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# Air-sea $CO_2$ flux in the Pacific Ocean for the period 1990–2009

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## Abstract

Air-sea $CO_2$ fluxes over the Pacific Ocean are known to be characterized by coherent large-scale structures that reflect not only ocean subduction and upwelling patterns, but also the combined effects of wind-driven gas exchange and biology. On the largest scales, a large net $CO_2$ influx into the extra-tropics is associated with a robust seasonal cycle, and a large net $CO_2$ efflux from the tropics is associated with substantial inter-annual variability. In this work, we have synthesized estimates of the net air-sea $CO_2$ flux from a variety of products drawing upon a variety of approaches in three sub-basins of the Pacific Ocean, i.e., the North Pacific extra-tropics (18° N–66° N), the tropical Pacific (18° S–18° N), and the South Pacific extra-tropics (44.5° S–18° S). These approaches include those based on the measurements of $CO_2$ partial pressure in surface seawater ($p$$CO_2$sw), inversions of ocean interior $CO_2$ data, forward ocean biogeochemistry models embedded in the ocean general circulation models (OBGCMs), a model with assimilation of $p$$CO_2$sw data, and inversions of atmospheric $CO_2$ measurements. Long-term means, inter-annual variations and mean seasonal variations of the regionally-integrated fluxes were compared in each of the sub-basins over the last two decades, spanning the period from 1990 through 2009. A simple average of the long-term mean fluxes obtained with surface water $p$$CO_2$ diagnostics and those obtained with ocean interior $CO_2$ inversions are $-0.47 \pm 0.13\,\mathrm{Pg\,C\,yr^{-1}}$ in the North Pacific extra-tropics, $+0.44 \pm 0.14\,\mathrm{Pg\,C\,yr^{-1}}$ in the tropical Pacific, and $-0.37 \pm 0.08\,\mathrm{Pg\,C\,yr^{-1}}$ in the South Pacific extra-tropics, where positive fluxes are into the atmosphere. This suggests that approximately half of the $CO_2$ taken up over the North and South Pacific extra-tropics is released back to the atmosphere from the tropical Pacific. These estimates of the regional fluxes are also supported by the estimates from OBGCMs after adding the riverine $CO_2$ flux, i.e., $-0.49 \pm 0.02\,\mathrm{Pg\,C\,yr^{-1}}$ in the North Pacific extra-tropics, $+0.41 \pm 0.05\,\mathrm{Pg\,C\,yr^{-1}}$ in the tropical Pacific, and $-0.39 \pm 0.11\,\mathrm{Pg\,C\,yr^{-1}}$ in the South Pacific extra-tropics. The estimates from the atmospheric $CO_2$ inversions show large variations amongst different inversion systems, but their median fluxes are consistent with the estimates from climatological $p\mathrm{CO_2}$sw data and $p\mathrm{CO_2}$sw diagnostics. In the South Pacific extra-tropics, where $CO_2$ variations in the surface and ocean interior are severely under-sampled, the difference in the air-sea $CO_2$ flux estimates between the diagnostic models and ocean interior $CO_2$ inversions is larger (0.18 $\mathrm{Pg\,C\,yr^{-1}}$). The range of estimates from forward OBGCMs is also large (−0.19 to −0.72 $\mathrm{Pg\,C\,yr^{-1}}$). Regarding inter-annual variability of air-sea $CO_2$ fluxes, positive and negative anomalies are evident in the tropical Pacific during the cold and warm events of the El Niño Southern Oscillation in the estimates from $p\mathrm{CO_2}$sw diagnostic models and from OBGCMs. They are consistent in phase with the Southern Oscillation Index, but the peak-to-peak amplitudes tend to be higher in OBGCMs ($0.40 \pm 0.09\,\mathrm{Pg\,C\,yr^{-1}}$) than in the diagnostic models ($0.27 \pm 0.07\,\mathrm{Pg\,C\,yr^{-1}}$).

## 1 Introduction

The Pacific Ocean plays an important role in the climate system as a large sink for anthropogenic carbon dioxide ($CO_2$), and, thereby partially mitigates the large-scale effects of human $CO_2$ emissions into the atmosphere. Estimates of the net air-sea $CO_2$ flux based on measurements of partial pressure of $CO_2$ in near-surface seawater ($p\mathrm{CO_2}$sw) and in the marine boundary air show that the extra-tropics in the North and South Pacific are major oceanic sinks of atmospheric $CO_2$. Although the $CO_2$ uptake in these sub-basins is counteracted in part by the large $CO_2$ outgassing from the tropical zone, the integrated $CO_2$ uptake by the Pacific Ocean likely accounts for approximately one third of the global oceanic $CO_2$ uptake (Takahashi et al., 2009a; Wanninkhof et al., 2013). In addition, it is well recognized that $CO_2$ outgassing from the tropical Pacific exhibits large variations with the El Niño Southern Oscillation (ENSO). This large inter-annual variability in air-sea $CO_2$ fluxes within the tropical Pacific is thought to play a dominant role in the inter-annual variability in the global oceanic $CO_2$ uptake (e.g., Feely et al., 1999, 2002, 2006; Ishii et al., 2004; Takahashi et al., 2009a; Wanninkhof et al., 2013).





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However, it is still difficult to quantify the net air-sea $CO_2$ fluxes from $p CO_2$sw measurements alone. This is primarily because measurements are sparse in both space and time in many parts of the ocean, particularly in the Southern Hemisphere, and because air-sea $CO_2$ fluxes are not themselves a directly measured but are derived and are associated with large uncertainties. Therefore, it is useful to compare the observational results with simulations from ocean models and estimates based on a combination of carbon data and models for the purpose of assessing fluxes over large temporal and spatial scales. Even then, there has been relatively poor agreement between the various approaches for estimating net air-sea $CO_2$ fluxes in the Pacific Ocean (McKinley et al., 2004; Peylin et al., 2005).

In this work, we begin with a review of what is known about air-sea $CO_2$ fluxes over the sub-basins of the Pacific Ocean. We then present a synthesis of state-of-the-art assessments of net air-sea $CO_2$ flux over the past two decades spanning the years from 1990 through 2009. This effort brings together $CO_2$ flux estimates from a wide range of available approaches: a synthesized climatological $p CO_2$sw data set, diagnostic models that use empirical interpolation schemes applied to the data of $p CO_2$sw, oceanic inversion methods from measurements of ocean interior dissolved inorganic carbon (DIC) and ocean circulation models, prognostic ocean general ocean circulation models coupled with biogeochemical models (OBGCMs), a data-assimilation model with $p CO_2$sw, and atmospheric $CO_2$ inversion systems with measurements of atmospheric $CO_2$ mixing ratio and atmospheric transport models. The goal of this paper is to perform a consistent analysis for these available methods and to arrive at consensus estimates of regionally integrated air-sea $CO_2$ fluxes for each of the sub-basins of the Pacific Ocean with corresponding estimates of the associated uncertainties. In evaluating the different estimates of $CO_2$ flux variability, it is important to devote particular attention to the differences in how air-sea $CO_2$ fluxes vary over different timescales. We begin with a consideration of the time-averaged fluxes over the period from 1990 through 2009. We then consider the inter-annual and seasonal variability for the same period. This allows us to assess whether community efforts are converging. Finally, we seek





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to identify the factors that cause the differences in the estimate of the flux among the methods, so that the results presented here can serve to guide future research.

## 2 Overview of air-sea $CO_2$ flux in the Pacific Ocean

### 2.1 Tropical Pacific

The physical and biogeochemical properties in the surface layer of the tropical Pacific show a large contrast between the domains of the western "warm pool" and the eastern "cold tongue" (Figs. 1 and 2). The warm pool is characterized by high sea surface temperatures (SST > 29.5 °C) and low sea surface salinities (SSS < 34.8) due to the large solar heat influx and high annual precipitation. As a result of the stratification thus attained, nitrate is depleted and the concentration of DIC is low (< 1950 µmol kg$^{-1}$ when salinity-normalized at $S = 35$) in the surface layer of this region. Due to the near equilibration of surface water $p\mathrm{CO_2}$ with atmospheric $CO_2$, and the presence of low wind speeds, net air-sea $CO_2$ fluxes over the "warm pool" are relatively small (< 1 mmol m$^{-2}$ day$^{-1}$; e.g., Ishii and Inoue, 1995).

By contrast, surface water in the eastern tropical Pacific cold tongue region tends to be highly supersaturated with respect to atmospheric $CO_2$. This is associated with the wind-driven equatorial divergence and turbulent mixing that brings colder, saline and nutrient- and $CO_2$-rich subsurface waters to the surface. The cold tongue is characterized by lower SSTs ($22 < T$(°C) $< 29$), higher SSSs (> 35), and higher DIC concentrations (> 1980 µmol kg$^{-1}$ at $S = 35$) than in the western Pacific warm pool (e.g., Ishii et al., 2004; see also Figs. 1 and 2). A significant portion of the DIC in the upwelled water is either removed by biological uptake or released to the atmosphere during the course of the poleward and westward advection. Nevertheless, $p\mathrm{CO_2}$sw remains higher than the atmosphere ($p\mathrm{CO_2}$sw − $p\mathrm{CO_2}$air > 90 µatm) due to the effect of concurrent warming (e.g., Feely et al., 1999, 2002, 2006; Ishii et al., 2004; Takahashi et al., 2009a).





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The "cold tongue" in the eastern tropics extends to the west during the cold events of ENSO (La Niña) and retreats to the east during the warm events of ENSO (El Niño). ENSO drives changes in the distributions of DIC, SST, and salinity in surface water as well as the surface wind field, and causes large perturbations to $p\mathrm{CO_2}$sw and significant temporal variability in the $CO_2$ outgassing from the tropical Pacific (e.g., Feely et al., 1987, 2002, 2006; Inoue and Sugimura, 1992; Ishii et al., 2004). The ENSO-driven changes to the variables that control $p\mathrm{CO_2}$sw and the gas transfer coefficient have been simulated and analyzed in a modeling study of Doney et al. (2009a, b). Their analysis revealed that the largest variability in air-sea $CO_2$ flux in the equatorial Pacific occurs in the region spanning the Date Line to the coast of Peru (Fig. 3). The dominant driver of this variability is the variability in DIC (Fig. 4). Although it is partly offset by the counteracting effect of variability in SST, the effect of DIC-driven changes in $p\mathrm{CO_2}$sw is reinforced by the effect of wind-speed change and results in the large variability in the air-sea $CO_2$ flux (see Figs. 3 and 4). A number of studies with OBGCMs have examined biogeochemical processes and air-sea $CO_2$ fluxes over the tropical Pacific (e.g., Winguth et al., 1994; Le Quéré et al., 2000; Obata and Kitamura, 2003; McKinley et al., 2004, 2006; Wang et al., 2006; Christian et al., 2008; Doney et al., 2009a). These studies have shown the dominant role of the tropical Pacific in the global inter-annual variability in the oceanic $CO_2$ uptake.

Underlying the large interannual variability is a secular trend with increasing $p\mathrm{CO_2}$sw observed in this region over the past decades (Feely et al., 1999, 2006; Takahashi et al., 2003). The mean rate of $p\mathrm{CO_2}$sw increase is consistent with the rate of atmospheric $CO_2$ increase, but decadal modulations have also been reported (Takahashi et al., 2003; Feely et al., 2006; Ishii et al., 2009). The decadal variability of $p\mathrm{CO_2}$sw is possibly linked with changes in the shallow meridional overturning circulation (McPhaden and Zhang, 2002, 2004), but a mechanistic understanding of this connection is still in development.





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## 2.2 North Pacific extra-tropics

In the extra-tropics, the dominant timescale of variability is the seasonal cycle. The predominance of this signal is expressed not only in SST, but also in large seasonal variations of mixed layer depth. Such seasonal variations in physical state variables are then associated with important seasonal variability in ocean biogeochemistry and biological activity. The factors drive changes in DIC and $p\mathrm{CO_2}$sw. In the North Pacific, the seasonality of $p\mathrm{CO_2}$sw is particularly significant in the vicinity of the Kuroshio Extension Current and in the western subarctic zone including the western subarctic gyre and the Bering Sea (Takahashi et al., 2002) (Figs. 1 and 2). Throughout the majority of the North Pacific extra-tropics, particularly in the northern subtropical zone, cooling in winter is the dominant control on low $p\mathrm{CO_2}$sw although it is partly compensated for by increases in DIC associated with wintertime vertical mixing (e.g., Inoue et al., 1987; Takahashi et al., 1993; Ishii et al., 2001; Keeling et al., 2004). By contrast, seasonal variations in $p\mathrm{CO_2}$sw in the western subarctic zone are dominated by the seasonal variations of DIC associated with the enhanced convection in winter and the large net biological DIC consumption in summer (Takahashi et al., 1993; Tsurushima et al., 2002; Chierici et al., 2006).

At interannual to decadal timescales, the dominant mode of basin-scale variability is the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997). Positive PDO anomalies are associated with positive SST anomalies in the Alaskan Gyre and along the west coast of North America, and negative SST anomalies in the central and western North Pacific. While the PDO is expected to impact the distribution of DIC in the upper layers of the North Pacific, the integrated effect of PDO on air-sea $\mathrm{CO_2}$ fluxes remains poorly quantified. Drawing on output from a collection of OBGCMs, McKinley et al. (2006) argued for a correlation of air-sea $\mathrm{CO_2}$ fluxes in the North Pacific with the PDO. Extrapolating from the mechanistic interpretation presented by McKinley et al. (2006), one can posit the following paradigm for the amplitude of inter-annual variability in the extra-tropics of the North Pacific, in particular around the subtropical – subarctic frontal





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zone where the ocean is a strong $CO_2$ sink in winter: inter-annual variations in wintertime $p$$CO_2$sw are rather small, despite sizable interannual variability in SST, because the opposing effects of SST and DIC concentrations on $p$$CO_2$sw compensate each other. This paradigm is consistent with results from the repeated $p$$CO_2$sw measurements in the northern subtropics of the western North Pacific at 137° E (Midorikawa et al., 2006). This study demonstrated that the inter-annual variations in SST and DIC have a counteracting effect on $p$$CO_2$sw, and consequently the inter-annual variability in air-sea $CO_2$ flux is thought to be associated with the variability in the wind speed. The modeling study of Doney et al. (2009a) came to the same conclusion (Fig. 3). By contrast, larger amplitude interannual variability in $p$$CO_2$sw and air-sea $CO_2$ flux in the subarctic zone and in the eastern subtropics are driven primarily by variability in DIC.

Long-term trends towards increasing $p$$CO_2$sw have been observed since the early 1980s along a north-south time-series line to the south of Japan at 137° E (Inoue et al., 1995; Midorikawa et al., 2005, 2012) and at a time-series station near Hawaii (Keeling et al., 2004; Dore et al., 2009). The principal publications to date for basin-scale long-term trends in $p$$CO_2$sw are those of Takahashi et al. (2003, 2006) and Lenton et al. (2012), which used existing $p$$CO_2$ measurements spanning 1970–2004 and from the mid-1990s to the mid-2000s, respectively. These observations show that the mean rate of $p$$CO_2$sw increase is roughly consistent with the rate of atmospheric $CO_2$ increase, but it is variable both in space and time. Long-term time-series records of oceanic $CO_2$ appears to show a decrease in the positive trends in $p$$CO_2$sw and DIC in the eastern to southern rim of the subtropical cell and in its tropical branch after the strong warm event of ENSO in 1997–1998 (Dore et al., 2003, 2009; Keeling et al., 2004; Ishii et al., 2009; Midorikawa et al., 2012). A change in the subtropical cell (Qiu and Chen, 2010) is a likely driver, but the mechanism driving a decrease in the rate of increasing $p$$CO_2$sw and its possible linkage to PDO is not fully understood.

Regarding a potential change in the seasonal cycle in $p$$CO_2$sw, Rodgers et al. (2008) and Gorgues et al. (2010) argued that over decadal timescales in the North Pacific there is a divergence between the trend in winter and summer that occurs in the ab-





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sence of trends in the circulation state of the ocean, and consequently a decadal trend arises towards an increased seasonal cycle. These results found further support in the study of Nakano et al. (2011), who attributed this to the interaction between seasonal dynamics and the changes in carbonate chemistry in seawater with increasing $CO_2$. This underscores the importance of accounting for the full seasonal cycle when calculating long-term trends, as trends inferred from summer-biased measurements will introduce bias in trend estimates (Lenton et al., 2012).

### 2.3 South Pacific extra-tropics

As is the case for the extra-tropical North Pacific, the extra-tropical South Pacific is also a major sink for atmospheric $CO_2$. However, this region poses particular challenges to estimating air-sea $CO_2$ fluxes because of the paucity of $p$CO$_2$sw measurements over this vast sub-basin. The various gridded data products that have resulted from data synthesis activities of $p$CO$_2$sw have by necessity relied on interpolation over large spatial scales and for seasonality in this region. For this region, we may expect that the flux estimates that rely heavily on the $p$CO$_2$sw measurements (namely diagnostic approaches, $p$CO$_2$sw assimilation, as well as atmospheric $CO_2$ inversion studies) to exhibit relatively similar mean fluxes, and for estimates that rely either on forward ocean models or ocean $CO_2$ inversions (which rely more heavily on interior carbon measurements) to produce different time-mean fluxes.

## 3 Methods

We use a range of air-sea $CO_2$ flux products in the Pacific Ocean, with these products described below. They are mainly the products collected for the Regional Carbon Cycle Assessment and Processes (RECCAP) (Canadell et al., 2011), but they also include products that have been collected in the preparation of this study for the Pacific Ocean synthesis.





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## 3.1 Climatological $pCO_2$sw data and $pCO_2$sw diagnostic models

The evaluation of the air-sea $CO_2$ flux through gridded $pCO_2$sw data products was originally developed in the studies built on the database of T. Takahashi for shipboard $pCO_2$sw measurements (Tans et al., 1990; Takahashi et al., 1997). The database and the gridded data products have been repeatedly updated and widely used since. The analysis here draws upon the data set of gridded monthly climatological $pCO_2$sw in the reference year 2000 (Takahashi et al., 2009a) (hereafter referred to LDEO V2009; the climatological mean values for $4^\circ \times 5^\circ$ pixel areas are listed in http://www.ldeo.columbia.edu/res/pi/CO2). Diagnostic models of $pCO_2$sw proposed by Park et al. (2010) and Sugimoto et al. (2012) are also used to estimate seasonal and interannual variations in air-sea $CO_2$ fluxes. In Park et al. (2010), sub-annual $pCO_2$sw-SST relationships have been empirically-derived for each $4^\circ \times 5^\circ$ pixel of climatological monthly mean $pCO_2$sw of LDEO V2009 for the extra-tropics. For tropical Pacific, they used empirical $pCO_2$sw-SST equations that are updated from those of Feely et al. (2006) and are unique for three different time periods of 1979 through 1989, 1990 through mid-1998 and mid-1998 through 2008. Sugimoto et al. (2012) derived the relationships of $pCO_2$sw versus sea surface salinity, ocean color, year of observation as well as SST for each of 9 sub-regions of the Pacific Ocean from individual $pCO_2$sw data in Takahashi et al. (2008). These relationships were combined with satellite-derived fields of SST and other parameters to obtain the monthly fields of $pCO_2$sw. Results from another empirical technique using neural network (Nakaoka et al., 2013) that has been tested for the Atlantic Ocean (Telszewski et al., 2009) are also included for comparisons in the North Pacific extra-tropics (Table 1).

A gas transfer velocity ($k$) is commonly applied to climatological $pCO_2$sw data and diagnostic models to estimate the air-sea $CO_2$ flux employing the following functional form (Sweeney et al., 2007; Park et al., 2012):

$$k = 0.25(\mathrm{Sc}/660)^{-0.5}\langle U^2\rangle \qquad (1)$$





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where Sc is the Schmidt number of $CO_2$ at SST (Wanninkhof, 1992) and $\langle U^2 \rangle$ is monthly mean second moment of wind speed at 10 m height derived from 6 hourly winds at 25 km resolution of the Cross-Calibrated, Multi-Platform (CCMP) Ocean Surface Wind Product (http://podaac.jpl.nasa.gov/DATA_CATALOG/ccmpinfo.html) (Ardizzone et al., 2009; Atlas et al., 2011). The coefficient 0.25 is specific to the wind-product used to calculate the air-sea $CO_2$ flux. It has been optimized globally so that the change in the bomb-$^{14}$C inventory in the ocean matches atmospheric $^{14}$C invasion rate.

The $CO_2$ flux ($F$) is then calculated by the conventional equation for the bulk method:

$$\begin{aligned} F &= k \cdot (C_{sw} - K_0 \cdot pCO_2\text{air}) \\ &= k \cdot K_0 \cdot (pCO_2\text{sw} - pCO_2\text{air}) = k \cdot K_0 \cdot \Delta pCO_2 \end{aligned} \tag{2}$$

where $C_{sw}$ denotes the concentration of $CO_2$ in surface seawater and $K_0$ denotes $CO_2$ solubility in seawater at a given temperature and salinity. Following the widely used convention for $pCO_2$ climatologies and diagnostic models, this flux is positive when $CO_2$ is released from ocean to the atmosphere and is negative when it is absorbed into the ocean.

### 3.2 Ocean interior $CO_2$ inversion methods

Ocean interior $CO_2$ inversion methods use a Green function inverse method to infer regional air-sea $CO_2$ fluxes from ocean interior DIC observations and ocean general circulation models. This method was first presented by Gloor et al. (2003), and the results shown in this paper are the values of the contemporary flux, i.e., the sum of natural and anthropogenic fluxes presented in Gruber et al. (2009) for the RECCAP period 1990–2009. The inversion was originally done for 30 ocean regions, and then aggregated to 23 regions (10 regions in the Pacific) as described in Mikaloff Fletcher et al. (2006, 2007). These results are generally in good agreement with more recent ocean inverse estimates (Gerber et al., 2009; Gerber and Joos, 2010). The contemporary fluxes include riverine components of $CO_2$ flux, i.e., +0.08 Pg C yr$^{-1}$ in the North





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Pacific extra-tropics and +0.04 $\mathrm{Pg\,C\,yr^{-1}}$ in the tropical Pacific (Jacobson et al., 2007). The anthropogenic fluxes have been scaled to a 1990–2009 average. A skill score has been determined for each model to account for the substantial differences in the model's ability to correctly simulate the oceanic distribution of passive tracers, and the skill scores are used to calculate the weighted mean net air-sea $CO_2$ flux (Gruber et al., 2009). Since this method does not resolve seasonal and interannual variability, the results are only used to compare the 1990–2009 mean air-sea $CO_2$ fluxes.

### 3.3 Ocean biogeochemistry/general circulation models (OBGCMs)

This work also incorporates results from several prognostic ocean biogeochemistry/general circulation model simulations over the period of interest (Table 2). From a total of nine modeling results, seven were retrieved from the RECCAP website (http://www.globalcarbonproject.org/reccap/products.htm). These simulations include not only an account of seasonally- and inter-annually-varying air-sea fluxes of $CO_2$, but also prognostic representations of the processes that are deemed to be important in controlling trends and variations in the ocean carbon cycle. For each case, a prognostic biogeochemistry model is embedded in a physical ocean circulation model and run online. The surface forcing for the dynamical models consists of using atmospheric flux fields derived from a combination of reanalysis and remotely sensed products. Surface buoyancy forcing is accomplished through the use of bulk formulas or other methods for heat and freshwater fluxes, with a restoring of SSS towards climatological values being characteristic of most of the models. The models considered here are coarse resolution models that are neither eddy permitting nor eddy resolving.

Given that the models tend to have differences in their respective (i) underlying physical models, (ii) underlying biogeochemical models, (iii) surface forcing fields, and (iv) handling of river carbon discharge, they should be expected to produce different representations of the ocean carbon cycle. At this point in time, our primary objective will be to provide a description of their similarities and differences. The sensitivity of the





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modeled carbon cycle to each of these four differences will not be given any extensive consideration in this study. However, we will be providing at least a preliminary assessment of the sensitivity of the differences in the model results, particularly with respect to the sensitivity of the carbon cycle to physical forcing at the sea surface in Sect. 6.1.

### 3.4 Ocean $p\mathrm{CO_2}$sw data assimilation

The $p\mathrm{CO_2}$sw data set of LDEO V2009 has also been assimilated into an offline tracer transport model (OTTM; Valsala and Maksyutov, 2010). This assimilation system minimizes the model biases in the surface ocean $p\mathrm{CO_2}$ through a weak constraint given to its gridded monthly climatology of LDEO V2009, while a strong constraint is given to the in-situ ship-observed $p\mathrm{CO_2}$sw measurements whenever they are available in the LDEO database (Takahashi et al., 2009b). The weak constraint is further weighted by the inverse of the model interannual variance, which ensures that the model is constrained to the monthly climatological $p\mathrm{CO_2}$sw only in regions where the interannual variability is small. Assimilated data of $p\mathrm{CO_2}$sw and air-sea $\mathrm{CO_2}$ flux were constructed from 1996–2008 using this method, while here we present an extended record of the data starting from 1990. Prior to 1996, the data represent the model interannual variability summed to the monthly climatology derived from the assimilation period of 1996–2008.

### 3.5 Atmospheric $\mathrm{CO_2}$ inversion methods

An atmospheric $\mathrm{CO_2}$ inversion intercomparison project community was launched by the TransCom with the RECCAP initiative collecting a number of atmospheric $\mathrm{CO_2}$ inversion results and comparing those to synthesize general features of the recent state-of-the-art inversions. The results are archived at the web site https://transcom.lsce.ipsl.fr/ representing 14 different approaches. Peylin et al. (2013) selected 11 inversion results from those and presented long term mean, long term trend, interannual variations and mean seasonal variations separately for land and ocean regions in the tropics and northern and southern extra-tropics.

The analysis here includes the inversion results for the Pacific Ocean region from the total of six atmospheric $CO_2$ inversions with outputs longer than 17 yr for decadal mean flux and ten models for mean seasonal variations (Table 3). It should be noted here that they differ in the atmospheric $CO_2$ datasets (i.e., observational constraints), atmospheric transport models, spatial resolution of the optimized flux and inversion methods. Most of the inversions used climatological air-sea $CO_2$ flux data from some versions of the LDEO monthly climatology as a prior air-sea $CO_2$ flux estimate, and therefore regionally-integrated or seasonal variations of posterior net air-sea $CO_2$ fluxes have been constrained by it to a greater or lesser extent depending on the inversion method.

## 4 Regions of assessment

We provide regionally-integrated net air-sea $CO_2$ fluxes over three sub-basins of the Pacific Ocean that are zonally partitioned. They are the zone to the north of 18° N including the Bering Sea (< 66° N), the tropical zone bounded by 18° N and 18° S, and the southern zone bounded by 18° S and 44.5° S (Fig. 5). The region to the south of 44.5° S is discussed in Lenton et al. (2013). The boundaries separating these three sub-basins are chosen to be consistent with previous publications, grouping 10 prescribed ocean regions of the ocean $CO_2$ inversions in the Pacific (Mikaloff Fletcher et al., 2006, 2007; Gruber et al., 2009). As such, these divisions are fairly consistent with dynamical boundaries separating the subtropical gyres of the North Pacific and the South Pacific from the Tropical Pacific.

The tropical zone defined here (18° N–18° S) includes a small part of the equatorward flanks of the subtropical gyres of the North and South Pacific. Nevertheless the tropical zone mainly consists of the "warm pool" in the west and the "cold tongue" in the east. The "cold tongue" includes both the equatorial and the Peruvian divergence systems. In Sect. 5.1, the inter-annual variability in the western tropical zone to the west of 160° W and the eastern tropical zone to the east of 160° W are presented separately in order





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to see the effect of westward expansions of the "cold tongue" during the ENSO cold events.

The North Pacific to the north of 18° N encompasses most of the subtropical gyre and the entire subarctic zone. As shown in Figs. 1 and 2, seasonality of $p\mathrm{CO_2}$sw reverses within this domain. In winter, $p\mathrm{CO_2}$sw decreases to considerable $CO_2$ undersaturation with respect to atmospheric $CO_2$ in the northern subtropics due to the large effect of seasonal cooling, and $p\mathrm{CO_2}$sw increases to the point of supersaturation in the subarctic region due to the large effect of DIC increase associated with vertical convection. In contrast, during the summer, $p\mathrm{CO_2}$sw increases to being in near equilibrium with respect to the atmosphere in the subtropics due to seasonal warming and $p\mathrm{CO_2}$sw decreases in the subarctic region due to the DIC decrease associated with biological production. The air-sea $CO_2$ flux shown for the North Pacific extra-tropics is the integrated flux over these two sub-domains.

The extra-tropical zone between 18° S and 44.5° S covers most of the subtropical gyre in the South Pacific. We note that the southern boundary at 44.5° S lies in the vicinity of the Subtropical Convergence Zone and the surface water in this zone is highly undersaturated with respect to atmospheric $CO_2$ (Metzl et al., 1999; Inoue, 2000; Takahashi et al., 2009a). Therefore, the estimates of air-sea $CO_2$ flux in the South Pacific extra-tropics is expected to depend largely on the choice of the southern boundary. We will examine this using LDEO V2009 climatological fluxes in Sect. 5.3.

For each sub-basin, regionally-integrated air-sea $CO_2$ fluxes were calculated as monthly means from each product: monthly climatological $p\mathrm{CO_2}$sw data of LDEO V2009, $p\mathrm{CO_2}$sw diagnostic models, Ocean BGC models, a $p\mathrm{CO_2}$sw data-assimilation, and atmospheric $CO_2$ inversions. Decadal and longer-term mean fluxes were then calculated over the intervals 1990–1999 and 2000–2009, as well as over the combined period 1990–2009. They were subsequently compiled for each of the approaches: the average ± range for $p\mathrm{CO_2}$sw diagnostic models and the median ± median absolute deviation (MAD) for OBGCMs and atmospheric $CO_2$ inversions were calculated for decadal and longer-term means. In addition, skill-weighted mean values have been given for





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ocean $CO_2$ inversions. However, no remarkable differences are seen in decadal mean flux between 1990–1999 and 2000–2009 in all approaches (Fig. A1). Therefore, in the following sections, we will not mention the decadal means but present only longer-term means for the period 1990–2009 and inter-annual variability and mean seasonal variability for this period. For the $p\mathrm{CO_2}$sw data assimilation and the atmospheric $CO_2$ inversions, mean fluxes were calculated for 1990–2008 since modeling products in the year 2009 were not available. In the cases of $p\mathrm{CO_2}$sw diagnostic models and OBGCMs, differences in the mean fluxes between for 1990–2009 and for 1990–2008 were minimal ($< 0.01\,\mathrm{Pg\,C\,yr^{-1}}$). Therefore we assume that the comparison of mean and median fluxes for 1990–2009 of diagnostic models and OBGCMs and those for 1990–2008 of $p\mathrm{CO_2}$sw data assimilation and atmospheric $CO_2$ inversions are not problematic in the following discussions.

## 5 Results

### 5.1 Tropical Pacific 18° S–18° N

From the LDEO V2009 climatological $p\mathrm{CO_2}$sw data set (which has been filtered to remove the ENSO warm events) and CCMP monthly wind speed, the annual $CO_2$ flux from the tropical Pacific is estimated to be $+0.51 \pm 0.24\,\mathrm{Pg\,C\,yr^{-1}}$ in year 2000 (Table 4; Fig. 6), which comprises a small efflux ($+0.06\,\mathrm{Pg\,C\,yr^{-1}}$) from the western tropical sector (west of 160° W), and a larger efflux ($+0.45\,\mathrm{Pg\,C\,yr^{-1}}$) from the eastern tropical sector. The mean of the time-averaged air-sea $CO_2$ flux from the two diagnostic models (which include the ENSO warm events) considered here ($+0.52 \pm 0.09\,\mathrm{Pg\,C\,yr^{-1}}$) is consistent with the estimates from the climatological $p\mathrm{CO_2}$sw and the atmospheric $CO_2$ inversions ($+0.53 \pm 0.08\,\mathrm{Pg\,C\,yr^{-1}}$). These diagnostic models also showed a peak-to-peak amplitude of inter-annual variability of $0.27 \pm 0.07\,\mathrm{Pg\,C\,yr^{-1}}$ between 1990 and 2009 (Table 4; Fig. 7). Negative anomalies of air-sea $CO_2$ flux are associated with the ENSO warm events. The negative anomaly is particularly significant in the eastern trop-





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ical sector during the ENSO strong warm event that occurred in 1997–1998 when the negative anomaly of the flux reached a level of −0.29 and −0.14 $\text{Pg C yr}^{-1}$, depending on the diagnostic model. This strong warm event was immediately followed by the persistent ENSO cold event in 1998–2000 with quite large positive anomalies in the western sector (up to +0.11 and +0.17 $\text{Pg C yr}^{-1}$) as well as in the eastern sector (+0.08 and +0.09 $\text{Pg C yr}^{-1}$) (Fig. 7). The results of $p\text{CO}_2\text{sw}$ data assimilation also shows the inter-annual variability of air-sea $CO_2$ flux that is associated with the ENSO, although the negative anomaly during the 1997–1998 warm event was smaller (−0.14 $\text{Pg C yr}^{-1}$).

The median ± MAD of time-averaged air-sea $CO_2$ flux in the tropical Pacific evaluated by OBGCMs is +0.39 ± 0.04 $\text{Pg C yr}^{-1}$ for 1990–2009. This is in good agreement with the estimate from ocean interior $CO_2$ inversion methods (+0.37 ± 0.12 $\text{Pg C yr}^{-1}$), but is 0.13 $\text{Pg C yr}^{-1}$ smaller than the estimate from the two diagnostic models considered here. Given the considerable range of flux estimates in both diagnostic models (+0.44 to +0.61 $\text{Pg C yr}^{-1}$) and the OBGCMs (+0.25 to +0.55 $\text{Pg C yr}^{-1}$), it is important to compare the $p\text{CO}_2\text{sw}$ fields and gas exchange coefficients to provide more clarity about these differences. One of the potential sources of the differences is the choice of wind-speed product used to force the ocean circulation in OBGCMs as well as to compute the $CO_2$ gas transfer velocities that are used both in diagnostic models and OBGCMs. Park et al. (2010) have shown that the global mean air-sea $CO_2$ flux changes by as much as 20 % with the choice of wind-speed products and coefficient for gas transfer velocity for gas exchange in their diagnostic model. We will discuss this in more detail in Sect. 6.1 for a diagnostic model and for an OBGCM.

In terms of the phase of inter-annual variability, the results from most OBGCMs are consistent with those from diagnostic models demonstrating larger $CO_2$ efflux during the ENSO cold events and smaller efflux during the warm events. However, OBGCMs appear more sensitive to the ENSO warm and cold events (Table 4 and Fig. 7), particularly during the 1995–1996 cold event and during the 1997–1998 warm event. The reason for the larger ENSO sensitivity in OBGCMs than diagnostic models is yet to be determined but is likely to be attributable to the larger response of the $p\text{CO}_2\text{sw}$ field



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to these ENSO events. Diagnostic models may more or less smooth out the variability through the regression analyses of $p\mathrm{CO_2}$sw as a function of SST and other parameters that are used to correct the implicated under-sampling in observations. To date, two modeling studies have evaluated the skill of the diagnostic method originally developed by Lee et al. (1998) in simulating air-sea $CO_2$ flux variability over the tropical Pacific. For both studies, skill was evaluated through an Observing System Simulation Experiment (OSSE), whereby the modeled ocean is sampled at the same space-time coordinates as the real-world observations, the same empirical relationships are employed to estimate air-sea $CO_2$ fluxes, and this estimate is compared with the fully resolved explicit fluxes in the model. The study of Park et al. (2010) found that their diagnostic model underestimates 15–20 % of the variability, while it overestimates 25–30 % for the full fluxes, in the tropical Pacific. The study of Christian et al. (2008) also argued that the variability in the flux are significantly underestimated by the diagnostic model for the tropical Pacific, with estimated variability being of order 50 % the amplitude of the full variability.

The estimates of air-sea $CO_2$ flux in the tropical Pacific from the atmospheric $CO_2$ inversions used in this study showed large monthly fluctuations that are not seen in the estimates from diagnostic models and OBGCMs (Fig. 7). Inverse estimates for tropical regions are subject to a high degree of uncertainty due to the limited number of observing stations in this region. Nevertheless, many of the atmospheric $CO_2$ inverse models show a decrease in outgassing during the strong ENSO warm event in 1997–1998 and an increase during the persistent cold event over 1998–2000.

### 5.2 North Pacific extra-tropics 18° N–66° N

The mean annual air-sea $CO_2$ flux in the North Pacific extra-tropics north of 18° N is estimated to be $-0.44 \pm 0.21\,\mathrm{Pg\,C\,yr^{-1}}$ in year 2000 from the LDEO V2009 climatological $p\mathrm{CO_2}$sw fields, and $-0.47$ and $-0.57\,\mathrm{Pg\,C\,yr^{-1}}$ for the period 1990–2009 from diagnostic models of Park et al. (2010) and Sugimoto et al. (2012), respectively. An independent diagnostic model of Nakaoka et al. (2013) that uses a non-linear empirical neural network technique also suggests similar, but slightly smaller, mean influx ($-0.40\,\mathrm{Pg\,C\,yr^{-1}}$) into this domain for the period 2002–2008. This is 0.07 and $0.20\,\mathrm{Pg\,C\,yr^{-1}}$ smaller, respectively, than the $CO_2$ flux estimates from the two other diagnostic models for the same period. Since the same wind product has been used to calculate gas exchange coefficient, these differences in the flux estimate are attributable to the differences in the $p\mathrm{CO_2sw}$ field. The strong $CO_2$ uptake in the North Pacific is dominated by the uptake in the northern subtropics and subtropical-to-subarctic transition zone in winter (Fig. 2), where the effect of cooling on $p\mathrm{CO_2sw}$ is stronger than the effect of DIC increase due to vertical mixing (Ishii et al., 2001; Takahashi et al., 2002). The mean annual air-sea $CO_2$ flux in the same sub-basin from $p\mathrm{CO_2sw}$ data assimilation ($-0.37\,\mathrm{Pg\,C\,yr^{-1}}$) is somewhat smaller, but those from ocean interior $CO_2$ inversions ($-0.42\pm0.08\,\mathrm{Pg\,C\,yr^{-1}}$) and atmospheric $CO_2$ inversions ($-0.48\pm0.08\,\mathrm{Pg\,C\,yr^{-1}}$) are consistent with the range of estimates from LDEO V2009 climatological $p\mathrm{CO_2sw}$ fields and diagnostic models. The net $CO_2$ sink estimated by the OBGCMs ($-0.57\pm0.02\,\mathrm{Pg\,C\,yr^{-1}}$) is the strongest among the estimates from the various approaches.

The peak-to-peak difference of the inter-annual variability in the air-sea $CO_2$ fluxes derived from diagnostic models over the North Pacific extra-tropics is small ($0.12\,\mathrm{Pg\,C\,yr^{-1}}$) (Fig. 8 and Table 4). This is also the case for the OBGCMs. Most of these models show slightly positive anomalies ($\sim 0.1\,\mathrm{Pg\,C\,yr^{-1}}$) for the period of 1999–2001 when the PDO index tended to be negative, but the relationship between the anomaly of $CO_2$ flux and the PDO is not discernible for other periods. The amplitude of inter-annual variability is somewhat larger in the flux estimate from $p\mathrm{CO_2sw}$ data assimilation ($0.19\,\mathrm{Pg\,C\,yr^{-1}}$). Valsala et al. (2012) posited a "four-region" structure in the North Pacific air-sea $CO_2$ flux variability related to the PDO in the data assimilation product. However, the correlation of the air-sea $CO_2$ flux with PDO is not clear when averaged over the North Pacific extra-tropics. Most of the atmospheric $CO_2$ inversions show negative anomalies around the late 1994, 1997, 2001 and early 2007 and positive anomalies around early 1994 and 2004. However, the correlations with the PDO index are also not obvious. No consistent pattern of inter-annual variability of air-sea $CO_2$ flux is seen among the different approaches in the North Pacific extra-tropics suggesting that the various methods have not yet converged in their representation of interannual-to-decadal variability over the North Pacific extra-tropics.

With regard to the seasonality of air-sea $CO_2$ flux in the North Pacific extra-tropics, results from the three diagnostic models are consistent in that they all show very small net air-sea $CO_2$ flux in summer (July–September: $+0.03 \pm 0.10\,\mathrm{Pg\,C\,yr^{-1}}$) and a larger influx into the ocean in winter (January–March: $-0.86 \pm 0.20\,\mathrm{Pg\,C\,yr^{-1}}$) (Fig. 9). The difference in the net annual air-sea $CO_2$ flux among these diagnostic models is mainly attributable to the difference in the flux estimates in the cold time of year (December–April). The phase of seasonality in the $p\mathrm{CO_2sw}$ data assimilation product is consistent with diagnostic models but does shows a net $CO_2$ efflux in summer ($+0.56\,\mathrm{Pg\,C\,yr^{-1}}$ in July). All OBGCMs presented in this work also show well-defined seasonality with large $CO_2$ sink in winter ($-1.8$ to $-0.9\,\mathrm{Pg\,C\,yr^{-1}}$) and slightly negative or moderately positive flux in summer ($-0.1$ to $+0.6\,\mathrm{Pg\,C\,yr^{-1}}$). By contrast, in the atmospheric $CO_2$ inversions large sub-annual variations are found in the air-sea $CO_2$ flux but its seasonality remains poorly resolved

### 5.3 South Pacific extra-tropics 44.5° S–18° S

The climatological net annual air-sea $CO_2$ flux at the year 2000 evaluated from LDEO V2009 climatological $p\mathrm{CO_2sw}$ was $-0.29 \pm 0.14\,\mathrm{Pg\,C\,yr^{-1}}$, and the long-term mean net annual air-sea $CO_2$ flux over two decades after 1990 was $-0.28 \pm 0.00\,\mathrm{Pg\,C\,yr^{-1}}$ from the diagnostic models and $-0.29 \pm 0.08\,\mathrm{Pg\,C\,yr^{-1}}$ from the atmospheric $CO_2$ inversions, respectively. The strength of the $CO_2$ sink in the South Pacific extra-tropics is smaller than that in the North Pacific extra-tropics ($-0.44$ to $-0.52\,\mathrm{Pg\,C\,yr^{-1}}$) estimated by the same approaches. The smaller sink in the South Pacific extra-tropics is, in part, ascribed to its smaller area defined here (18° S–44.5° S; $3.88 \times 10^7\,\mathrm{km^2}$) than in the North Pacific extra-tropics (18° N–66° N; $4.43 \times 10^7\,\mathrm{km^2}$), but it is primarily attributable to the difference in the zonal distributions of air-sea $CO_2$ flux. In the North





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Pacific extra-tropics, there is a band serving as a strong $CO_2$ sink that extends over the mid-latitudes from the region off of Japan to off of the west coast of North America (Figs. 2 and 5). The $CO_2$ sink is particularly strong around the subtropical-to-subarctic transition zone where the net annual air-sea $CO_2$ flux from climatological $p$CO$_2$sw of LDEO V2009 reaches $-2.9\,\mathrm{mol\,m^{-2}\,yr^{-1}}$. The western South Pacific extra-tropics near Australia and New Zealand is also a region of $CO_2$ sink but its strength is moderate (ca. $-2.1\,\mathrm{mol\,m^{-2}\,yr^{-1}}$). In addition, the eastern South Pacific extra-tropics is a weak sink or even a weak source of $CO_2$ to the atmosphere. However, it has to be noted that the South Pacific extra-tropics is severely undersampled for $p$CO$_2$sw in winter (Takahashi et al., 2009a) and the uncertainty in the air-sea $CO_2$ flux is thereby considerably larger there than it is over the North Pacific. The integrated flux over this region is also sensitive to the choice of the southern boundary of the South Pacific extra-tropics. In this analysis, the Southern Pacific extra-tropics has been defined to be the region north of 44.5° S to 18° S. According to the LDEO V2009 climatological $p$CO$_2$sw, the net annual $CO_2$ flux integrated over the zonal band between 42° S and 46° S in the South Pacific is $-0.056\,\mathrm{Pg\,C\,yr^{-1}}$. This accounts for *ca.* 20 % of net $CO_2$ sink in the South Pacific extra-tropics.

The results from OBGCMs for the South Pacific extra-tropics show a large range of estimates ($-0.19$ to $-0.71\,\mathrm{Pg\,C\,yr^{-1}}$) for the net air-sea $CO_2$ flux averaged over the period 1990–2009. In addition, their median ($-0.39 \pm 0.11\,\mathrm{Pg\,C\,yr^{-1}}$) indicate a $0.1\,\mathrm{Pg\,C\,yr^{-1}}$ larger sink than the estimates from other methods that are constrained by the data of $p$CO$_2$sw. The larger range of air-sea $CO_2$ fluxes here presents a marked contrast to the rather small range of OBGCM estimates in the North Pacific extra-tropics ($-0.57 \pm 0.02\,\mathrm{Pg\,C\,yr^{-1}}$). In the South Pacific extra-tropics, the estimates of air-sea $CO_2$ flux by OBGCMs split up into two groups. For the two models that show a larger $CO_2$ sink, the averaged (± standard deviation) flux is $-0.71 \pm 0.02\,\mathrm{Pg\,C\,yr^{-1}}$, while for the remaining six models a smaller $CO_2$ sink of $-0.32 \pm 0.08\,\mathrm{Pg\,C\,yr^{-1}}$ is seen, which is more consistent with LDEO V2009 climatology and diagnostic models.





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The weighted average of net air-sea $CO_2$ flux from the ocean $CO_2$ inversions is $-0.46 \pm 0.10\,\mathrm{Pg\,C\,yr^{-1}}$. This shows nearly a $0.2\,\mathrm{Pg\,C\,yr^{-1}}$ larger sink than the estimates from climatological $p\mathrm{CO_2sw}$ and diagnostic models. Only the results from ocean interior $CO_2$ inversions show a larger oceanic $CO_2$ sink in the South Pacific extra-tropics than in the North Pacific extra-tropics.

With regard to the inter-annual variations, no remarkable change is seen in the diagnostic models (Fig. 8 and Table 4). The estimate from Sugimoto et al. (2012) shows small positive anomalies for 1995–1997 and small negative anomalies for 2006–2008, but they are within $\pm 0.1\,\mathrm{Pg\,C\,yr^{-1}}$. Two of the OBGCMs, i.e., "BER" and "UEA_NCEP1" (Table 2), show a larger $CO_2$ sink in the South Pacific extra-tropics and also show larger interannual variation with the amplitude of $\pm 0.23$ and $\pm 0.38\,\mathrm{Pg\,C\,yr^{-1}}$, but the variations are smaller ($< 0.13\,\mathrm{Pg\,C\,yr^{-1}}$) in the other six OBGCMs that show smaller net $CO_2$ sinks. The latter six OBGCMs consistently show negative anomalies averaging about $-0.1\,\mathrm{Pg\,C\,yr^{-1}}$ in 1998. However, no such anomaly is seen in 1998 in the diagnostic models. In general, the atmospheric $CO_2$ inversions again show much larger inter-annual variations than the other approaches. This is likely to be attributed to the few data covering land regions in the Southern Hemisphere, such that the atmospheric $CO_2$ inversions aren't able to effectively distinguish between air-land $CO_2$ flux and air-sea $CO_2$ flux. Most of the results from the atmospheric $CO_2$ inversions show large negative anomalies in 1997–1998 ($-0.23\,\mathrm{Pg\,C\,yr^{-1}}$ on the average) that are larger than the anomalies found in the OBGCMs.

### 5.4 All Pacific Ocean regions 44.5° S–66° N

The time-averaged air-sea $CO_2$ flux for 1990–2009 described in the previous sections reveals a quite large range of variation among the different approaches when integrated over the Pacific Ocean between 44.5° S and 66° N (Table 4 and Fig. 6). It ranges from the weak $CO_2$ sink of $-0.22\,\mathrm{Pg\,C\,yr^{-1}}$ that was estimated from LDEO V2009 climatological $p\mathrm{CO_2sw}$ to a stronger sink of $-0.57 \pm 0.12\,\mathrm{Pg\,C\,yr^{-1}}$ from OBGCMs.





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The results from diagnostic models ($-0.27 \pm 0.13\,\mathrm{Pg\,C\,yr^{-1}}$), $p\mathrm{CO_2}$sw data assimilation ($-0.33\,\mathrm{Pg\,C\,yr^{-1}}$) and atmospheric $CO_2$ inversions ($-0.26 \pm 0.10\,\mathrm{Pg\,C\,yr^{-1}}$) that are more or less constrained by the measurements of $p\mathrm{CO_2}$sw are consistent with the estimate from the climatological $p\mathrm{CO_2}$sw. However, the estimate from the ocean interior $CO_2$ inversions ($-0.52 \pm 0.18\,\mathrm{Pg\,C\,yr^{-1}}$) is similar to the results from the OBGCMs. The smaller efflux from the tropics and larger or comparable influxes into the extra-tropics cooperatively contribute to the estimate of larger $CO_2$ influx into the Pacific from the OBGCMs and ocean interior $CO_2$ inversions. It is also interesting to note that the efflux from the tropics tends to be balanced by the influx into the North Pacific extra-tropics in the diagnostic models, but it tends to be balanced with the influx into the South Pacific extra-tropics in the estimate from the OBGCMs.

In regard to the inter-annual variability, the diagnostic models and OBGCMs are consistent with each other in that the ENSO-driven change in the tropical zone is playing a dominant role in the Pacific, and changes in the extra-tropics are minor as mentioned in Sects. 5.2 and 5.3 (Fig. 10). In the $p\mathrm{CO_2}$sw data assimilation and atmospheric $CO_2$ inversions, the inter-annual variability in the air-sea $CO_2$ flux in the tropical zone is also large, but the effect of ENSO events is less clear.

## 6 Discussion

### 6.1 Dependence of air-sea $CO_2$ flux upon wind-product

The accuracy of wind field products over the ocean is fundamental to evaluating air-sea $CO_2$ fluxes in all approaches considered in this study. In atmospheric $CO_2$ inversions, the wind field directly controls the transport of $CO_2$. The wind field in the marine boundary layer is also required to drive ocean circulation in the prognostic OBGCMs, the $p\mathrm{CO_2}$sw data assimilation, and the ocean interior $CO_2$ inversions. In addition, various gas transfer velocities for $CO_2$ at the air-sea interface have been given empirically as a function of wind speed at 10 m above sea surface and are needed to calculate air-sea $CO_2$ flux from the diagnostic models, forward OBGCMs and $p$$CO_2$sw data assimilation. Several wind products are available and have been used for these purposes. They include NCEP/NCAR Reanalysis 1 (Kalnay et al., 2007), NCEP/DOE Reanalysis 2 (Kanamitsu et al., 2002), ECMWF 40 yr Re-analysis (Uppala et al., 2005), JMA-CRIEPI JRA25/JCDAS (Onogi et al., 2007), and JPL CCMP Ocean Surface Wind Product (Ardizzone et al., 2009; Atlas et al., 2011). However, it is unclear how the choice of these wind products influences the resultant estimate of air-sea $CO_2$ fluxes, since a model is usually run with a single wind product and no comprehensive inter-comparison exercise has been made in terms of the difference in the wind fields.

In this section, we briefly describe the impact of the difference in wind products on the estimates of air-sea $CO_2$ flux for a diagnostic model (Sugimoto et al., 2012) and an OBGCM (Buitenhuis et al., 2010). The wind products used here are NCEP/NCAR Reanalysis 1 (NCEP1), which has been often used to force the forward ocean models, and JPL CCMP Ocean Surface Wind Components (CCMP) that we used to calculate the air-sea $CO_2$ flux with the LDEO V2009 climatological $p$$CO_2$sw and with $p$$CO_2$sw diagnostic models in this work. In the diagnostic model, the gas transfer velocity for $CO_2$ that was applied in Eq. (1) with the CCMP wind product has a functional form that depends on the monthly mean second moment of wind speed $\langle U^2 \rangle$ and an empirical coefficient of 0.25. On the other hand, monthly mean wind speed squared, $\langle U \rangle^2$, and the coefficient 0.39 has been applied to the $CO_2$ gas transfer velocity of Wanninkhof (1992) with the NCEP1 wind product. In the OBGCM, daily wind speed and a coefficient appropriate for short term wind speed of 0.3 has been applied with both wind products.

Mean seasonal variations and deseasonalized trends of regional mean wind speed, regionally-integrated air-sea $CO_2$ flux from a diagnostic model (Sugimoto et al., 2012) and that from a OBGCM (Buitenhuis et al., 2010) are shown in Figs. 11 and 12 for each sub-basin of the Pacific Ocean. The seasonality in the regional mean wind speed in the extra-tropics, i.e., stronger in winter and weaker in summer, is clearly seen in Fig. 11. For the inter-annual variability in regional mean wind speed (Fig. 12), a positive anomaly in 1997–1998 in the South Pacific and a trend towards increasing wind speed





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in the South and tropical Pacific are observed. It is also evident that the mean wind speed in CCMP is always stronger than that in NCEP1. The difference is larger in the tropics (18° S–18° N) than in the extra-tropics. In the tropical Pacific, the monthly mean wind speed from CCMP and from NCEP1 varied in parallel to each other, and the time-averaged wind speed over the period 1990–2008 was 6.4 $m\,s^{-1}$ in CCMP and 5.5 $m\,s^{-1}$ in NCEP1. In the North Pacific extra-tropics, no significant difference was seen in the regional mean wind speed in early summer (May–July). However, in winter (December–February), mean CCMP wind speed (8.9 $m\,s^{-1}$) is 0.6 $m\,s^{-1}$ stronger than that of NCEP1.

The difference in wind field influences the estimate of air-sea $CO_2$ fluxes from the diagnostic model and the forward ocean model in different ways. In the diagnostic modeling of Sugimoto et al. (2012), the air-sea $CO_2$ flux in the tropical Pacific calculated with the stronger CCMP wind field and that calculated with the weaker NCEP1 wind field agreed well both in the mean flux and in temporal variability (see Figs. 11 and 12). In spite of the large difference in the mean wind speed between CCMP and NCEP1, the difference in the air-sea $CO_2$ flux estimates was no more than 0.02 $Pg\,C\,yr^{-1}$ on average. This result indicates that the different formulations and coefficients of proportionality for CCMP and for NCEP1 in gas transfer velocity that have been individually calibrated with the bomb $^{14}C$ inventory in the global ocean successfully helped to account for the difference in the wind fields in the tropical Pacific. However, in the North and South Pacific extra-tropics, the weaker wind field of NCEP1 rather yielded stronger $CO_2$ sinks than the stronger wind field of CCMP. The differences are 0.07 $Pg\,C\,yr^{-1}$ for the South Pacific extra-tropics, 0.17 $Pg\,C\,yr^{-1}$ for the North Pacific extra-tropics, and 0.23 $Pg\,C\,yr^{-1}$ when integrated over the sub-basins of the Pacific. These differences in the estimate of regional air-sea $CO_2$ fluxes due to the use of different combinations of wind field and gas transfer velocity are comparable to, and therefore, may explain a large part of the difference in the estimate between diagnostic models and OBGCMs (Table 4 and Fig. 6) in the extra-tropics and in the entire Pacific.





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In contrast, the OBGCM of Buitenhuis et al. (2010) that has been forced with the stronger winds of CCMP (Table 2) yielded a stronger $CO_2$ source in the tropical Pacific and stronger sinks in the extra-tropics than that forced with the weaker winds of NCEP1. The magnitude of inter-annual variability in $CO_2$ outgassing in the tropical Pacific is also greater when the CCMP wind field was used to force the ocean model. The difference in the integrated $CO_2$ outflux amounted to $0.22\,\mathrm{Pg\,C\,yr^{-1}}$ in the tropical Pacific, and the differences in the integrated $CO_2$ sinks in the extra-tropics amounted to $0.13\,\mathrm{Pg\,C\,yr^{-1}}$ in the South Pacific and $0.09\,\mathrm{Pg\,C\,yr^{-1}}$ in the North Pacific. Moreover, it is interesting to note that these differences are offsetting between the tropical source and extra-tropical sinks, and consequently the difference in the flux estimate integrated over the Pacific sub-basins is minor ($< 0.02\,\mathrm{Pg\,C\,yr^{-1}}$).

Importantly, the sensitivity described here for one forward OBGCM is likely due to different responses of the subtropical cell overturning to the wind stress component of forcing. For example, if the subtropical overturning strength were to be determined by the strength of the trade winds across 12° N and 12° S, the differences in zonal wind stress at these latitudes could sustain differences in the overturning strength of the subtropical cells, and as a linear advection issue the supply of carbon in the upwelling cold tongue. This may also find expression in increased subduction rates in the extra-tropical source regions, as illustrated by Fig. 1 of Rodgers et al. (2003). The response of simulated fields of DIC and $p\mathrm{CO_2}$sw to the different wind forcing would of course vary from OBGCM to OBGCM. However, the results from this model suggest that the smaller $CO_2$ efflux from the tropical Pacific estimated by the forward OBGCMs than by the diagnostic models (Table 4 and Fig. 6) may reflect not only difference in winds used for calculating gas exchange, but also biases in the dynamical component of the wind forcing for the forward models (surface wind stress).





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## 6.2 The "best estimates" of air-sea $CO_2$ flux in the Pacific regions

A focus of this effort was to obtain "best estimates" of time-averaged net air-sea $CO_2$ flux in each of the three sub-basins as well as over the entirety of the Pacific Basin by synthesizing the estimates from a variety of approaches. However, it is now clear that the synthesis of the estimates for the air-sea $CO_2$ flux in the Pacific Ocean does not provide a robust or convincing quantitative path to define a "best estimate". Rather, this synthesis exercise has provided an important first step towards assembling the information that will be needed for future efforts to construct a best estimate.

A quantitative assessment building on the results presented here would certainly require skill weighting in the construction of a model-mean or a model-median value of the air-sea $CO_2$ flux. Although this type of quantitative effort will not be conducted here, we can make loose use of the expression "best estimate" to describe a flux diagnostic that is consistent with what is calculated with the other RECCAP efforts for the other major ocean basins (Schuster et al., 2013; Lenton et al., 2013). This "best estimate" for the air-sea $CO_2$ flux is taken as the average of the results from (a) the diagnostic models and (b) the ocean interior $CO_2$ inversions. Both of these approaches are anchored in observational constraints, with the data sources used by these two approaches being independent. As such, this "best estimate" is a simple average of the results obtained with surface $p\mathrm{CO_2}$sw constraints and results obtained with ocean interior tracer constraints. The uncertainty is then calculated from the uncertainties of the estimate from the diagnostic models ($\sigma_{\mathrm{dia}}$) and ocean $CO_2$ inversions ($\sigma_{\mathrm{ocn\_inv}}$) as $\{(1/2)^2 \cdot \sigma_{\mathrm{dia}}^2 + (1/2)^2 \cdot \sigma_{\mathrm{ocn\_inv}}^2\}^{1/2}$. Given the large uncertainty inherent in the gas transfer velocity calculated from the wind speed products, the uncertainty in the air-sea $CO_2$ flux from the diagnostic models ($\sigma_{\mathrm{dia}}$) was assumed to be 50 % of the flux (Wanninkhof et al., 2013). The estimates thus appraised are examined for consistency with the results from OBGCMs. For comparison with the estimates here, the riverine $CO_2$ flux, $+0.08\,\mathrm{Pg\,C\,yr^{-1}}$ in the North Pacific extra-tropics and $+0.04\,\mathrm{Pg\,C\,yr^{-1}}$ in the tropical Pacific (Jacobson et al., 2007), are added to the results from 6 of 8 OBGCMs in which



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riverine carbon discharge has not been taken into account (Table 2), and median ± MAD of flux estimates from all 8 OBGCs were recalculated.

In the North Pacific extra-tropics, the "best estimate" thus calculated is $-0.47 \pm 0.13\,\mathrm{Pg\,C\,yr^{-1}}$. This is consistent with the estimates from the OBGCMs ($-0.49 \pm 0.02\,\mathrm{Pg\,C\,yr^{-1}}$) (Table 4). Good consistency is also seen for the tropical Pacific where the "best estimate" is $+0.44 \pm 0.14\,\mathrm{Pg\,C\,yr^{-1}}$ and the median of the estimates from the OBGCMs is $+0.41 \pm 0.05\,\mathrm{Pg\,C\,yr^{-1}}$. In the South Pacific extra-tropics, the difference in the estimate of air-sea $CO_2$ flux between the diagnostic models and ocean interior $CO_2$ inversions is larger ($0.18\,\mathrm{Pg\,C\,yr^{-1}}$) than those in other regions (0.10 to $0.15\,\mathrm{Pg\,C\,yr^{-1}}$). The variation in the estimates among different OBGCMs ($-0.39 \pm 0.11\,\mathrm{Pg\,C\,yr^{-1}}$) is also large. However, the mean of the estimates from the diagnostic models and ocean interior $CO_2$ inversions ($-0.37 \pm 0.08\,\mathrm{Pg\,C\,yr^{-1}}$) is comparable to the estimate from the OBGCMs ($-0.39 \pm 0.11\,\mathrm{Pg\,C\,yr^{-1}}$). Finally, the "best estimate" of air-sea $CO_2$ flux for the entire Pacific basins to the north of 44.5° S ($-0.40 \pm 0.21\,\mathrm{Pg\,C\,yr^{-1}}$) was estimated from the results of diagnostic models ($-0.27 \pm 0.38\,\mathrm{Pg\,C\,yr^{-1}}$) and ocean interior $CO_2$ inversions ($-0.52 \pm 0.18\,\mathrm{Pg\,C\,yr^{-1}}$). Given the quite large uncertainties and discrepancies between these estimates, it was not possible to obtain estimates with small uncertainty. The estimate for the entire Pacific Ocean basin is in reasonable agreement with the sum of the estimates from the OBGCMs after adding the riverine $CO_2$ flux ($-0.45 \pm 0.18\,\mathrm{Pg\,C\,yr^{-1}}$).

Absolutely critical to future efforts to reduce uncertainty in estimating the Pacific carbon sink will be future expansion of the ocean carbon observing system. As we have seen through our synthesis, two central priorities in expanding the observing system should be improving our characterization of seasonal variability and more extensive data sampling of the South Pacific. For $p\mathrm{CO_2sw}$ based flux estimate, an important component of efforts to better estimate $CO_2$ uptake will involve the combined use of sea surface $p\mathrm{CO_2}$ measurements and diagnostic modeling. Similarly, it will be important to continue collection of hydrographic measurements of $CO_2$ chemistry given not only their intrinsic value but also their value to ocean inversion efforts. For both cases, the combined data/model analysis in the future will benefit greatly from the implementation and operation of autonomous platforms such as profiling floats and wave gliders mounted with the emerging technology of $CO_2$ and biogeochemical sensors (e.g., Martz et al., 2010; Fiedler et al., 2013). The autonomous platforms will certainly require coordinated efforts with the accurate measurements and calibration that are provided only by hydrographic measurements from research on oceanographic cruises. Additional measurements can be provided by efforts on voluntary observing ships. It is especially important that measurements are extended to fill in the data gaps in the Pacific Ocean with well considered and planned sampling strategies (e.g., Lenton et al., 2009), particularly in the Southern Hemisphere, and in the seasonal variability.

In addition to the aforementioned monitoring efforts, prognostic ocean modeling will also play an important role in continued development of process-understanding of the controls on physical-biogeochemical coupling in the ocean. As prognostic ocean models are also the basis of both ocean inversion studies and ocean biogeochemical assimilation efforts, they will directly benefit from better process representation in the models used for ocean carbon inversions. As was seen in Fig. 6, the largest discrepancies among the simulations with OBGCMs were found in the South Pacific. We will not identify the underlying cause of the discrepancies in detail within the context of this synthesis, but such discrepancies can result from differences in ocean model resolution, ocean physical parameterizations, and the representation of ocean biogeochemical processes. As was seen with the UEA model, important differences can also arise from differences in surface forcing fields.

## 7 Conclusions

In this study, a synthesis has been conducted of available observational products and modeling efforts to characterize the air-sea $CO_2$ fluxes over the Pacific Ocean basin. Consideration has been given to three regions, namely the extra-tropical North Pacific, the tropical Pacific, and the extra-tropical South Pacific. Consideration has also been





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given not only to the time-mean fluxes, but also to seasonal variability and interannual variability. With regard to the time-averaged air-sea $CO_2$ flux for 1990–2009, the estimates from all approaches are consistent in the sign of the flux for the tropical Pacific (efflux) as well as for the extra-tropics of the North and South Pacific (influx) (Table 4 and Fig. 6). In a considerable number of cases, the regional estimates agree within 0.1 $PgCyr^{-1}$. Some larger discrepancies are also seen between different approaches as well as among different models within the same approach.

In the tropical Pacific, time-averaged air-sea $CO_2$ fluxes over 1990–2009 from ocean interior $CO_2$ inversions ($+0.37 \pm 0.12\,PgCyr^{-1}$) and OBGCMs ($+0.39 \pm 0.04\,PgCyr^{-1}$) agree well, but they are smaller than estimates derived from $p$CO$_2$sw diagnostic models ($+0.52 \pm 0.25\,PgCyr^{-1}$). Nevertheless the differences are not significant if we consider the 50 % flux uncertainty associated with the gas exchange coefficient and under-sampling (Wanninkhof et al., 2013). Since the wind speed in the tropical Pacific has a quite large offset ($\sim 1\,ms^{-1}$) among the wind products and the estimate of air-sea $CO_2$ flux from OBGCMs is considered to be sensitive to the choice of wind product that forces the model, the improvement of wind-speed products could be one of the key issues in reconciling the discrepancies in the flux estimate among these models. For the inter-annual variability, its peak-to-peak amplitude in the OBGCMs ($0.40 \pm 0.09\,PgCyr^{-1}$) is larger than that of diagnostic models ($0.27 \pm 0.07\,PgCyr^{-1}$). The amplitude is also sensitive to the choice of wind product in the OBGCMs. The skill of the diagnostic models that potentially underestimate inter-annual variability also needs further examinations.

In the North Pacific extra-tropics, where $p$CO$_2$sw diagnostic models and ocean interior $CO_2$ inversions are relatively well constrained by the data of $p$CO$_2$sw and ocean interior DIC, the agreement of time-averaged air-sea $CO_2$ fluxes over 1990–2009 between these approaches ($-0.52 \pm 0.25\,PgCyr^{-1}$ and $-0.42 \pm 0.08\,PgCyr^{-1}$) is fair. The estimates from OBGCMs ($-0.57 \pm 0.02\,PgCyr^{-1}$) were more negative, but become consistent with other two approaches when riverine flux ($+0.08\,PgCyr^{-1}$) is added to the estimate. By contrast, the South Pacific extra-tropics remain severely under-sampled



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for $p$CO$_2$sw and ocean interior DIC. The discrepancy in the time-averaged air-sea $CO_2$ fluxes inferred from the $p$CO$_2$sw diagnostic model ($-0.28 \pm 0.13\,\mathrm{Pg\,C\,yr^{-1}}$) and ocean interior $CO_2$ inversions ($-0.46 \pm 0.10\,\mathrm{Pg\,C\,yr^{-1}}$) is larger than in the North Pacific. The discrepancy needs to be reconciled in the future primarily through the increase in measurements both at the surface and in the interior of the ocean. The development of an improved observation network that incorporates deployments of autonomous instruments with $CO_2$ and biogeochemical sensors has a great potential to contribute to filling in the data gaps in the Pacific Ocean, particularly in the Southern Hemisphere. The estimate from the OBGCMs ($-0.39 \pm 0.11\,\mathrm{Pg\,C\,yr^{-1}}$) in the South Pacific extra-tropics is rather consistent with the estimate from ocean interior $CO_2$ inversions, but its uncertainty is greater than twice as large as those in other sub-basins of the Pacific Ocean. The causes of such a large variation among OBGCMs in the South Pacific extra-tropics are yet to be clarified.

For the entire Pacific Ocean between 44.5° S and 66° N, the time-averaged net air-sea $CO_2$ flux for 1990–2009 is $-0.27 \pm 0.38\,\mathrm{Pg\,C\,yr^{-1}}$ from $p$CO$_2$sw diagnostic models, $-0.52 \pm 0.18\,\mathrm{Pg\,C\,yr^{-1}}$ from ocean interior $CO_2$ inversions, and $-0.57 \pm 0.12\,\mathrm{Pg\,C\,yr^{-1}}$ from the OBGCMs. The discrepancies in the estimates in the sub-basins of the Pacific between the diagnostic models and other two approaches were reinforced when integrated over the Pacific Ocean. On the other hand, the diagnostic models and OBGCMs are consistent with each other in that the inter-annual variability in the air-sea $CO_2$ flux in the Pacific Ocean is dominated by the inter-annual variability in the tropical zone and is associated with the ENSO events.

The model with $p$CO$_2$sw data assimilation gave the smallest efflux in the tropics ($+0.27\,\mathrm{Pg\,C\,yr^{-1}}$) and the smallest influx into the extra-tropics of the North Pacific ($-0.37\,\mathrm{Pg\,C\,yr^{-1}}$) and the South Pacific ($-0.24\,\mathrm{Pg\,C\,yr^{-1}}$) among the approaches considered in this study. The phase of inter-annual annual variability in the tropics is not always consistent with the results from diagnostic models and OBGCMs.

Atmospheric $CO_2$ inversions gave the largest variety of estimates of time-averaged air-sea $CO_2$ flux in each of the North Pacific extra-tropics ($-0.48 \pm 0.08\,\mathrm{Pg\,C\,yr^{-1}}$), trop-





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ics ($+0.53\pm0.08\,\mathrm{Pg\,C\,yr^{-1}}$), and the South Pacific extra-tropics ($-0.29\pm0.08\,\mathrm{Pg\,C\,yr^{-1}}$). The median of the estimates is fairly consistent with the estimate from the LDEO V2009 climatological $p\mathrm{CO_2}$sw and $p\mathrm{CO_2}$sw diagnostic models, possibly because of the use of the climatological $CO_2$ flux from $p\mathrm{CO_2}$sw in the flux priors.

## Appendix

### Decadal mean air-sea $CO_2$ fluxes

Regionally-integrated and time-averaged net air-sea $CO_2$ flux ($\mathrm{Pg\,C\,yr^{-1}}$) for 1990–1999 and 2000–2009 in the Pacific Ocean regions from various approaches are shown in Table A1.

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**Table 1.** List of diagnostic models included in this study.

| Abbreviation | Reference | Period evaluated[1] |
|---|---|---|
| Park_2010 | Park et al. (2010) | 1990–2009 |
| Sugimoto_2012 | Sugimoto et al. (2012) | 1990–2009 |
| Nakaoka_2013[2] | Nakaoka et al. (2013) | 2002–2008 |

[1] Period evaluated in this study. For "Park_2010" and "Sugimoto_2012", models have been run for longer.
[2] For the North Pacific extra-tropics only.



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**Table 2.** List of prognostic ocean biogeochemistry/general circulation models and a $p\mathrm{CO_2}$sw data-assimilation system included in this study.

| Name | Abbreviation | Reference | Period evaluated[1] |
|---|---|---|---|
| CCSM-BEC | BEC | Doney et al. (2009a, b) | 1990–2009 |
| MICOM-HAMOCCv1 | BER | Assmann et al. (2010) | 1990–2009 |
| CSIRO-BOGCM | CSIRO | Matear and Lenton (2008) | 1990–2009 |
| CCSM-ETHk15[2] | ETHk15 | Graven et al. (2012) | 1990–2007 |
| MOM4-TOPAZ | GFDL | Dunne et al. (2012) | 1990–2004 |
| NEMO-PISCES[3] | IPSL | Aumont and Bopp (2006) | 1990–2009 |
| MRI.COM | MRI | Nakano et al. (2011) | 1990–2007 |
| NEMO-PlankTOM5NCEP[3,4] | UEA_NCEP1 | Buitenhuis et al. (2010) | 1990–2009 |
| NEMO-PlankTOM5CCMP[3,5] | UEA_CCMP | Buitenhuis et al. (2010) | 1990–2009 |
| OTTM-CO2 assimilation | OTTM | Valsala and Maksyutov (2010) | 1990–2008 |

[1] Period evaluated in this study. Models have been run for longer.
[2] ETHk15: CCSM-ETH model with a prescribed global average gas transfer velocity of 15 $\mathrm{cm\,h^{-1}}$.
[3] River carbon discharge has been considered.
[4] UEA_NCEP: NEMO-PlankTOM5 model with NCEP core forcing.
[5] UEA_CCMP: NEMO-PlankTOM5 model with NCEP core forcing (heat, precipitation etc.) but using CCMP winds for both ocean circulation and gas exchange.





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**Table 3.** List of atmospheric $CO_2$ inversions included in this study.

| Abbreviation | Reference | Period evaluated[1] |
|---|---|---|
| LSCE an v2.1 | Piao et al. (2004) | 1996–2004 |
| LSCE var v1.0 | Chevallier et al. (2010) | 1990–2008 |
| C13 CCAM law | Rayner et al. (2008) | 1992–2008 |
| C13 MATCH rayner | Rayner et al. (2008) | 1992–2008 |
| CTracker US | Peters et al. (2007) | 2001–2008 |
| JENA s96 v3.3 | Rödenbeck (2005) | 1996–2008 |
| RIGC patra | Patra et al. (2005) | 1993–2006 |
| JMA, 2010 | Maki et al. (2010) | 1990–2008 |
| TRCOM mean 9008 | Baker et al. (2006) | 1990–2008 |
| NICAM niwa | Niwa et al. (2012) | 1990–2007 |

[1] Period evaluated in this study. Some inversions may have been run for longer time.

**Table 4.** Mean and amplitude of inter-annual variability (IAV) of regionally-integrated air-sea $CO_2$ flux in the sub-basins of the Pacific Ocean for the period 1990–2009. All units are $Pg\,C\,yr^{-1}$.

| Regions [Area ($km^2$)] | | LDEO V2009[1] | $p$CO$_2$sw Diag. Models[2] | Ocean Inv.[3] | OBGC Models[4] | OBGCM + River -ine flux[5] | $p$CO$_2$sw Data Assim.[6] | Atm. Inv.[4,6] | Best Estimates[7] |
|---|---|---|---|---|---|---|---|---|---|
| North Pacific extra- | Mean | −0.44 ±0.21 | −0.52 ±0.05 ±0.25 | −0.42 ±0.08 | −0.57 ±0.02 | −0.49 ±0.02 | −0.37 | −0.48 ±0.08 | −0.47 ±0.13 |
| tropics 66–18° N [$4.43 \times 10^7$] | IAV[8] | | 0.12 ±0.02 | | 0.11 ±0.02 | | 0.19 | 0.41 ±0.06 | |
| Tropical Pacific 18° N | Mean | +0.51 ±0.24 | +0.52 ±0.09 ±0.25 | +0.37 ±0.12 | +0.39 ±0.04 | +0.41 ±0.05 | +0.27 | +0.53 ±0.08 | +0.44 ±0.14 |
| –18° S [$6.65 \times 10^7$] | IAV[8] | | 0.27 ±0.07 | | 0.40 ±0.09 | | 0.34 | 0.48 ±0.15 | |
| South Pacific extra- | Mean | −0.29 ±0.14 | −0.28 ±0.00 ±0.13 | −0.46 ±0.10 | −0.39 ±0.11 | −0.39 ±0.11 | −0.24 | −0.29 ±0.08 | −0.37 ±0.08 |
| tropics 18–44.5° S [$3.88 \times 10^7$] | IAV[8] | | 0.08 ±0.03 | | 0.12 ±0.03 | | 0.11 | 0.64 ±0.11 | |
| All Pacific 66° N | Mean | −0.22 ±0.35 | −0.27 ±0.13 ±0.38 | −0.52 ±0.18 | −0.57 ±0.12 | −0.45 ±0.18 | −0.33 | −0.26 ±0.10 | −0.40 ±0.21 |
| –44.5° S [$14.96 \times 10^7$] | IAV[8] | | 0.25 ±0.20 | | 0.49 ±0.08 | | 0.45 | 1.08 ±0.29 | |

[1] Climatological flux at year 2000 calculated with CCMP wind product. Error of ±48 % are applied to the flux in each sub-basin and error of ± sqrt{$\sum$(error in a sub-basin)$^2$} is applied for the all Pacific. Components of error assumed is ±10 % for $\Delta p CO_2$, ±30 % for gas transfer scaling factor (0.25), ±20 % for wind speed, ±25 % for normalization for the year 2000, ±15 % for under sampling (based on SST).
[2] Mean ± half of range of two diagnostic models of Park et al. (2010) and Sugimoto et al. (2012) and ±48 % error as in the LDEO V2009 climatological fluxes.
[3] Skill-weighted average ± standard deviation of the ocean $CO_2$ inversions.
[4] Median ± median absolute deviation of estimates from various models and inversions.
[5] No errors in the riverine $CO_2$ flux is considered.
[6] For the period 1990–2008.
[7] The estimates given here are the mean values from the $p$CO$_2$sw diagnostic models and ocean inversions. Uncertainty has been given as sqrt{$(0.5)^2 \cdot$ sigma(Diag)$^2$ + $(0.5)^2 \cdot$ sigma(OcnInv)$^2$}
[8] Inter-annual variability, i.e., peak-to-peak difference of the annual mean flux.





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**Table A1.** Decadal mean values of the regionally-integrated air-sea $CO_2$ flux in three sub-basins of the Pacific Ocean evaluated from various approaches. All units are $Pg\,C\,yr^{-1}$.

| Regions [Area (km$^2$)] | Period | $p$CO$_2$sw Diag. Models[1] | Ocean Inv.[2] | OBGC Models[3] | $p$CO$_2$sw Data Assim.[4] | Atm. Inv.[3,4] |
|---|---|---|---|---|---|---|
| North Pacific extra-tropics 66–18° N [4.43 × 10$^7$] | 1990–1999 | −0.51 ± 0.03 | −0.42 ± 0.08 | −0.58 ± 0.03 | −0.37 | −0.44 ± 0.08 |
| | 2000–2009 | −0.53 ± 0.07 | −0.42 ± 0.08 | −0.57 ± 0.02 | −0.36 | −0.53 ± 0.07 |
| Tropical Pacific 18° N–18° S [6.65 × 10$^7$] | 1990–1999 | +0.49 ± 0.07 | +0.37 ± 0.12 | +0.36 ± 0.06 | +0.27 | +0.53 ± 0.07 |
| | 2000–2009 | +0.56 ± 0.11 | +0.36 ± 0.12 | +0.41 ± 0.04 | +0.28 | +0.54 ± 0.07 |
| South Pacific extra-tropics 18–44.5° S [3.88 × 10$^7$] | 1990–1999 | −0.27 ± 0.01 | −0.46 ± 0.10 | −0.38 ± 0.11 | −0.23 | −0.26 ± 0.09 |
| | 2000–2009 | −0.29 ± 0.01 | −0.46 ± 0.10 | −0.40 ± 0.12 | −0.24 | −0.29 ± 0.07 |
| All Pacific 66° N–44.5° S [14.96 × 10$^7$] | 1990–1999 | −0.29 ± 0.08 | −0.51 ± 0.18 | −0.61 ± 0.12 | −0.34 | −0.22 ± 0.17 |
| | 2000–2009 | −0.25 ± 0.18 | −0.52 ± 0.18 | −0.53 ± 0.13 | −0.33 | −0.35 ± 0.07 |

[1] Mean ± half of range of two diagnostic models of Park et al. (2010) and Sugimoto et al. (2012).
[2] Skill-weighted average ± standard deviation of the ocean $CO_2$ inversions.
[3] Median ± median absolute deviation of estimates from various models and inversions.
[4] For the period 1990–2008.





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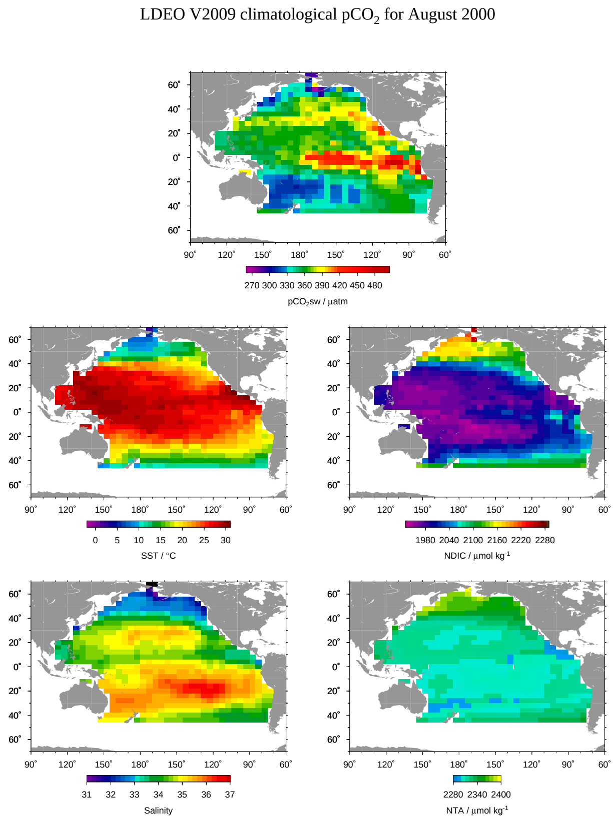


**Fig. 1.** Data-based climatology in August of *p*$CO_2$sw (top panel), temperature (middle left), salinity (bottom left) from LDEO V2009 (Takahashi et al., 2009a), and salinity-normalized ($S$ = 35) DIC (middle right) calculated with total alkalinity derived from Lee et al. (2006) (bottom right).





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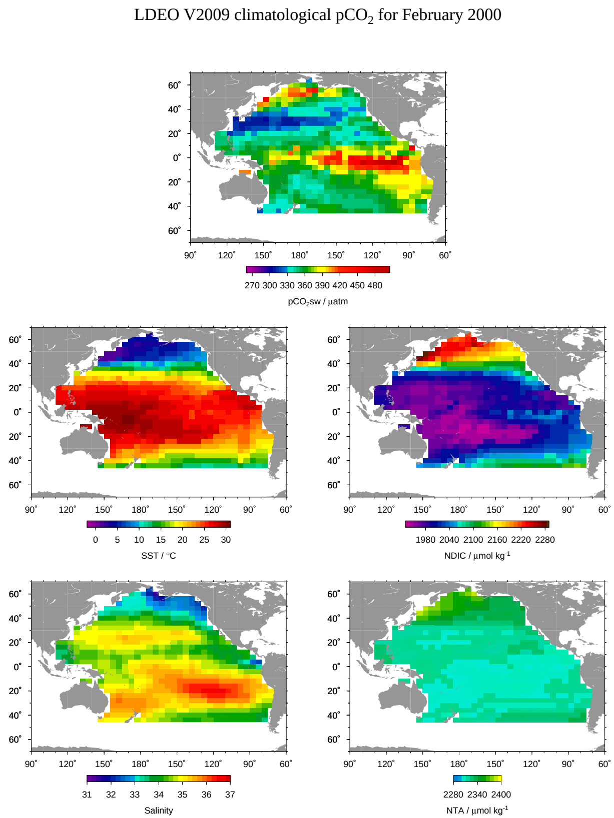


**Fig. 2.** Same as Fig. 1 but in February.





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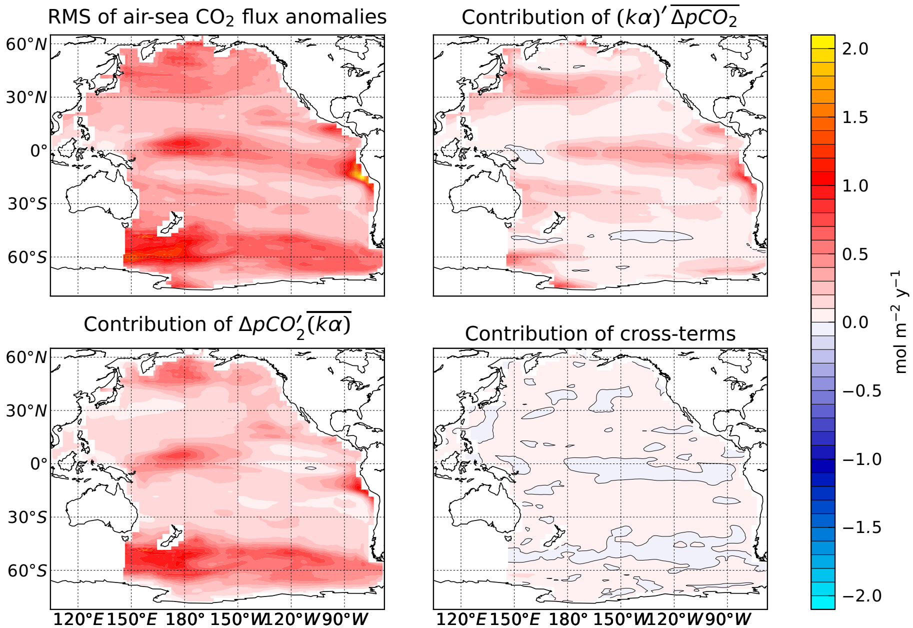


**Fig. 3.** Partitioning of the mechanisms driving inter-annual variability in air–sea $CO_2$ flux (mol m$^{-2}$ yr$^{-1}$) in the Pacific Ocean in the CCSM ocean BEC model (Doney et al., 2009a). The panels show the root mean square (rms) of the model deseasonalized $CO_2$ flux anomalies (1990–2009) (top left) and the contributions from gas transfer velocity (wind speed and ice cover) (top right), surface-water $\Delta pCO_2$ (lower left), and the cross-correlation of gas transfer and $pCO_2$sw anomalies (lower right).

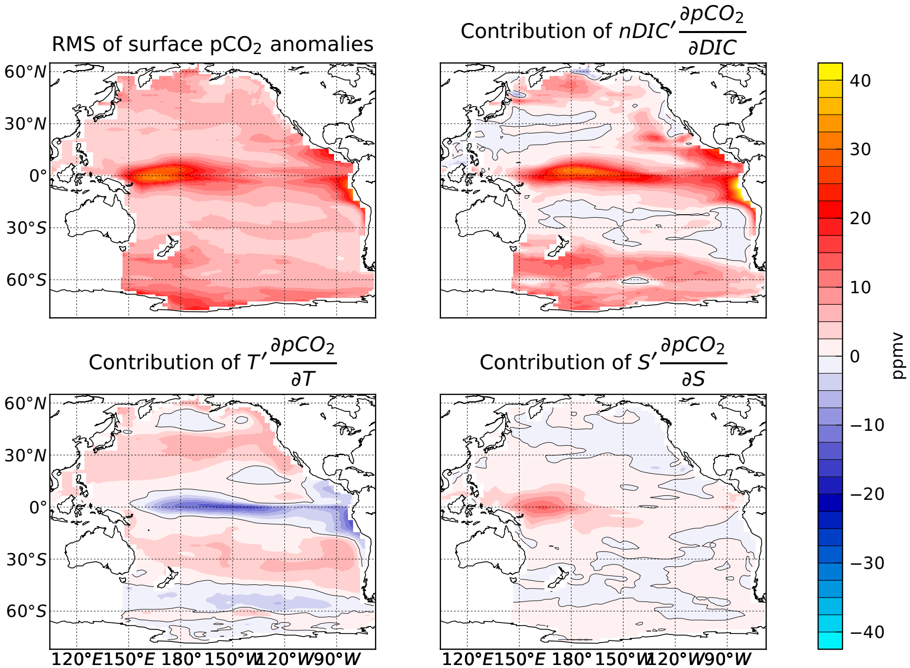


**Fig. 4.** Partitioning of the mechanisms driving inter-annual variability in the air–sea $pCO_2$ difference, $\Delta pCO_2$ (µatm), in the CCSM BEC ocean model ppmv is numerically equal to $\Delta pCO_2$ (µatm) when the barometric pressure is assumed to be 1.0 atmosphere. The panels show the root mean square of the model deseasonalized surface-water $\Delta pCO_2$ anomalies (1990–2009) (top left) and the contributions from surface-water salinity-normalized dissolved inorganic carbon (DIC; top right), temperature (lower left), and fresh water/salinity (lower right). The contributions from salinity-normalized alkalinity (not shown) are generally negligible.





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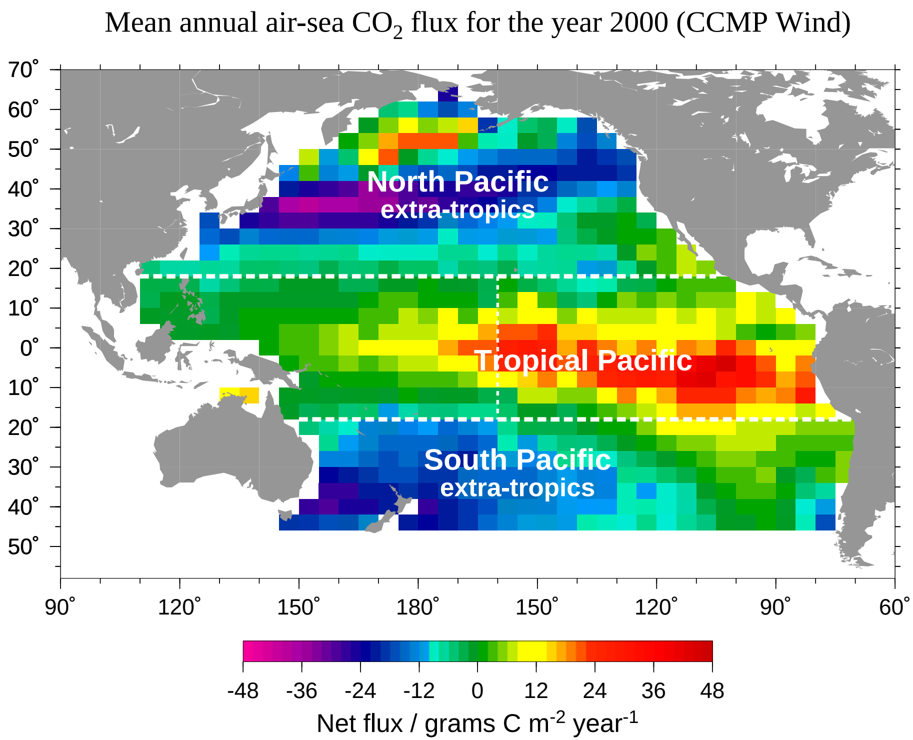


**Fig. 5.** Three Pacific Ocean sub-basins defined according to the aggregation of 14 Pacific regions of the ocean interior $CO_2$ inversions (Mikaloff Fletcher, 2006). Superposed is the mean annual air-sea $CO_2$ flux for the year 2000 calculated with climatological $p$$CO_2$sw of LDEO V2009 (Takahashi et al., 2009a) and CCMP wind (Ardizzone et al., 2009; Atlas et al., 2011). Positive fluxes are out of ocean and negative fluxes are into the ocean.





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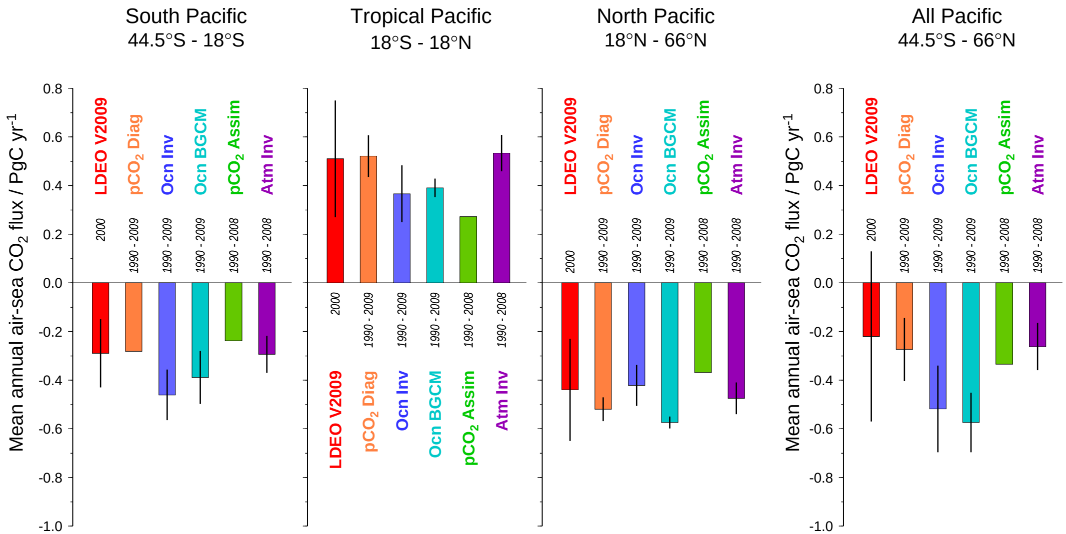


**Fig. 6.** Summary of regionally-integrated and time-averaged net air-sea $CO_2$ flux (Pg C $yr^{-1}$) in the Pacific Ocean sub-basins shown in Fig. 5.





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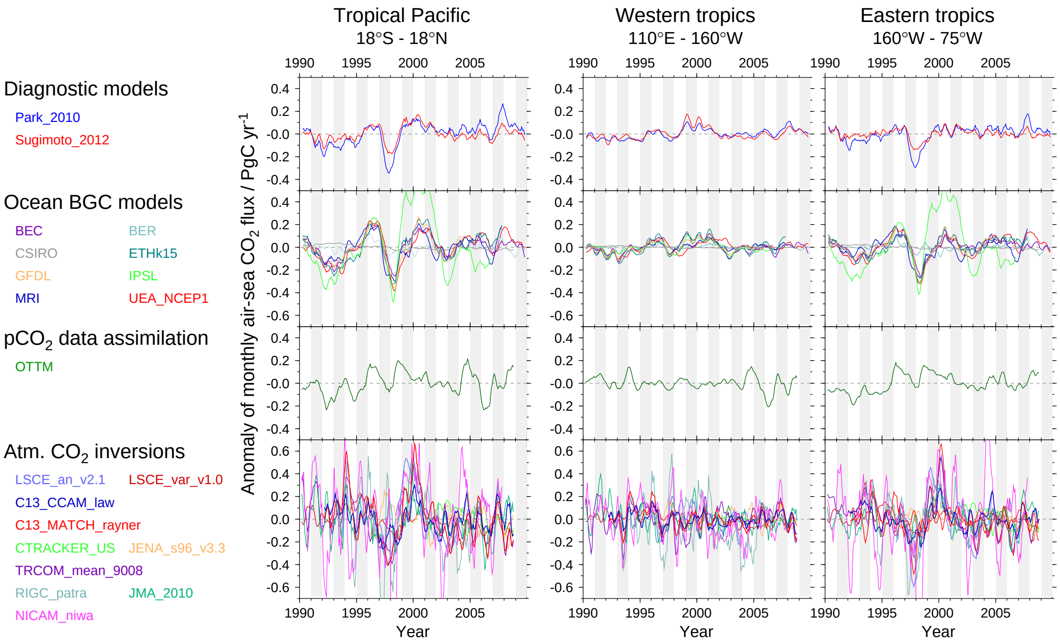


**Fig. 7.** Trend of air-sea $CO_2$ flux anomalies (5 months running means) in the tropical Pacific (18° S–18° N) (left panel) for 1990–2009, and its components in the western tropical sector to the west of 160° W (middle) and in the eastern tropical sector to the east of 160° W.





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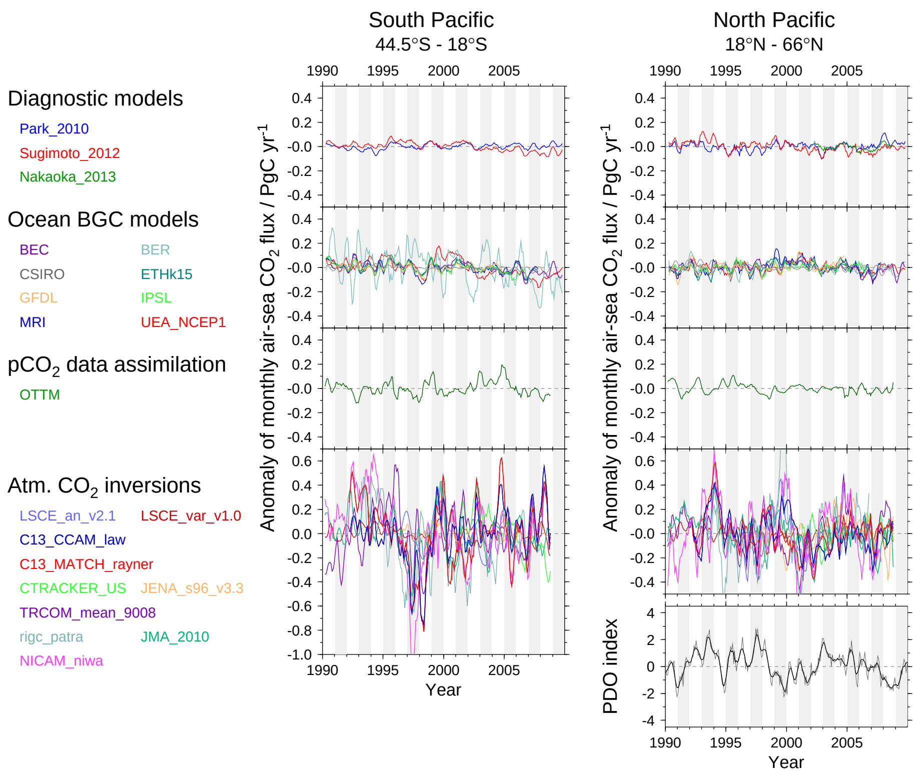


**Fig. 8.** Trend of air-sea $CO_2$ flux anomalies (5 month running means) in the South Pacific extra-tropics (44.5° S–18° S) (left) and in the North Pacific extra-tropics (18° N–66° N) (right) for 1990–2009. Also shown at the bottom of right panel is the Index of Pacific Decadal Oscillation (http://ds.data.jma.go.jp/tcc/tcc/products/elnino/decadal/pdo.html).





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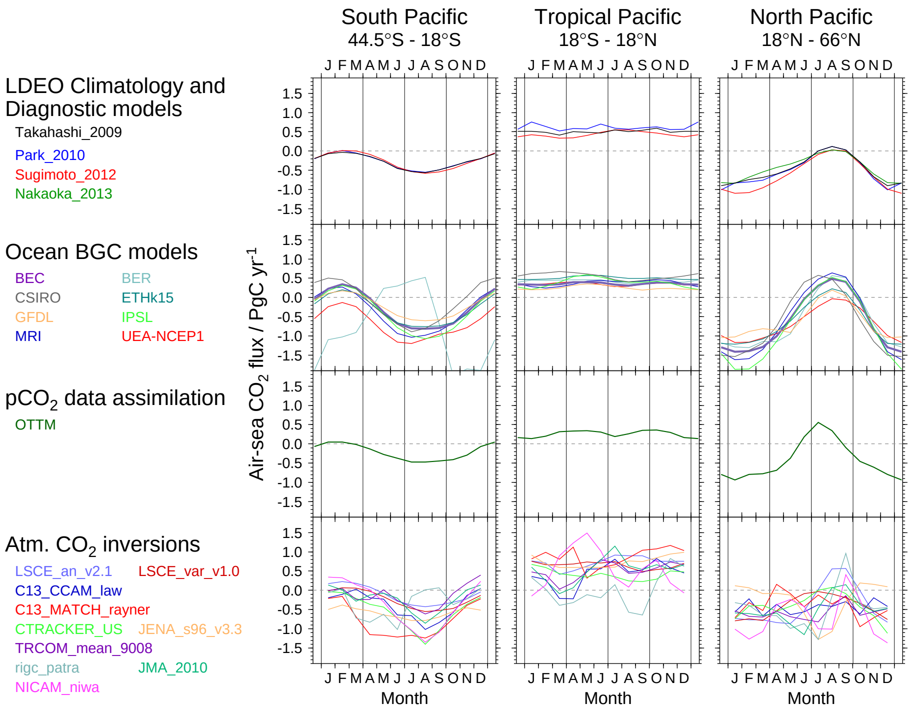


**Fig. 9.** Mean monthly variations of air-sea $CO_2$ flux in the South Pacific extra-tropics (44.5° S–18° S) (left), in the tropical Pacific (18° S–18° N) (middle) and in the North Pacific extra-tropics (18° N–66° N) (right) for 1990–2009.

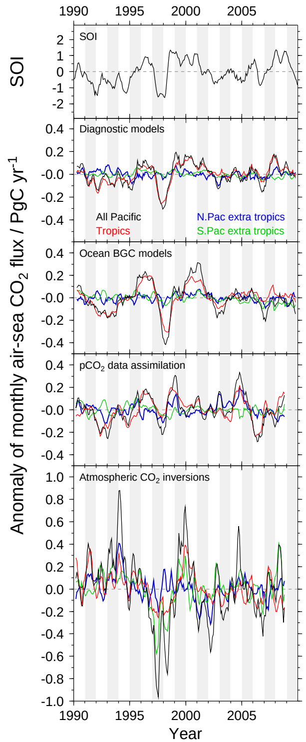


**Fig. 10.** Trend (5 month running means) of the Southern Oscillation Index (SOI) (http://ds.data.jma.go.jp/tcc/tcc/products/elnino/index/), and the air-sea $CO_2$ flux anomalies over the Pacific (44.5° S–66° N; black), in the tropical Pacific (18° S–18° N; red), in the North Pacific extra-tropics (18° N–18° S; blue), and in the South Pacific extra-tropics (green).

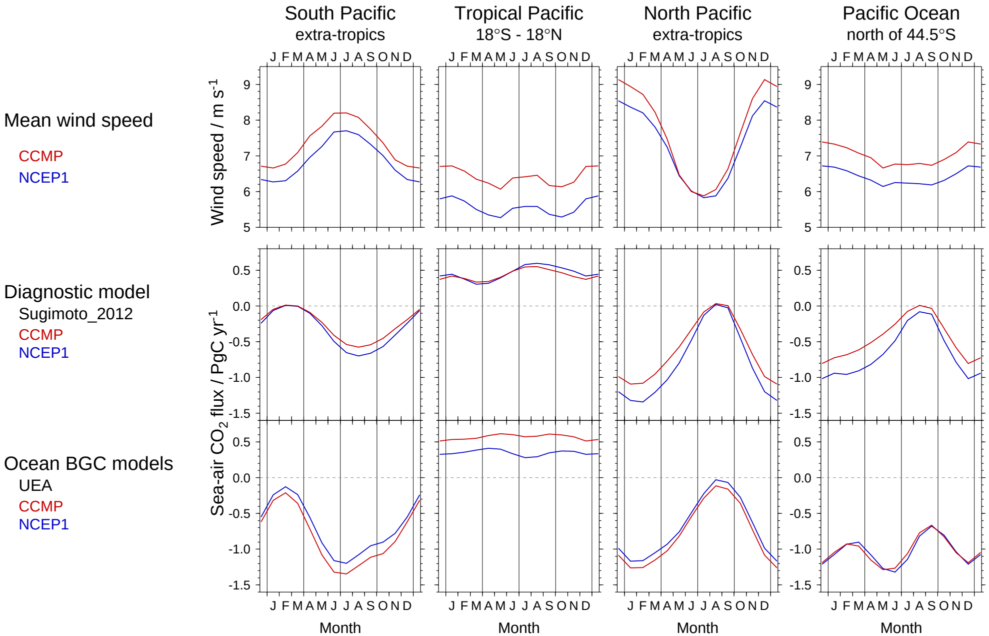


**Fig. 11.** Mean monthly wind speed of NCEP/NCAR Reanalysis 1 (NCEP1) (Kalnay et al., 2007) and Cross-Calibrated, Multi-Platform (CCMP) Ocean Surface Wind Product (http://podaac.jpl.nasa.gov/DATA_CATALOG/ccmpinfo.html) (Ardizzone et al., 2009; Atlas et al., 2011) (top), and mean monthly air-sea $CO_2$ flux for 1990–2009 calculated with the diagnostic model of Sugimoto et al. (2012) (middle) and with the ocean biogeochemistry/general circulation model of Buitenhuis et al. (2010) (bottom) using the surface wind fields of NCEP1 and CCMP in the South Pacific extra-tropics (left), in the tropical Pacific (middle-left), in the North Pacific extra-tropics (middle-right), and in the all Pacific regions shown in Fig. 5.





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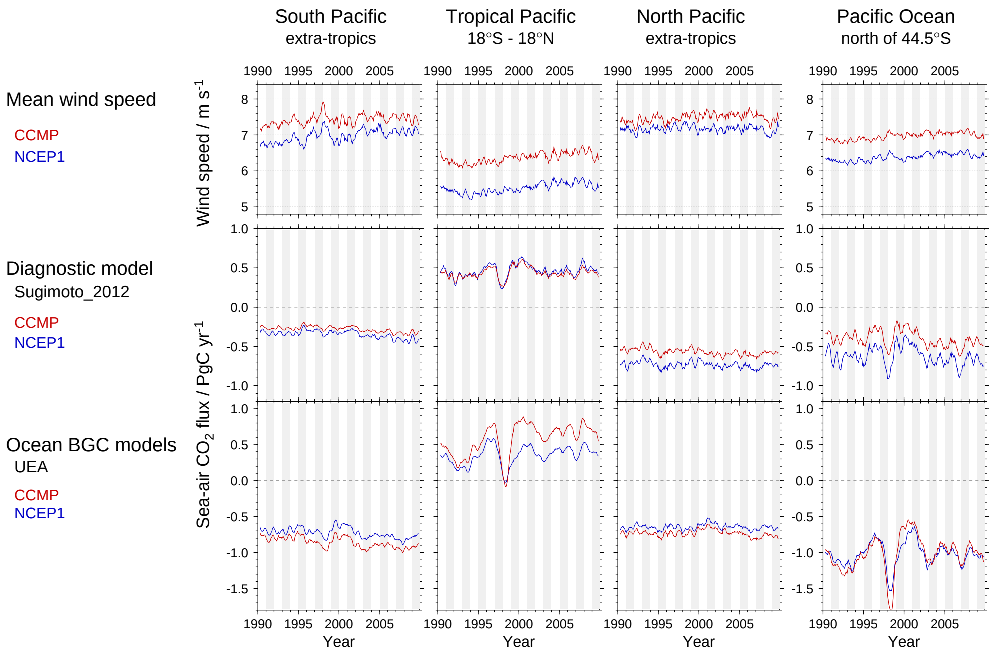


**Fig. 12.** Trend (5 month running means) of the monthly anomalies of wind speed of NCEP/NCAR Reanalysis 1 (NCEP1) (Kalnay et al., 2007) and Cross-Calibrated, Multi-Platform (CCMP) Ocean Surface Wind Product (http://podaac.jpl.nasa.gov/DATA_CATALOG/ccmpinfo.html) (Ardizzone et al., 2009; Atlas et al., 2011) (top), and air-sea $CO_2$ flux for 1990–2009 calculated with the diagnostic model of Sugimoto et al. (2012) (middle) and with the ocean biogeochemistry/general circulation model of Buitenhuis et al. (2010) (bottom) using the surface wind fields of NCEP1 and CCMP in the South Pacific extra-tropics (45° S–18° S) (left), in the tropical Pacific (middle-left), in the North Pacific extra-tropics (18° N–66° N) (middle-right), and in the whole Pacific regions shown in Fig. 5.





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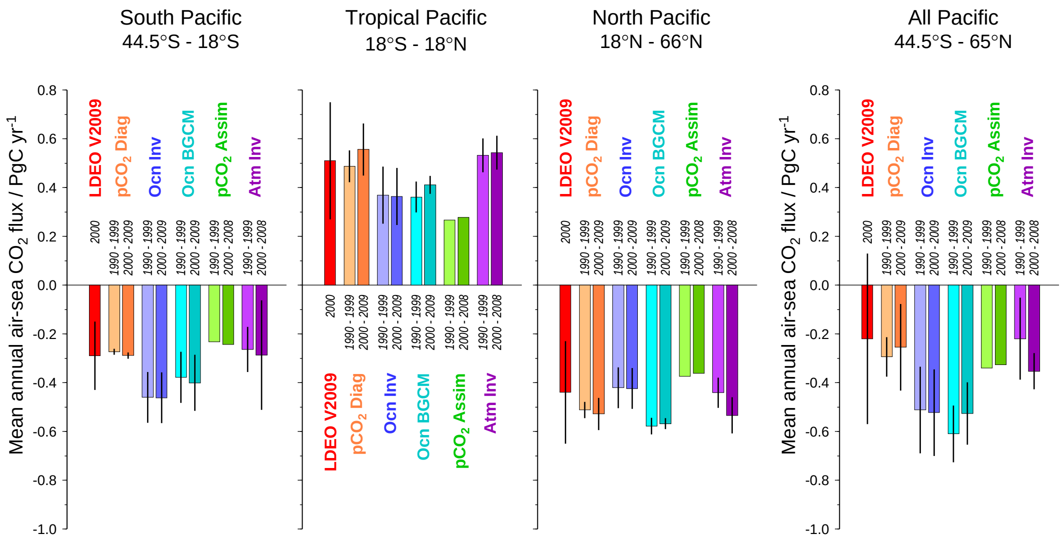


**Fig. A1.** Summary of regionally-integrated and time-averaged net air-sea $CO_2$ flux ($PgC\,yr^{-1}$) for 1990–1999 and 2000–2009 in the Pacific Ocean regions shown in Fig. 5.





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