2$ calculated using the LUT (Fig. 6).

3.2 $p\text{CO}_2$ and fluxes by region – the influence of wind and upwelling

Using the LUT, SMOS and OSTIA data, the small scale variability and features within the ETPO can be resolved more thoroughly than possible through SOCAT data alone. Bimonthly averaged (between July 2010–June 2014) SSS, SST, $p\text{CO}_2$, CO_2 flux and windspeed are plotted in Fig. 7. Jet wind velocities over the Gulf of Tehuantepec are at their peak between October to February, thus optimising Ekman upwelling. During these months the SSS of the Gulf increases, whilst the SST decreases (due to the influences of upwelling). Increased $p\text{CO}_2$ is observed across most of the Gulf of Tehuantepec, with peak outgassing occurring as a narrow band, centred underneath the axis of the jet wind. We suggest that this is due to the complimentary nature of ocean physics in this region – high windspeeds promote Ekman upwelling of high DIC deep water, whilst increasing the k component within the air–sea flux parameterisation, thereby maximising CO_2 outgassing.

4 Discussion

The Gulf of Papagayo shares many similarities with the Tehuantepec Gulf, however, jet winds are more zonally aligned (reducing Ekman pumping strengths, Alexander et al., 2012), and occur later into winter compared to Tehuantepec; being strongest between November–February. This results in a lag between peak wintertime $p\text{CO}_2$ values seen

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



within the two gulfs (Fig. 7). The Gulf of Papagayo also features elevated $p\text{CO}_2$ values during the summer months which are not seen in either the Gulf of Tehuantepec or Panama. This is due to westerly winds maintaining the vorticity of the Costa Rican dome structure, enabling the continued upwelling of deep water at the core of this dome throughout the summer season (Grotsky et al., 2014; Kessler, 2006; Fig. 7).

The seasonally averaged wind velocities in the Gulf of Panama superficially resemble the wind patterns in the two northern gulfs- jet winds are observed between January-February (Fig. 7). However, excluding a small area of ocean directly underneath the wind jet axis during winter, the region remains a small net sink of carbon throughout the year (Fig. 8). We suggest that this contrast between the Panama gulf and the rest of the ETPO is due to the high rainfall within this region, resulting in the dilution of DIC as described above.

4.1 $p\text{CO}_2$ and the influence of rainfall

A significant proportion of atmospheric water exported from above the Atlantic basin into the Pacific basin is precipitated into the ETPO. This rainfall is intensified within the ETPO during the summertime, due to a northwards shift of the ITCZ towards the Panama coast (Xie et al., 2005). As a result, the Gulf of Panama receives net precipitation of 180–220 cm per year, with peak rainfall of 20 mm per day during July-August (Alory et al., 2012). This large freshwater flux, coupled with light southerly winds between March and June, and the southerly (thus downwelling promoting), winds between July and December result in the semi-permanent stratification of the water-column (Fig. 7, Alory et al., 2012). Qualitatively, this stratification can be observed in the SOCAT and Argo data in Fig. 4 by the scarcity of deep water observations in the Gulf of Panama compared to the Gulfs of Tehuantepec and Papagayo. Additionally, the thermocline is observed at deeper depths in the Argo profiles taken in the Gulf of Panama, compared to either the Gulf of Tehuantepec or Papagayo (Fig. 4).

The influence of stratification/rainfall on $p\text{CO}_2$ values are seen in Fig. 7. Here, the lowest salinity and $p\text{CO}_2$ are observed during the summer months (during peak rain-

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



fall season, Alory et al., 2012). This low $p\text{CO}_2$, stratified system persists until January, when intensification of the south equatorial current results in the export of the low $p\text{CO}_2$ /fresh surface layer towards the south west (as seen by the elongation of the freshpool during January–March SSS). This export of water, coupled with the dry-season within the Panama Gulf appears to weaken the stratification, with the result that sporadic jet winds enable Ekman upwelling, thus increasing SSS (Alory et al., 2012; Fig. 7) and average $p\text{CO}_2$ values between January–April (Fig. 8).

4.2 Air–sea fluxes from July 2010–June 2014

The ETPO between 2010–June 2014 had an annual average $p\text{CO}_2$ of $425 \mu\text{atm}$ ($\Delta p\text{CO}_2$ $41 \mu\text{atm}$) and a outgassing CO_2 flux of $1.6 \text{ mmol m}^{-2} \text{ d}^{-1}$ ($0.59 \text{ mol m}^{-2} \text{ yr}^{-1}$, Fig. 8/Table 1). The ETPO, (the boundaries of which defined as per Fig. 1), has an area of $1.9 \times 10^{12} \text{ m}^2$, therefore, net outgassing from the ETPO equates to $11 \times 10^{12} \text{ g}$ (0.01 Pg) of carbon per year.

To examine both monthly and inter-annual variability in air–sea fluxes, the ETPO is split into subregions; the Gulfs of Tehuantepec, Papagayo, Panama and the South Equatorial Current (Figs. 1 and 8 and Table 1). Here, although variable, the consistent seasonal cycle between summer and winter observed in the bimonthly data in Fig. 7 is present for each individual year. Peak outgassing within the Gulfs of Tehuantepec, Papagayo and the South Equatorial Current occurs during winter (November–February). This contrasts the Gulf of Panama, which is a net sink of atmospheric CO_2 except during occasional upwelling events during January–February. However, ingassing within the Gulf of Panama is insufficient to offset outgassing from the rest of the ETPO and the region remains a net source of CO_2 year round.

In addition to seasonal variability, interannual variability is observed within the ETPO. Across the entire basin, the ETPO during the years 2010–2012 were on average cooler and saltier than subsequent years (especially 2012–2013). This resulted in higher annually averaged outgassing and $p\text{CO}_2$ values (Fig. 8, Table 1). Concurrently, during 2010–2012, the Gulf of Panama featured a low salinity mimima (of 28), and a small

freshpool footprint (the boundary of which is defined by the 33 isohaline, Alory et al., 2012). The extreme low salinity observed within the eastern region of Gulf of Panama during 2010 to 2012 resulted in lower $p\text{CO}_2$ and hence stronger ingassing than 2012–2014 values, highlighting the effect of rainwater dilution of DIC. However, this localised ingassing was insufficient to offset the higher outgassing observed in the other regions of the ETPO (Fig. 8, Table 1). Also noteworthy is that average wind velocities were similar during 2010 and 2012 compared to subsequent years, so the increased outgassing observed during these years cannot be attributed to increased wind-mediated Ekman upwelling of deep water alone.

We note that 2010 to the end of 2011 featured La Niña conditions, and suggest that this could be causal to the differences in CO_2 fluxes, higher salinities and higher $p\text{CO}_2$ observed during this time period. A study using NCEP (National Centers for Environmental Prediction) reanalysis data within this region observed cooler waters and more frequent upwelling of sub-thermocline deep water during La Niña events. A shallower thermocline depth exists during La Niña events, which results from the uplift of the water-column by intensified coastal Kelvin waves. Mechanistically, this shallow thermocline reduces the strength of upwelling required for the expression of deep water at the surface, thus decreasing average SSTs observed during La Niña periods (Alexander et al., 2012).

Furthermore, Alexander et al.'s (2012) study also suggests that although jet winds are the first order control on SST and thermocline depth during winter; it is changes in the thermocline depth (rather than changes in jet winds) that result in observed El Niño Southern Oscillation (ENSO) variability in SST within the ETPO. Variability in thermocline depth, and by association, variability in the ease by which deep water can be advected towards the surface suggest that (ENSO) could drive variability within the surface $p\text{CO}_2$ observations. However, with only four years of data, and no El Niño phase for intercomparison we are unable to draw definitive conclusions.

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



4.3 Previous work

The results in Figs. 7 and 8 represent the first attempt to quantify $p\text{CO}_2$ within the ETPO region at high resolution from observations. Previous work encompassing the ETPO include three basin wide or global $p\text{CO}_2/p\text{CO}_2$ studies, by Takahashi et al., (2009), Ischii et al. (2014) and Landschützer et al. (2014). These studies, as discussed in the introduction, are based on the extrapolation of $p\text{CO}_2$ directly, or the extrapolation of $p\text{CO}_2$ using a neural network technique, and feature spatial resolutions of $4^\circ \times 5^\circ$, $4^\circ \times 5^\circ$ and $1^\circ \times 1^\circ$ respectively. The resolutions used in the Takahashi et al. (2009) and Ischii et al. (2014) studies result in the ETPO being depicted by only 3 points, which, although able to separate the low $p\text{CO}_2$ Panama Gulf region from the higher values observed in the Gulfs of Tehuantepec and Papagayo, are too coarse to identify any features (such as upwelling) within the region. The improved resolutions featured in Landschützer et al. (2014) resulted in some mesoscale features being observed, such as the increased $p\text{CO}_2$ values within the Gulf of Papagayo. However, this work was not able to identify the low $p\text{CO}_2$ conditions within the Gulf of Panama, most likely due to the same biogeochemical province description being applied to the entire ETPO. In addition to these studies, our work distinguishes the importance of jet winds in increasing $p\text{CO}_2$, and quantifies the strong inter annual variability within the region.

5 Conclusions

Estimating surface $p\text{CO}_2$ and air sea fluxes of CO_2 within the global oceans has advanced considerably over the past decade, assisted by the assembly of large standardised atlases of surface observations (such as the SOCAT database). However, although these databases boast measurements in the millions, a challenge remains in gauging seasonal or sub-mesoscale $p\text{CO}_2$ variability in the oceans. Sampling though the use of commercial volunteering observation vessels may introduce bias, with most data existing within the narrow confines of the major global shipping lanes or for only

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



a few months of the year. However, the quantification of $p\text{CO}_2$ within TS space, used in conjunction with observations of surface SST and SSS (made possible through the recent availability of SSS from satellite) has proved highly useful in improving our understanding of $p\text{CO}_2$ variability at much improved spatial and temporal resolutions. We have demonstrated a technique using SOCAT data to identify the $p\text{CO}_2$ signatures of water-masses within the ETPO region, namely, the high $p\text{CO}_2$ deep water, the near equilibrium ETPO surface water, and the undersaturated rainfall diluted surface waters. From this, we used a LUT technique, in order to produce a description of the $p\text{CO}_2$ content of an ETPO surface water using satellite SST and SSS.

The highest outgassing and surface $p\text{CO}_2$ were observed during the winter period (November–March), in the Gulfs of Tehuantepec, Papagayo and in the south equatorial current. The first order control on these upwelling events and hence $p\text{CO}_2$ in the ETPO are strong wind jets blowing through low altitude gaps in the Central American cordillera. The Gulf of Panama remained net undersaturated on average, due to dilution effects of heavy rainfall and the stratification of the water column. Although wind jets were observed in the Gulf of Panama, the exceptionally low density of the water within this region appears to limit upwelling, and any upwelling that occurs is directly underneath the wind jet axis. Inter-annual variability was observed within the region, with the location of the western extent of the freshpool moving westwards considerably between 2010 and 2014. Previous work within this region suggest that changes in thermocline depth related to ENSO are likely to influence $p\text{CO}_2$ within this region. The region is a net contributor to atmospheric CO_2 , with average sea to air fluxes (over the four years of observations) of $1.6 \text{ mmol m}^{-2} \text{ d}^{-1}$, with all regions of the ETPO outgassing year-round, except the rainfall diluted Gulf of Panama/Freshpool region.

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New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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5

BGD

12, 4595–4625, 2015

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Table 1. Annual averaged values for each region, as reported in Fig. 8.

Date/ Location	$\Delta p\text{CO}_2$ (μatm)	CO_2 Flux ($\text{mmol m}^{-2} \text{d}^{-1}$)	Windspeed (m s^{-2})	Salinity (PSU)	Temperature ($^{\circ}\text{C}$)
2010–2011 ETPO	41.2	1.58	4.81	33.3	27.9
2010–2011 SEC	41.7	1.67	5.18	33.2	27.7
2010–2011 Panama	−4.1	−0.18	4.57	28.3	27.4
2010–2011 Papagayo	34.1	0.97	4.17	33.7	28.2
2010–2011 Tehuantepec	40.9	1.58	3.72	33.6	29.0
2011–2012 ETPO	44.2	1.59	4.59	33.3	27.9
2011–2012 SEC	44.2	1.66	4.97	33.1	27.7
2011–2012 Panama	−3.6	−0.20	4.02	28.4	27.5
2011–2012 Papagayo	35.4	1.04	4.01	33.7	27.9
2011–2012 Tehuantepec	49.5	1.66	3.90	33.7	28.4
2012–2013 ETPO	36.6	1.48	4.88	33.2	28.6
2012–2013 SEC	37.0	1.54	5.11	33.0	28.3
2012–2013 Panama	−2.0	−0.10	3.92	28.7	27.3
2012–2013 Papagayo	33.2	1.16	4.49	33.6	28.5
2012–2013 Tehuantepec	33.8	1.52	4.08	33.6	29.3
2013–2014 ETPO	40.7	1.56	4.67	33.3	28.0
2013–2014 SEC	39.5	1.65	5.09	33.0	27.8
2013–2014 Panama	−2.2	0.09	3.88	27.5	27.7
2013–2014 Papagayo	38.0	1.18	4.15	33.7	28.0
2013–2014 Tehuantepec	47.3	1.58	3.92	33.7	28.6

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

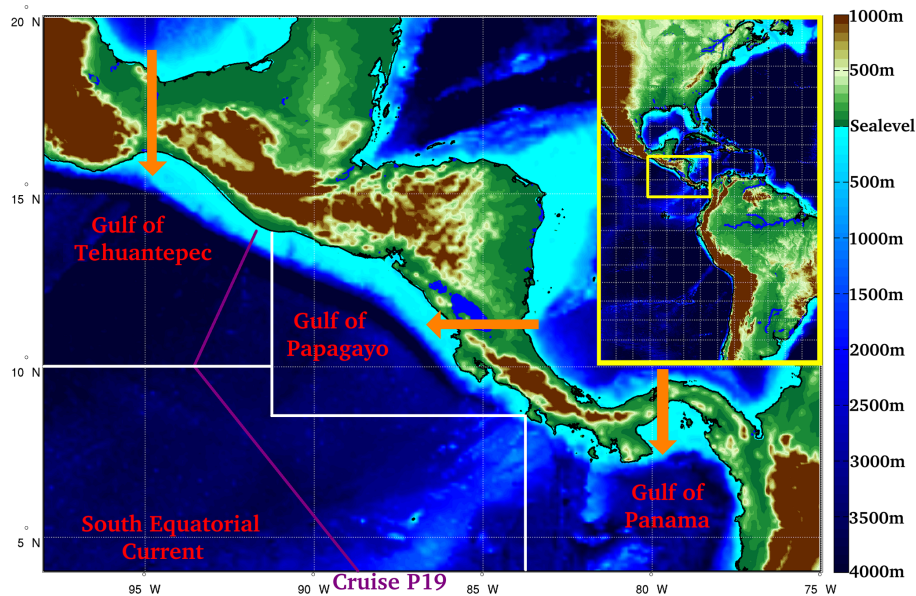


Figure 1. The topography of the Eastern Tropical Pacific Ocean and Isthmus of Panama, plotted using GEBCO bathymetry. The three gulfs with the region, Panama, Papagayo and Tehuantepec, and the South Equatorial Current are marked in white. The transect of WOCE cruise P19 is indicated in purple. The path of jet winds are marked by orange arrows.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



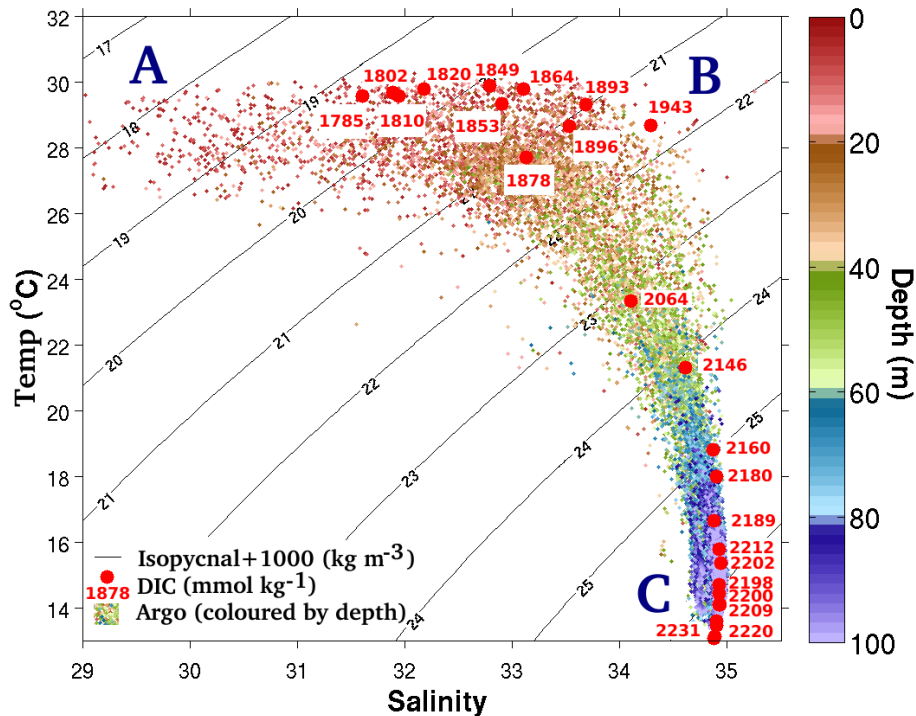


Figure 2. The total library of Argo profiles collected within the ETPO in the upper 100 m. Numbered red dots indicate the total dissolved carbon concentration at specific TS values as measured during cruise WOCE cruise p19.

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



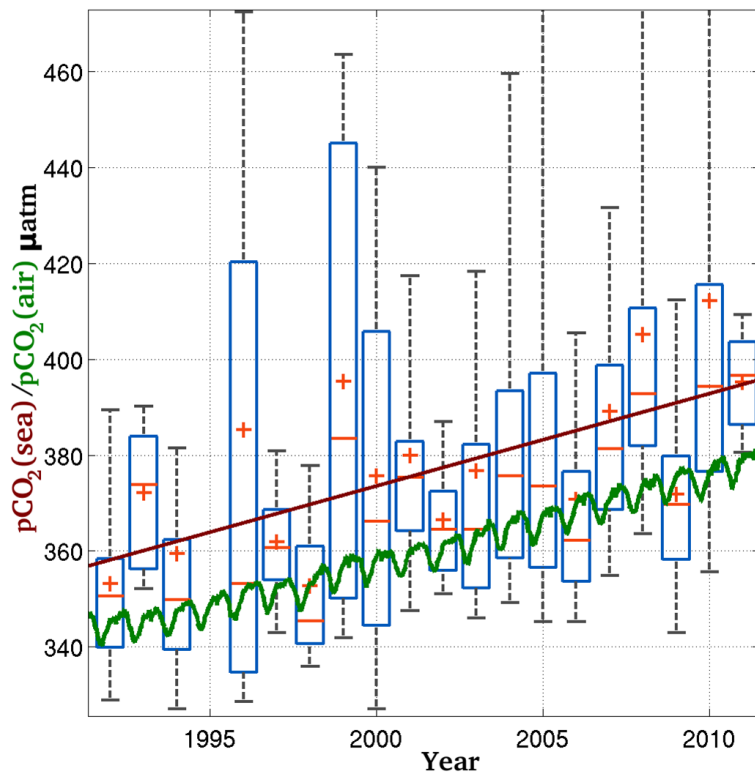


Figure 3. Yearly binned $p\text{CO}_2$ measurements from the SOCAT database. The interpolated average rate of ETPO $p\text{CO}_2$ increase is shown as a red line, with the Mauna Loa $p\text{CO}_2$ data shown as a green line. The blue boxes represent the 5th and 95th percentile $p\text{CO}_2$, with the small red lines indicating yearly averages, and the red crosses yearly median.

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

Back Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

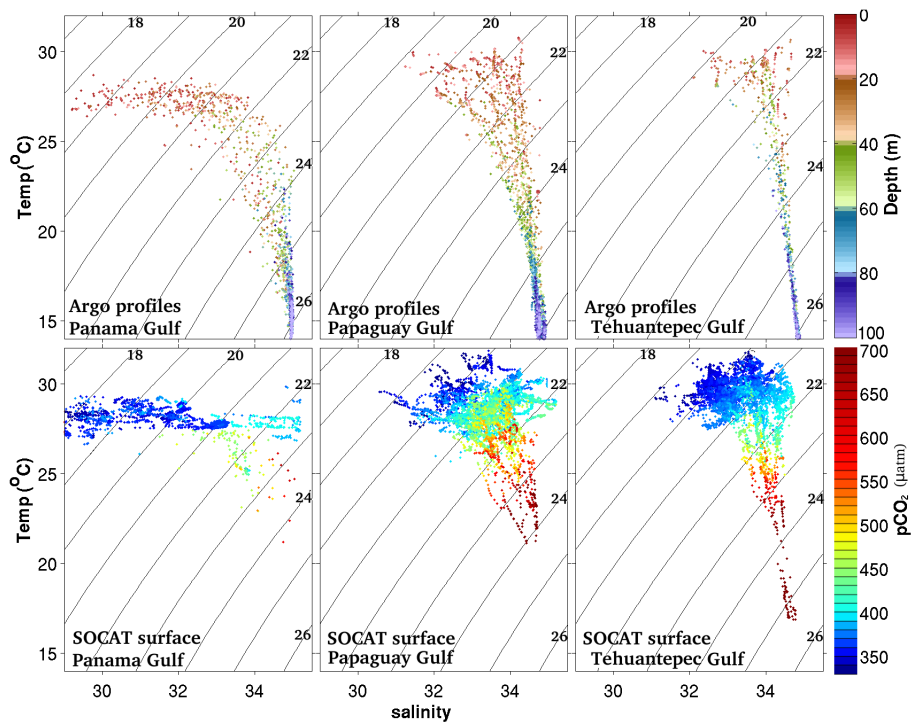


Figure 4. TS data from the three main gulfs within the ETPO both from Argo floats (coloured by depth-top three plots) and from surface SOCAT T and S data, (coloured by $p\text{CO}_2$ -bottom three plots).

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

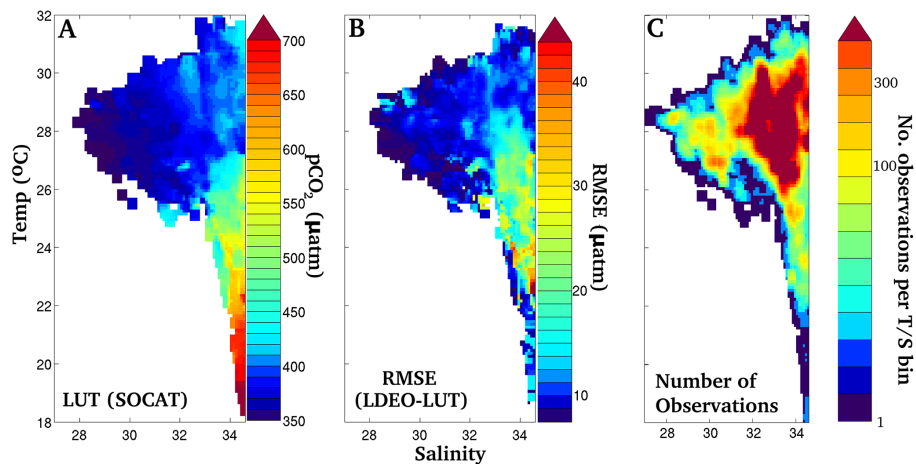


Figure 5. (a) The Look Up Table derived from ETPO SOCAT $p\text{CO}_2$ measurements. (b) The root mean squared error of the LUT–LDEO $p\text{CO}_2$ observations. (c) The number of LDEO measurements per $0.1^\circ \times 0.1^\circ$ salinity/temperature bins.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

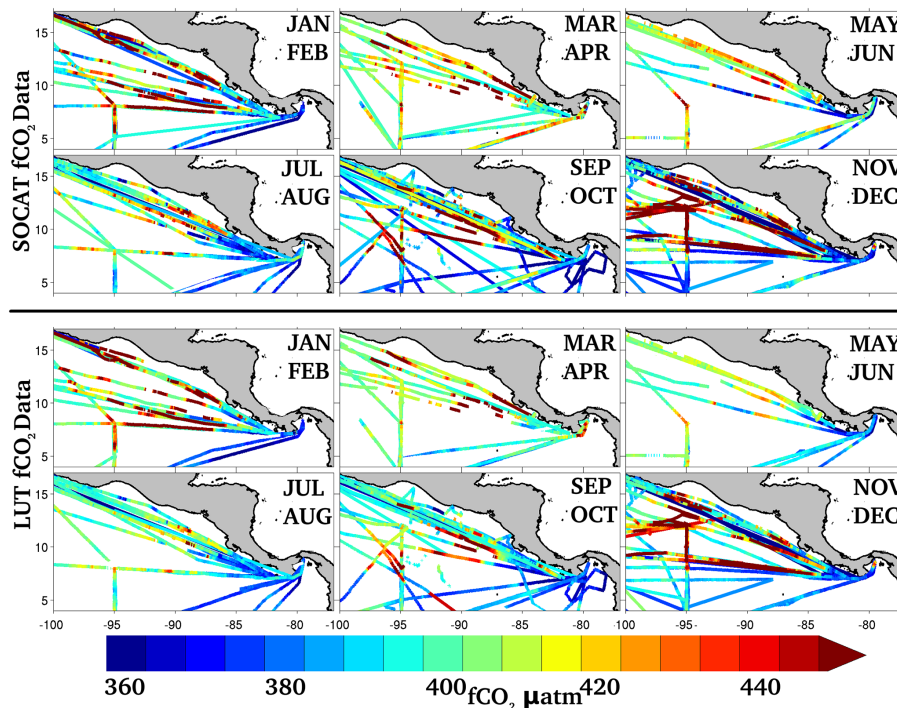


Figure 6. SOCAT $p\text{CO}_2$ bimonthly (January and February, March and April, May and June, July and August, September and October, November and December) observations within the ETPO corrected for the annual CO_2 increase. Bottom rows: LUT derived bimonthly $p\text{CO}_2$ calculated using SOCAT SST and SSS observations.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



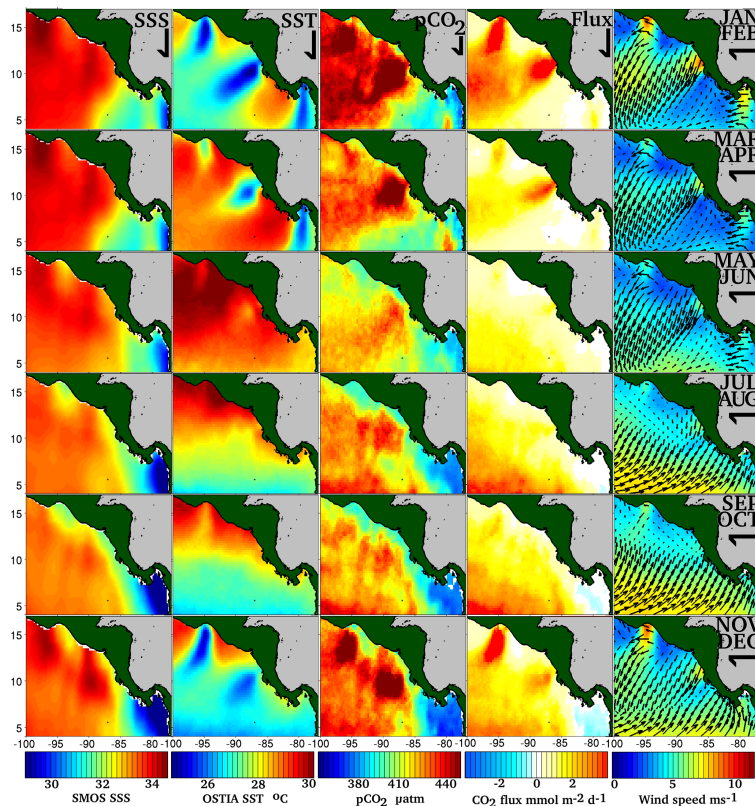


Figure 7. July 2010–June 2014 average SSS, SST, $p\text{CO}_2$, air–sea fluxes and wind vectors for the ETPO, split bimonthly (January and February, March and April, May and June, July and August, September and October, November and December).

New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

[Title Page](#)

[Abstract](#) [Introduction](#)

[Conclusions](#) [References](#)

[Tables](#) [Figures](#)

[◀](#) [▶](#)

[◀](#) [▶](#)

[Back](#) [Close](#)

[Full Screen / Esc](#)

[Printer-friendly Version](#)

[Interactive Discussion](#)



New insights of $p\text{CO}_2$ variability in the ETPO

C. W. Brown et al.

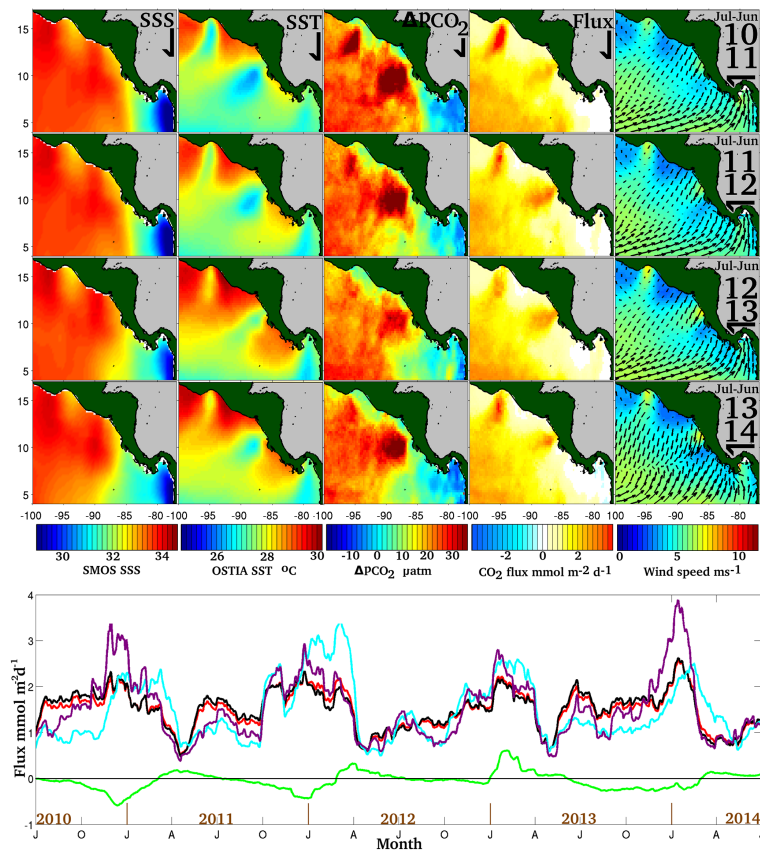


Figure 8. Upper: yearly average SSS, SST $\Delta p\text{CO}_2$, air-sea fluxes and wind vectors for the ETPO for July to June 2010 and 2011, 2011 and 2012, 2012 and 2013 and 2013 and 2014. Lower: the continuous $p\text{CO}_2$ fluxes from the entire ETPO (red line), the Gulfs of Tehuantepec (purple), Papagayo (blue), Panama (green) and the South Equatorial Current (black).

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

