

# Oceanic sources and sinks of atmospheric CO<sub>2</sub>

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## Abstract.

We present new estimates of the contemporary net air-sea CO<sub>2</sub> fluxes based on the inversion of interior ocean carbon observations using a suite of 10 Ocean General Circulation Models. The comparison of these inverse estimates with those based on a greatly expanded climatology of the air-sea difference of the partial pressure of CO<sub>2</sub> reveal a consistent description of the regional distribution of annual mean sources and sinks of atmospheric CO<sub>2</sub>. This distribution is characterized by outgassing in the tropics, uptake in mid-latitudes, and comparatively small fluxes in the high-latitudes. In particular, both estimates point toward a substantially smaller present CO<sub>2</sub> sink in the Southern Ocean than previous estimates [Takahashi *et al.*, 2002; Gurney *et al.*, 2002, 2004; Watson and Orr, 2003]. The inversion permits us to attribute this small sink to a near cancellation between a substantial outgassing of natural CO<sub>2</sub> and a strong uptake of anthropogenic CO<sub>2</sub>. Globally, the inverse estimate of the uptake of anthropogenic CO<sub>2</sub> by the ocean amounts to  $2.20 \pm 0.25$  Pg C yr<sup>-1</sup> for a nominal year of 1995, in excellent agreement with most global estimates. Interpreted in the context of the global anthropogenic CO<sub>2</sub> budget for the 1990s, this oceanic uptake rate constrains the balance of the terrestrial biosphere for this decade to a net sink of  $9 \pm 5$  Pg C.

## 1. Introduction

The exchange of carbon dioxide (CO<sub>2</sub>) between the atmosphere and ocean is a critical process of the global carbon cycle and an important determinant of the future of the Earth system [Fung *et al.*, 2005; Friedlingstein *et al.*, 2006]. From ~1800 until 1994, the ocean removed about  $118 \pm 19$  Pg C (1 Pg = 10<sup>15</sup> g) from the atmosphere [Sabine *et al.*, 2004]. This is equivalent to about 50% of the CO<sub>2</sub> emitted into the atmosphere from the burning of fossil fuels or about 30% of

the total anthropogenic CO<sub>2</sub> emissions, which additionally include emissions from land use change and cement production [Houghton, 2003b; DeFries *et al.*, 2002; Achard *et al.*, 2002; Marland *et al.*, 2006]. With this removal of anthropogenic CO<sub>2</sub>, the ocean constitutes the only net sink over the last 200 years, as the terrestrial biosphere is most likely a source when integrated over this period [Sabine *et al.*, 2004]. While the current sink strengths of the ocean and the land biosphere are similar [Prentice *et al.*, 2001], model projections suggest that the land sink may decrease during this century or, perhaps, even turn into a source [Cox *et al.*, 2000]. In contrast, the oceanic sink for atmospheric CO<sub>2</sub> will likely continue to grow [Orr *et al.*, 2001], highlighting the crucial role of the ocean as the ultimate sink for anthropogenic CO<sub>2</sub>. However, quantitative estimates of the oceanic sink strength and its regional distribution have remained uncertain, particularly for key regions such as the Southern Ocean [Caldeira and Duffy, 2000; Orr *et al.*, 2001; Watson and Orr, 2003; Roy *et al.*, 2003].

A fundamental challenge is that the contemporary exchange flux of CO<sub>2</sub> across the air-sea interface consists of a natural CO<sub>2</sub> flux component, i.e., a CO<sub>2</sub> flux that already existed in pre-industrial times and is assumed to have changed little since, and an anthropogenic CO<sub>2</sub> flux component that is driven by the anthropogenic perturbation in atmospheric CO<sub>2</sub>. The fluxes of both anthropogenic and natural CO<sub>2</sub> are expected to change considerably in the future in response to continually rising atmospheric CO<sub>2</sub> and global climate change [Sarmiento *et al.*, 1998; Joos *et al.*, 1999; Matear and Hirst, 1999; Plattner *et al.*, 2001; Gruber *et al.*, 2004]. Therefore, in order to predict the future of the oceanic carbon sink and consequently the future evolution of atmospheric CO<sub>2</sub>, we need a quantitative estimate of the current sources and sinks for atmospheric CO<sub>2</sub>, as well as a mechanistic understanding of both components of the CO<sub>2</sub> flux.

Until recently, the ability to separately estimate these two flux components from observations was limited to the very indirect approach of computing the horizontal flux divergence of the oceanic transport of the respective carbon component (e.g., Holfort *et al.* [1998]). A more direct and often

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used approach, albeit without ability to separate the net flux into its natural and anthropogenic CO<sub>2</sub> components, is the measurement of the air-sea difference of the partial pressure of CO<sub>2</sub> ( $p\text{CO}_2$ ), which by combining with bulk gas exchange parameterizations yields an estimate of the net flux [Takahashi et al., 1997, 1999, 2002]. Another method to estimate net air-sea fluxes is the inversion of atmospheric CO<sub>2</sub> observations [Bolin and Keeling, 1963; Enting and Mansbridge, 1989; Keeling et al., 1989; Tans et al., 1990; Gurney et al., 2002, 2004]. The air-sea CO<sub>2</sub> flux constraints by these approaches have remained somewhat limited, owing to a combination of methodological uncertainties (e.g., uncertainty in the parameters for the bulk parameterization [Wanninkhof, 1992; Krakauer et al., 2006], uncertain atmospheric transports [Gurney et al., 2004]), and insufficient data coverage [Gloor et al., 2000; Takahashi et al., 2002].

A region of particularly large discrepancies between different flux estimates is the Southern Ocean, which we define here as the oceanic region south of 44°S, with some estimates indicating a very large uptake flux (order of 1 Pg C yr<sup>-1</sup> and higher [Takahashi et al., 1997, 1999]) and others pointing toward a much lower uptake [Roy et al., 2003]. This represents a critical gap in our understanding of the ocean carbon cycle, as simulations indicate that the Southern Ocean likely will dominate the ocean’s CO<sub>2</sub> flux response during this century [Sarmiento et al., 1998; Orr et al., 2001], and also that it is the leading candidate region to have controlled the glacial-interglacial changes in atmospheric CO<sub>2</sub> (e.g., Sigman and Boyle [2000]; Toggweiler [1999]; Gildor et al. [2002]).

We synthesize here the results from a recently developed ocean inversion method [Gloor et al., 2003; Mikaloff Fletcher et al., 2006a, b], which provides air-sea CO<sub>2</sub> flux estimates that are based on inorganic carbon observations from the ocean interior, and are therefore entirely independent from estimates based on observations of the air-sea  $p\text{CO}_2$  difference or those based on atmospheric CO<sub>2</sub>. An added benefit is that the ocean inversion separately estimates the air-sea fluxes of natural and anthropogenic CO<sub>2</sub>, which when added together give an estimate of the contemporary net air-sea flux of CO<sub>2</sub>. The anthropogenic CO<sub>2</sub> flux results have been presented and discussed by Mikaloff Fletcher et al. [2006a], while Mikaloff Fletcher et al. [2006b] summarized and discussed the natural CO<sub>2</sub> flux results. This paper here focuses on the contemporary air-sea CO<sub>2</sub> flux results and highlights the global-scale implications. The results presented here are quantitatively very similar to those reported by Jacobson et al. [2006b] on the basis of a joint atmosphere-ocean inversion, since the oceanic constraints in this joint inversion are much stronger than the atmospheric ones, resulting in only small shifts from an ocean-only inversion. Relative to Jacobson et al. [2006b], the fluxes reported here are based on a larger set of oceanic models, thereby providing an improved assessment of the uncertainties stemming from the ocean models. An additional new element is that we compare here our inverse estimates of the contemporary air-sea CO<sub>2</sub> flux with new estimates based on an updated and greatly expanded analysis of oceanic  $p\text{CO}_2$  observations by Takahashi et al. [2006]. These two independent sets provide an unprecedentedly consistent and precise depiction of the global distribution of the long-term annual mean sources and sinks of atmospheric CO<sub>2</sub>, with regional uncertainties of generally less than 0.1 Pg C yr<sup>-1</sup>. In particular, we show that these two estimates indicate a small net uptake flux of CO<sub>2</sub> in the Southern Ocean, with a magnitude that is substantially smaller than previous estimates.

## 2. Data and Methods

We provide here an abbreviated description of the ocean inversion method, the data, and how we assessed uncertainties. The reader interested in more details is referred to the underlying publications by Mikaloff Fletcher et al. [2006a]

and Mikaloff Fletcher et al. [2006b] and their respective on-line supplementary materials.

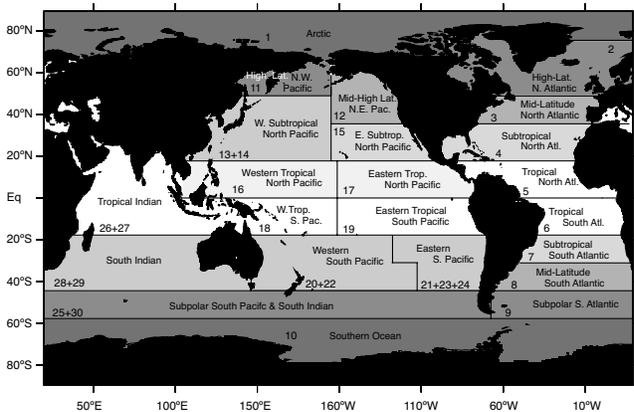
### 2.1. Ocean Inversion

The ocean inversion method is based on the premise that the ocean interior distribution of a soluble gas directly reflects the exchange of this gas across the air-sea interface, provided that it has no sources or sinks in the interior. Therefore, if ocean transport and mixing can be reversed, then regionally resolved air-sea fluxes of this gas can be inferred from ocean interior data by applying this “inverse” transport to the data [Gloor et al., 2001, 2003; Gruber et al., 2001]. In our case, this “inverse” transport is determined by releasing passive dye tracers at the surface of 30 prescribed regions in an Ocean General Circulation Model (OGCM) (Figure 1). The model is integrated forward in time, resulting in the spreading of the dye tracers from the surface into the ocean’s interior. The dye tracers are then sampled at locations and times corresponding to the space-time distribution of the observations. Due to their very different atmospheric histories, separate dye tracer simulations are undertaken for natural and for anthropogenic CO<sub>2</sub> [Gloor et al., 2003; Mikaloff Fletcher et al., 2006a, b]. In the final inversion step, the sampled dye tracers,  $\bar{c}_i$ , often referred to as basis functions, are combined linearly such that they match a set of observations,  $\bar{c}_{\text{obs}}$ , most closely, i.e., the expression

$$\left(\bar{c}_{\text{obs}} - \sum_i \lambda_i \bar{c}_i\right)^T \mathbf{cov}^{-1} \left(\bar{c}_{\text{obs}} - \sum_i \lambda_i \bar{c}_i\right), \quad (1)$$

is minimized. Here, T represents the transpose operator, the multipliers  $\lambda_i$  are the (unknown) scaling factors,  $\bar{c}_i$  are the basis function fields sampled at the observation locations, and  $\mathbf{cov}$  is the error covariance matrix of the data, used to weigh the contribution of different observations. The final air-sea flux estimates are then obtained by multiplying the scaling factors with the amount of dye tracer released in each region to simulate the basis functions.

Owing to a combination of data limitation and signal dispersion, we found that the inversion cannot reliably resolve the partitioning of the air-sea fluxes into all 30 regions. Following Mikaloff Fletcher et al. [2006b], we addressed this limitation by combining, after the inversion, those regions whose partitioning is ill constrained. We identified these



**Figure 1.** Global map of the 23 regions resolved by the ocean inversion. Numbers refer to the 30 regions, for which basis functions were originally computed [Mikaloff Fletcher et al., 2003].

regions on the basis on an analysis of the matrix of region-region flux covariances. This results in our reporting of flux estimates for 23 regions (Figure 1) (see *Mikaloff Fletcher et al.* [2006a,b] for details).

## 2.2. Data and derived tracers

Such an inverse approach has been difficult to exploit for ocean interior dissolved inorganic carbon (*DIC*) data until recently, because such observations were historically sparse and often inaccurate, and because *DIC* is not a conservative tracer in the ocean, requiring methods to separate the imprint of the air-sea flux of natural and anthropogenic CO<sub>2</sub> from the large *DIC* variations caused by the biological transformations.

The first problem has been overcome by the recent availability of a high-quality, global-scale data set of inorganic carbon parameters and related tracers [*Key et al.*, 2004], based largely on the World Ocean Circulation Experiment (WOCE)/Joint Global Ocean Flux Study (JGOFS) global CO<sub>2</sub> survey [*Wallace*, 2001], which have been archived by the Global Ocean Data Analysis Project (GLODAP). We augmented the GLODAP dataset with several historical cruises with high accuracy data, yielding over 68,000 observations with excellent spatial coverage and full depth resolution.

We have resolved the second problem by the formulation of two quasi-conservative tracers,  $\Delta C_{\text{gas ex}}$ , and  $C_{\text{ant}}$ , whose concentration variations are directly related to the air-sea exchange of natural and anthropogenic carbon, respectively [*Gruber et al.*, 1996; *Gruber and Sarmiento*, 2002]. The basic tenet underlying these tracers is the Redfield ratio concept, i.e., the observation that the biological transformations altering *DIC* are also altering several other tracers with nearly constant stoichiometric ratios [*Redfield*, 1958; *Redfield et al.*, 1963]. With this assumption, it is possible to combine *DIC*, *Alk*, and phosphate ( $\text{PO}_4^{3-}$ ) linearly in such a way that biological changes in the three components cancel each other and the resulting tracer

$$\Delta C_{\text{gas ex}} = \frac{S_o}{S} (DIC - r_{C:P} \cdot \text{PO}_4^{3-}) - \frac{S_o}{2 \cdot S} (Alk + r_{N:P} \cdot \text{PO}_4^{3-}) - C_{\text{ant}} - \text{const}, \quad (2)$$

is conserved (see *Gruber and Sarmiento* [2002] for a detailed derivation). Here  $r_{C:P}$  and  $r_{N:P}$  are stoichiometric carbon-to-phosphorus and nitrogen-to-phosphorus ratios for photosynthesis and remineralization. The constant is arbitrarily chosen such that the mean surface  $\Delta C_{\text{gas ex}}$  is zero [*Gruber and Sarmiento*, 2002]. The value of this constant has no implications for our results, as the inversion only interprets spatial gradients of  $\Delta C_{\text{gas ex}}$ . A salinity normalization ( $S/S_o$ ) is added in order to account for the concentration/dilution effect generated by evaporation and precipitation.

As there is no exchange of phosphorus and alkalinity across the air-sea interface, and since we have subtracted the anthropogenically driven increase in *DIC* ( $C_{\text{ant}}$ ), changes in  $\Delta C_{\text{gas ex}}$  must be driven by the air-sea gas exchange of natural CO<sub>2</sub>. A minor complication arises due to the input of carbon and phosphorus by rivers, which, depending on the carbon-to-phosphorus ratio, can also cause non-conservative changes in  $\Delta C_{\text{gas ex}}$  and hence must be properly accounted for in our analysis (see below). For  $C_{\text{ant}}$ , we use the estimates reported by *Key et al.* [2004], which are based on the  $\Delta C^*$  method of *Gruber et al.* [1996]. The  $\Delta C_{\text{gas ex}}$  tracer is closely related to the quasi-conservative tracer  $\Sigma\text{CO}_2^*$  of *Broecker and Peng* [1992], but it includes a more detailed consideration of  $C_{\text{ant}}$ .

Since  $\Delta C_{\text{gas ex}}$  and  $C_{\text{ant}}$  are derived quantities, we have to consider the impact of errors and biases in these tracers on our inverse flux estimates. *Matsumoto and Gruber* [2005]

reviewed and assessed the uncertainties associated with the  $\Delta C^*$ -derived  $C_{\text{ant}}$  in great detail and concluded that many of the potential biases tend to cancel so that the distribution of  $C_{\text{ant}}$  tends to be relatively robust. They identified, however, a tendency for  $C_{\text{ant}}$  to be overestimated in the upper ocean and to be underestimated in the deeper ocean, with a possible overestimation of the global inventory of about 7%. *Mikaloff Fletcher et al.* [2006a] investigated the impact of these uncertainties and biases on the inverse estimates of the air sea fluxes of anthropogenic CO<sub>2</sub> and demonstrated that the impact of the possible vertical distribution error of  $C_{\text{ant}}$  is small but that any change in the global inventory will lead to a nearly equal relative change in the global uptake flux of anthropogenic CO<sub>2</sub>. Hence, a possible 7% reduction of the global inventory of anthropogenic CO<sub>2</sub> will cause a nearly 7% reduction in the global uptake flux of anthropogenic CO<sub>2</sub> in our inversion. Without a careful reevaluation of the global inventory of anthropogenic CO<sub>2</sub>, it is premature, however, to adjust our flux estimate.

The main source of systematic error in  $\Delta C_{\text{gas ex}}$  is a possible bias in the carbon-to-phosphorus stoichiometric ratio,  $r_{C:P}$ , used to remove from *DIC* the contribution of organic matter formation and decomposition (see (1)). *Mikaloff Fletcher et al.* [2006b] investigated the impact of this uncertainty, as well as that caused by uncertainty of  $C_{\text{ant}}$ , on the inferred estimates of the air-sea flux of natural CO<sub>2</sub> and showed that the resulting differences in the estimated air-sea fluxes of natural CO<sub>2</sub> are generally smaller than the differences between the different ocean general circulation models, which we use as an estimate of the uncertainty of the inverse flux estimates.

## 2.3. Ocean transport models

In order to account for the uncertainty in ocean transport stemming from potential biases in the employed OGCMs, we use dye tracer simulations from a suite of 10 OGCMs, which span nearly the entire range of model behavior exhibited by the current generation of global-scale OGCMs used to model ocean biogeochemical processes [*Matsumoto et al.*, 2004; *Doney et al.*, 2004]. These dye-tracer simulations were undertaken by six different modeling groups: Princeton (PRINCE), Massachusetts Institute of Technology (MIT), Bern-Switzerland (Bern3D), Jet Propulsion Laboratory (ECCO), National Center for Atmospheric Research (NCAR), and University of Liège-Belgium (UL) (described in detail in *Mikaloff Fletcher et al.* [2006a]). Princeton provided results from five different configurations of their model [*Gnanadesikan et al.*, 2002, 2004]. The use of this large range of models permits a much improved assessment of model transport uncertainties compared to *Gloor et al.* [2003]. The inverse estimates from the different models were aggregated to a mean flux, whereby each model was weighted according to a skill score [*Taylor*, 2001] to account for the substantial differences in the model's ability to correctly simulate the oceanic distribution of passive tracers. For the natural CO<sub>2</sub> inversions, we used a skill score based on natural radiocarbon, while for the anthropogenic CO<sub>2</sub> inversions, we used a score based on chlorofluorocarbons (CFC). In both cases, observations were taken from GLODAP [*Key et al.*, 2004].

Despite large differences in model architecture, setup, and forcing, the employed OGCMs share certain shortcomings. This means that the weighted mean fluxes cannot necessarily be viewed as unbiased estimates. The most important shared shortcoming of all employed OGCMs is their coarse resolution, which requires the parameterization of mesoscale processes. Furthermore, all of the models use z-coordinates as their vertical discretization, leading to potential problems

when flow over steep topography has to be resolved, for example. Common problems in the resulting ocean circulation fields are (i) a too southerly formation region for North Atlantic Deep Water (NADW), which results in a too shallow southward flowing of NADW, (ii) equatorial current systems that are often only partially resolved, and (iii) a strong sensitivity of the modeled circulation fields to small changes in the parameters for subgrid-scale parameterization.

#### 2.4. River fluxes

The steady-state outgassing of natural CO<sub>2</sub> caused by the input of organic and inorganic carbon by rivers [Sarmiento and Sundquist, 1992] must be properly considered in order to be able to compare our inverse estimates of the air-sea CO<sub>2</sub> fluxes with those based on the air-sea difference of  $p\text{CO}_2$ . It appears that our inversion of  $\Delta C_{\text{gas ex}}$  only partially resolves this steady-state outgassing flux of natural CO<sub>2</sub>. This conclusion is based on our finding of a global near-zero flux of natural CO<sub>2</sub> in the absence of our imposing such a constraint. We interpret this as evidence that our natural CO<sub>2</sub> flux estimates are based on a nearly “perfect” inversion (see discussion in appendix A), i.e., we assume that the  $\Delta C_{\text{gas ex}}$  data reflect both the addition of inorganic carbon from rivers (either directly as *DIC* or by the remineralization of organic carbon), which the inversion erroneously interprets as an uptake of CO<sub>2</sub> from the atmosphere, as well as the loss of inorganic carbon due to the outgassing of this riverine CO<sub>2</sub>, which the inversion interprets correctly as an air-sea flux. This requires an adjustment of the “raw” natural CO<sub>2</sub> flux estimates to obtain a correct estimate of the global net air-sea flux of CO<sub>2</sub> (see appendix A for a more detailed discussion).

Following Jacobson *et al.* [2006a], we adopt a global total outgassing of 0.45 Pg C yr<sup>-1</sup>, which is partitioned among all regions using spatially resolved estimates of the carbon input by rivers [Ludwig *et al.*, 1996]. The magnitude of this river outgassing-induced adjustment and its spatial allocation is very uncertain, resulting in our assigning a  $\pm 50\%$  uncertainty to these flux adjustments.

### 3. Results and Discussion

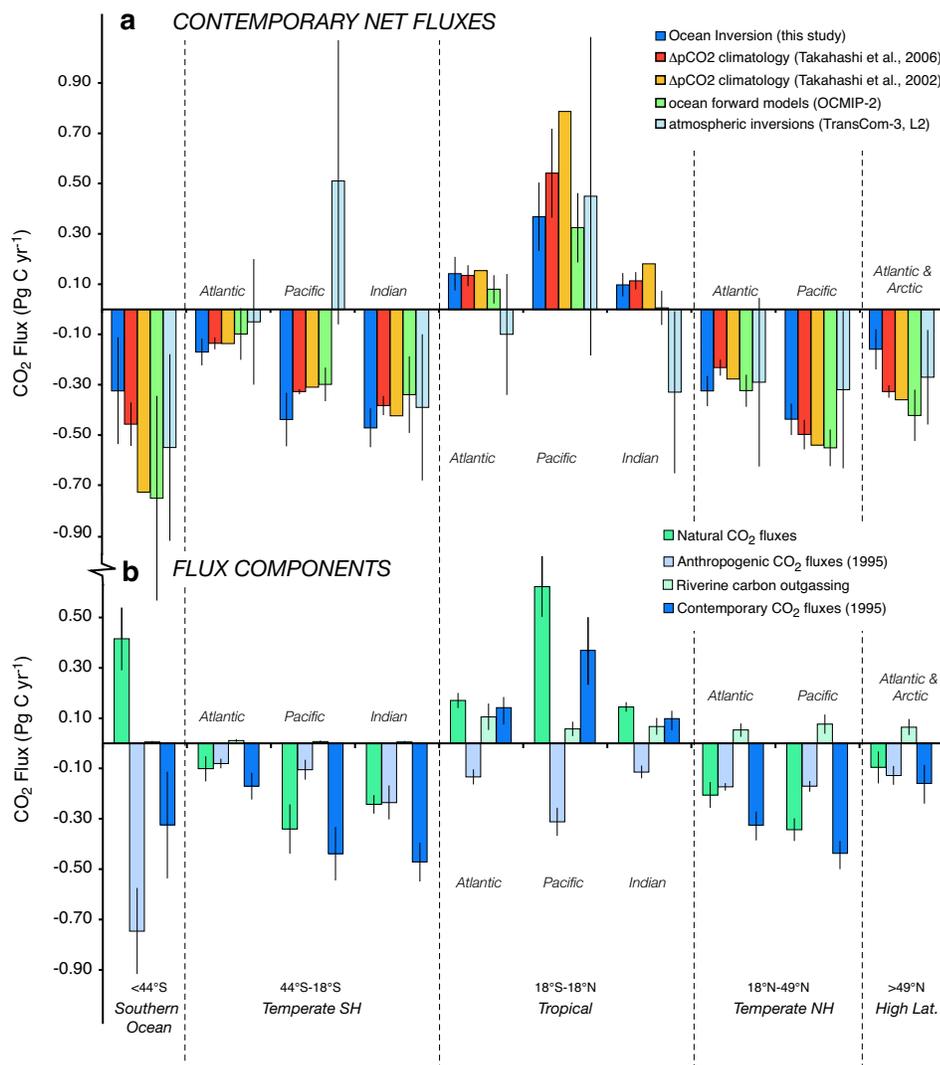
#### 3.1. Contemporary fluxes

The contemporary air-sea CO<sub>2</sub> fluxes, aggregated to 10 large-scale regions for clarity, show the familiar annual-mean pattern of outgassing of CO<sub>2</sub> in the tropical regions and uptake of atmospheric CO<sub>2</sub> nearly everywhere else (Figure 2a) (A listing of the estimated fluxes for all 23 regions and for a nominal year of 1995 is given in Table S1 in the supplementary material). The strongest outgassing is diagnosed in the tropical Pacific, where we find an outgassing of nearly 0.4 Pg C yr<sup>-1</sup>. The temperate latitude regions in the different ocean basins contribute nearly equally to the oceanic sink for atmospheric CO<sub>2</sub>, with an uptake flux between  $\sim 0.2$  and  $\sim 0.5$  Pg C yr<sup>-1</sup>. The high-latitude regions, i.e., the North Atlantic and Arctic ( $>49^\circ\text{N}$ ), and the Southern Ocean ( $<44^\circ\text{S}$ ), are comparatively weak sink regions. At the hemispheric scale, we find that the southern hemisphere extratropics ( $<18^\circ\text{S}$ ) is the largest sink region, with an uptake of  $1.4 \pm 0.2$  Pg C yr<sup>-1</sup>, mostly reflecting its large surface area. By contrast, the contemporary uptake flux of the northern hemisphere extratropics ( $>18^\circ\text{N}$ ) accounts for only  $0.9 \pm 0.1$  Pg C yr<sup>-1</sup> (Table S1). The total outgassing in the tropics ( $18^\circ\text{S}$  to  $18^\circ\text{N}$ ) amounts to  $0.6 \pm 0.2$  Pg C yr<sup>-1</sup>, resulting in a global contemporary net uptake flux of CO<sub>2</sub> of  $1.7 \pm 0.3$  Pg C yr<sup>-1</sup>. As will be discussed below in more detail, this global integral consists of an anthropogenic CO<sub>2</sub> uptake flux of  $2.2 \pm 0.2$  Pg C yr<sup>-1</sup> [Mikaloff Fletcher *et al.*, 2006a] (for a nominal year of 1995), and a natural CO<sub>2</sub> outgassing flux of  $0.5 \pm 0.2$  Pg C yr<sup>-1</sup>, with the latter almost entirely driven by the outgassing of river derived CO<sub>2</sub>.

The comparison of our inverse estimates of the contemporary air-sea CO<sub>2</sub> flux with a set of recent estimates reveals a relatively consistent pattern in the northern hemisphere, but large differences in the southern hemisphere, particularly in the Southern Ocean (Figure 2b). Included in this comparison are three very different types of flux estimates: The first two are based on compilations of oceanic  $p\text{CO}_2$  observations [Takahashi *et al.*, 2002, 2006], the third stems from simulation results by 13 ocean biogeochemistry models that participated in the 2nd phase of the Ocean Carbon-cycle Model Intercomparison Project (OCMIP-2) [Watson and Orr, 2003], and the fourth is based on a seasonal inversion of atmospheric CO<sub>2</sub> undertaken by the TransCom-3 intercomparison project [Gurney *et al.*, 2004]. Since the latter two estimates involve models to a substantial degree, we first focus our discussion on the comparison with the most recent  $p\text{CO}_2$ -based estimate of Takahashi *et al.* [2006].

This new estimate is based on a greatly expanded and improved  $p\text{CO}_2$  climatology of the surface ocean, with the most important change being the addition of nearly 2 million new observations, bringing the total to nearly 2.9 million observations (see appendix B for additional information). Contrary to earlier versions of this climatology, all observations were included, regardless of the state of the El Niño/Southern Oscillation. The largest improvement in data coverage occurred in the Southern Ocean, where the previous climatology [Takahashi *et al.*, 2002] had relatively few observations, particularly in winter time. While the central flux estimate shown in Figure 2a was computed using a square dependence of the gas transfer velocity on the wind speed [Wanninkhof, 1992], we also computed the air-sea CO<sub>2</sub> flux using the linear wind speed dependence of Krakauer *et al.* [2006] and the cubic wind speed dependence of Wanninkhof and McGillis [1999], and used the standard deviation of these three estimates as an approximate uncertainty estimate. As all three parameterizations are based on the same global mean transfer velocity (about 20 cm hr<sup>-1</sup>), this uncertainty estimate is a lower bound, as it does not include the possible error that stems from uncertainties in this mean transfer velocity (see e.g., Naegler *et al.* [2006]).

At this aggregated scale, our inversely estimated contemporary air-sea CO<sub>2</sub> flux and the new  $p\text{CO}_2$ -based estimate agree with each other to generally within less than 0.1 Pg C yr<sup>-1</sup> (Figure 2a). One exception is the tropical Pacific, where the  $p\text{CO}_2$  based outgassing estimate exceeds that of the inversion by 0.17 Pg C yr<sup>-1</sup>. Since the tropical Pacific is among the best-sampled ocean regions in terms of  $p\text{CO}_2$ , sampling biases are likely not a problem, although the observation of regional differences in the decadal trends of oceanic  $p\text{CO}_2$  [Takahashi *et al.*, 2003] could lead to some biases in the  $p\text{CO}_2$  climatology. Significant uncertainty is also associated with the magnitude of the gas transfer velocity at low windspeed, where factors other than wind can control the gas transfer (e.g., McGillis *et al.* [2004]). Perhaps the most likely possibility is that the inversion estimates are biased low, possibly owing to the OGCMs used to compute the tracer basis functions having relatively low resolution. As a result, they may share a common deficiency in their modeled upper ocean circulation in the tropics, which includes a complex, small-scale pattern of currents and countercurrents. Another exception is the northern North Atlantic and the Arctic, where the inverse uptake estimate is 0.16 Pg C yr<sup>-1</sup> smaller than the  $p\text{CO}_2$ -based flux estimate. One reason for this discrepancy is the shared problem among all OGCMs employed here of having a tendency to produce NADW too far south, with little or no formation in the Nordic Seas, and with the southward flow of this water mass in the Atlantic occurring at too shallow depths [Doney *et al.*, 2004]. In the inversion, this circulation bias tends to cause an erroneous



**Figure 2.** Air-sea CO<sub>2</sub> fluxes for 10 regions, ordered by latitude and Ocean basin (positive: outgassing; negative: uptake) (a) Comparison of contemporary air-sea fluxes. Shown are the ocean inverse estimates (this study), two estimates based on observations of the air-sea difference of the partial pressure of CO<sub>2</sub> (Takahashi et al. [2002] and Takahashi et al. [2006]), one set of estimates based on results from the 13 ocean biogeochemistry models that participated in the second phase of the Ocean Carbon-cycle Model Intercomparison Project (OCMIP-2) [Watson and Orr, 2003], and the mean estimates from the TransCom-3 project based on seasonal inversions of atmospheric CO<sub>2</sub> [Gurney et al., 2004]. The plotted uncertainty for the *p*CO<sub>2</sub>-based estimate of Takahashi et al. [2006] is the standard deviation of the computed air-sea CO<sub>2</sub> flux using a linear, a square, or a cubic dependence of the gas transfer velocity on the wind speed. The uncertainty for the OCMIP-2 estimates reflect the (unweighted) standard deviation across the 13 models, while the uncertainties for the TransCom estimates were obtained by quadrature of the within and the between model uncertainties reported by Gurney et al. [2004]. (b) Weighted mean estimates of natural, anthropogenic, river induced, and contemporary air-sea fluxes of CO<sub>2</sub> based on our ocean inversion [Mikaloff-Fletcher et al., 2006a,b]. The results are aggregated to 10 regions from the 23 regions solved for in the inversion for reasons of clarity. Error bars denote the cross-model weighted standard deviation of the mean. The anthropogenic and contemporary CO<sub>2</sub> fluxes are for a nominal year of 1995.

southward shift in the region of uptake of both natural and anthropogenic CO<sub>2</sub> in the North Atlantic, explaining why the inverse uptake flux of CO<sub>2</sub> exceeds that based on *p*CO<sub>2</sub> in the temperate North Atlantic and underestimates the uptake in the northern North Atlantic and in the Arctic. This interpretation is confirmed by the substantially smaller difference of 0.07 Pg C yr<sup>-1</sup> when the sum of the fluxes from these two regions is compared.

Despite these significant differences, the agreement between the inversely estimated contemporary fluxes and those

estimated from the most recent *p*CO<sub>2</sub> climatology is remarkable, particularly considering that they are based on entirely independent constraints and assumptions. In the case of the ocean inversion, the flux estimates are based solely on information provided by ocean interior carbon data, which are interpreted with the help of transport and mixing estimates stemming from a suite of OGCMs. In the case of the new *p*CO<sub>2</sub>-based estimates, the main constraint are the *p*CO<sub>2</sub> observations, from which the flux is estimated using assumptions about the gas transfer velocity.

As much as these two estimates agree with each other, they differ substantially in the southern hemisphere from the other three estimates shown in Figure 2. Particularly large is the difference in the Southern Ocean south of 44°S, where the inverse and the new  $p\text{CO}_2$ -based estimates indicate a sink that is only about half the size of previous estimates, or about 0.3 to 0.4 Pg C yr<sup>-1</sup>. Given these large differences in the Southern Ocean, it is incumbent on us to demonstrate the robustness of our inverse estimates. *Mikaloff Fletcher et al.* [2006a,b] and *Jacobson et al.* [2006a] undertook detailed studies to determine the uncertainties of the inversely estimated fluxes resulting from uncertainties in the data, from assumptions associated with the determination of  $\Delta C_{\text{gas ex}}$  and  $C_{\text{ant}}$ , and from errors in the OGCMs used to construct the basis functions. These studies concluded that the inversely estimated fluxes, once aggregated to the large regions presented here, are remarkably robust, and that the cross-model uncertainties reported here are a good measure of the overall uncertainty of the flux estimates.

As discussed by *Takahashi et al.* [2006], the uncertainty associated with the  $p\text{CO}_2$  based flux estimates is mostly caused by possible sampling biases and uncertainties associated with the gas transfer velocity. The use of a linear dependence of the gas transfer velocity on wind speed, as recently proposed by *Krakauer et al.* [2006], decreases the  $p\text{CO}_2$  based fluxes south of 44°S by only 0.05 Pg C yr<sup>-1</sup>. The cubic dependence of *Wanninkhof and McGillis* [1999] increases the uptake flux substantially to 0.6 Pg C yr<sup>-1</sup>. However, *Krakauer et al.* [2006] demonstrated that a cubic dependence on the windspeed is inconsistent with carbon-13 and carbon-14 constraints, a conclusion supported by the recent in-situ measurements, which suggest a quadratic dependence [*Ho et al.*, 2006]. If we exclude the cubic dependence, the uncertainty in the flux estimate caused by the gas transfer velocity may not exceed 0.1 Pg C yr<sup>-1</sup>. The sampling bias is more difficult to assess, but we note that the new  $p\text{CO}_2$  climatology has over 1 million observations south of 44°S, with reasonably good coverage in all seasons. Using the concurrently sampled temperature as a proxy, *Takahashi et al.* [2006] estimate the potential bias in the annual mean oceanic  $p\text{CO}_2$  in the Southern Ocean to be around  $\pm 2 \mu\text{atm}$ , which translates into a possible flux bias of about  $\pm 0.15$  Pg C yr<sup>-1</sup>. Taken together, the uncertainty of the  $p\text{CO}_2$ -based estimate in the Southern Ocean is estimated to be around  $\pm 0.2$  Pg C yr<sup>-1</sup>.

Given that both our inverse CO<sub>2</sub> uptake of  $0.3 \pm 0.2$  Pg C yr<sup>-1</sup> as well as the new  $p\text{CO}_2$  uptake of *Takahashi et al.* [2006] of  $0.4 \pm 0.2$  Pg C appear to be robust estimates for the Southern Ocean CO<sub>2</sub> flux south of 44°S, can we explain why the other three estimates may be biased high? In the case of the earlier  $p\text{CO}_2$  based estimate of *Takahashi et al.* [2002], the bias can be relatively easily explained by the existence of a well known summer bias in this older climatology. As the Southern Ocean is characterized by neutral to supersaturated conditions in winter time, and undersaturated conditions in summertime (see also *Metzl et al.* [2001] and *Jabaud-Jan et al.* [2004]), a climatology that includes predominantly summertime data will be biased toward a larger uptake. In the case of the OCMIP-2 models, the much lower contemporary uptake of CO<sub>2</sub> appears to be primarily driven by a weaker outgassing flux of natural CO<sub>2</sub> and not by a stronger uptake of anthropogenic CO<sub>2</sub> [*Mikaloff Fletcher et al.*, 2006b]. *Mikaloff Fletcher et al.* [2006b] argue that the outgassing flux of natural CO<sub>2</sub> could be biased low in the OCMIP-2 models owing to errors in their common biological model [*Najjar and Orr*, 1998], particularly errors associated with the seasonal cycle [*Najjar et al.*, 2006]. The TransCom-3-based uptake estimate for the Southern Ocean is in between that of *Takahashi et al.* [2002] and that from our inversion. This can be explained by the requirement in atmospheric inversions for regularization techniques (such

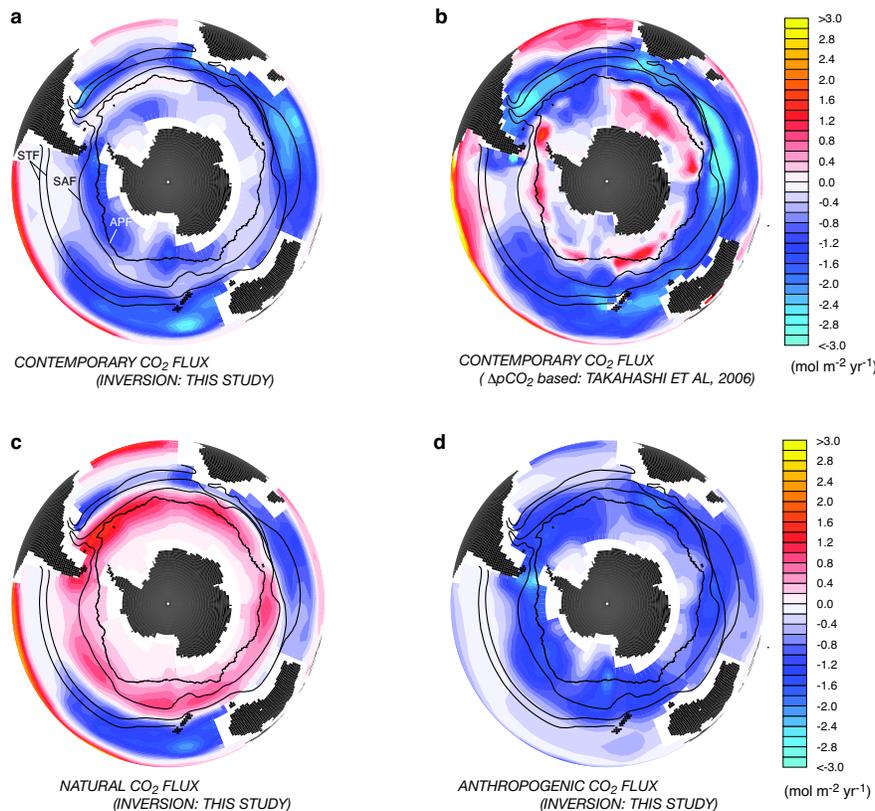
as prior constraints) in order to avoid unrealistic solutions (see, e.g., *Fan et al.* [1999]). In the case of TransCom-3, the air-sea flux estimates of *Takahashi et al.* [2002] were used as a prior constraint. Therefore, despite a strong tendency of the atmospheric inversion to reduce the Southern Ocean sink strength relative to *Takahashi et al.* [2002], any such reduction (or change from the prior) comes with a penalty. Given the limited information contained in the current atmospheric CO<sub>2</sub> network to constrain air-sea CO<sub>2</sub> fluxes (see detailed discussion by *Jacobson et al.* [2006a]), the prior constraint tends to dominate, keeping the shift small. Nevertheless, our finding here of a substantially smaller contemporary CO<sub>2</sub> sink in the Southern Ocean is a finding qualitatively consistent with all atmospheric CO<sub>2</sub> inversions [*Gurney et al.*, 2002; *Roy et al.*, 2003]. A recently developed new air-sea CO<sub>2</sub> flux estimate for the Southern Ocean based on a climatology of surface ocean DIC also supports a small contemporary CO<sub>2</sub> uptake, with a magnitude that is even smaller than ours, i.e.,  $0.1 \pm 0.3$  Pg C yr<sup>-1</sup> [*McNeil et al.*, 2005].

The excellent agreement between the two contemporary CO<sub>2</sub> flux estimates at the scale of the entire Southern Ocean breaks down when compared at more regional scales (Figure 3, panels a and b). The most distinct difference is found in the Antarctic region south of the Antarctic Polar Front (APF), where the  $p\text{CO}_2$ -based fluxes indicate an outgassing, while the inversion-based estimates show an uptake. In addition, the  $p\text{CO}_2$ -based fluxes reveal a stronger uptake in the Subantarctic Zone between the Subantarctic Front (SAF) and the Subtropical Fronts (STF). The smoother distribution of the inverse fluxes is expected, as it represents the weighted mean average of the estimates from 10 different OGCMs. In addition high covariances in the basis functions between a number of regions in the southern hemisphere required the aggregation of various regions to a larger region, thereby smoothing possible spatial gradients in the fluxes. The number of  $p\text{CO}_2$  observations south of the Polar Front, particularly in the vicinity of sea-ice, are also rather limited, making the air-sea CO<sub>2</sub> flux in this region sensitive to details of the extrapolation technique used to map the observations. Therefore, the flux estimates in the Antarctic region need to be considered more uncertain than elsewhere. In summary, while a convergence is reached for large-scale regions, the sub-regional distribution, particularly in the Southern Ocean, remains to be firmly established.

### 3.2. Attribution and Processes

Our inverse estimate of the contemporary air-sea CO<sub>2</sub> flux is constructed from two separate inversions, one for natural CO<sub>2</sub> [*Mikaloff Fletcher et al.*, 2006b] and one for anthropogenic CO<sub>2</sub> [*Mikaloff Fletcher et al.*, 2006a], permitting us to partition the contemporary flux into the two driving components. As is evident from Figure 2b, the contemporary flux arises as a result of a complex superposition of the anthropogenic and natural CO<sub>2</sub> fluxes, with the former going into the ocean everywhere, and the latter exhibiting a pattern of outgassing in the tropics and uptake in the mid-latitudes and northern high latitudes. By definition, the flux of river-derived natural CO<sub>2</sub> is out of the ocean everywhere, with a magnitude that is considerably smaller than either of the two other components.

The Southern Ocean differs from the high northern latitudes as it represents a strong source of natural CO<sub>2</sub> to the atmosphere. At the same time, this region represents the largest sink region for anthropogenic CO<sub>2</sub>. Thus, as displayed in greater detail in Figure 3 (panels c and d), the contemporary small sink in the Southern Ocean is caused by a near cancellation between a substantial outgassing flux of natural CO<sub>2</sub>, which presumably existed already in preindustrial times, and a large uptake flux of anthropogenic CO<sub>2</sub>,



**Figure 3.** Polar stereographic maps of air-sea CO<sub>2</sub> fluxes (mol m<sup>-2</sup> yr<sup>-1</sup>) (positive (red): outgassing, negative (blue): uptake). (a) Inverse estimate of the contemporary air-sea flux (nominal 1995), i.e., the sum of natural and anthropogenic CO<sub>2</sub>. (b) Oceanic pCO<sub>2</sub>-based estimate of the air-sea CO<sub>2</sub> flux based on Takahashi et al. [2006] (nominal year of 2000). (c) Inverse estimates of the air-sea flux of natural CO<sub>2</sub>, which is equivalent to the air-sea flux of CO<sub>2</sub> in preindustrial times. (d) Inverse estimate of the air-sea flux of anthropogenic CO<sub>2</sub> for a nominal year of 1995. The lines show the dominant fronts in the Southern Ocean, going equatorward from Antarctica: Antarctic Polar Front (APF), Subantarctic Front (SAF), and Southern and Northern Subtropical Fronts (STF). The maps for the inverse estimates were constructed by distributing the inverse flux estimates within each region according to the flux pattern that was used when releasing the dye tracers from the surface.

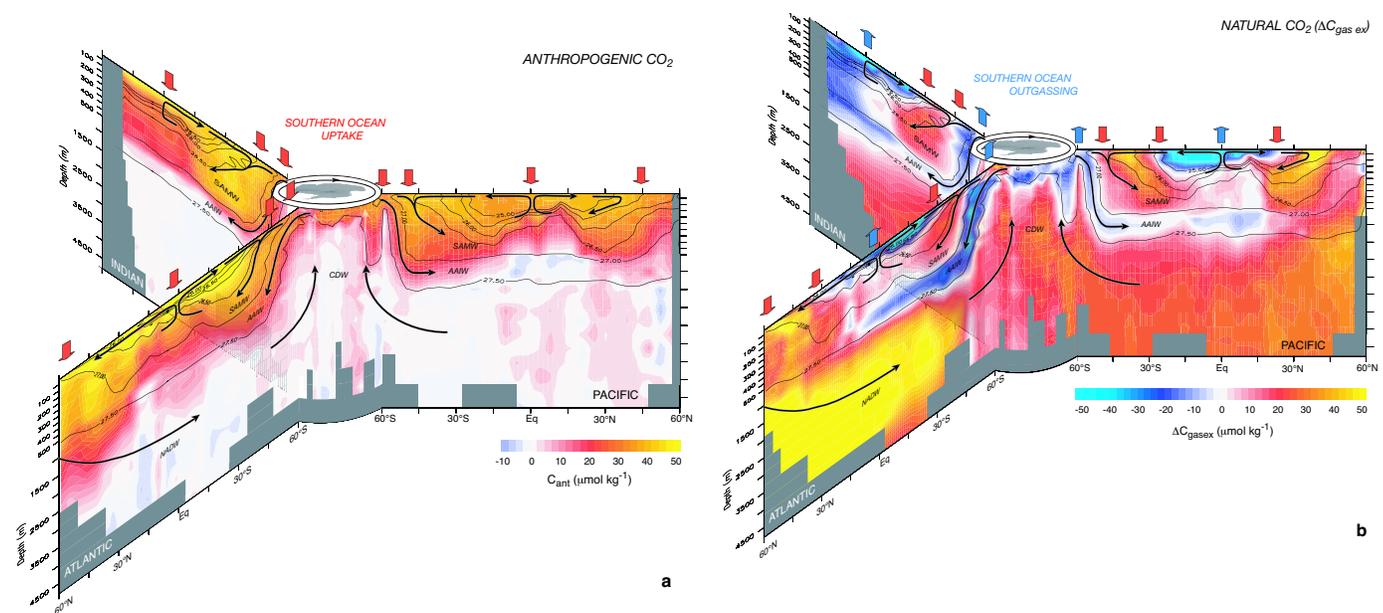
which has grown over time and is expected to increase further as atmospheric CO<sub>2</sub> continues to rise. Presently, the Southern Ocean takes up more than a third of the global anthropogenic CO<sub>2</sub>, yet it covers only 18% of the global surface ocean area.

High uptake rates of anthropogenic CO<sub>2</sub> are associated with those regions where waters that have not been in contact with the atmosphere for some time come to the surface and are exposed to the increased CO<sub>2</sub> in the atmosphere [Sarmiento et al., 1992; Orr et al., 2001; Mikaloff Fletcher et al., 2006a]. In the Southern Ocean, the upwelling of Circumpolar Deep Water (CDW) with very low concentrations of  $C_{\text{ant}}$  provides a large potential for the uptake of anthropogenic CO<sub>2</sub> (Figure 4a).  $C_{\text{ant}}$  is low in upwelling CDW because it stems in part from the mid-depth return flow of deep waters from the Indian and Pacific Ocean [Schmitz, 1995]. Most of this uptake potential is realized as the upwelled waters that are being pushed northward at the surface by the Ekman drift are exposed to the elevated atmospheric CO<sub>2</sub> in the presence of high windspeeds, which accelerate the uptake. The resulting anthropogenic CO<sub>2</sub>-rich waters are then transported to depth as Antarctic Intermediate Waters (AAIW) and Sub-Antarctic Mode Waters (SAMW) discernible in Figure 4a by the deep penetration of anthropogenic CO<sub>2</sub> at mid-latitudes of the Southern hemisphere (see also Gruber [1998] and Sabine et al. [2004]). Globally, this is the single most important pathway for the oceanic uptake of anthropogenic CO<sub>2</sub>, causing a circumpolar ring of

strong uptake flux centered in the Polar Frontal Zone (between the Polar Front and the Subantarctic Front) and in the Subantarctic Zone (Figure 3d).

W

The source and sink pattern of natural CO<sub>2</sub> inferred by the inversion can be traced back to the ocean interior distribution of  $\Delta C_{\text{gas ex}}$ , in particular its gradients (Figure 4b). Waters that have low concentrations of  $\Delta C_{\text{gas ex}}$  must have lost natural CO<sub>2</sub> to the atmosphere the last time they were in contact with the atmosphere, and vice versa for waters with high  $\Delta C_{\text{gas ex}}$  concentrations. Closer inspection of the  $\Delta C_{\text{gas ex}}$  distribution in the Southern Ocean reveals that the CDW upwelling in this region is characterized by elevated  $\Delta C_{\text{gas ex}}$  concentrations, whereas the  $\Delta C_{\text{gas ex}}$  concentrations in the core of the AAIW ( $\sigma_{\theta}$  between 27.00 and 27.50) are substantially depleted, reflecting the strong outgassing of natural CO<sub>2</sub> as waters flow northward in the Southern Ocean (Figure 4b). By contrast, SAMW has rather high  $\Delta C_{\text{gas ex}}$  concentrations, reflecting the uptake of natural CO<sub>2</sub> as some of the waters from the south continue to be pushed northward. Some of this high  $\Delta C_{\text{gas ex}}$ , particularly in the water masses shallower than SAMW, may also reflect the entrainment of southward flowing waters that took up natural CO<sub>2</sub> from the atmosphere during their transit in the temperate latitudes of the southern hemisphere (Figure 3b). Also discernible in Figure 4b is the high  $\Delta C_{\text{gas ex}}$  tongue of southward flowing North Atlantic Deep Water (NADW) in



**Figure 4.** Ocean interior distributions of the tracers reflecting the exchange of CO<sub>2</sub> across the air-sea interface, displayed as global-scale section plots organized around the Southern Ocean in the center. (a) Distribution of anthropogenic CO<sub>2</sub>,  $C_{\text{ant}}$ , estimated from measurements of dissolved inorganic carbon and other tracers using the  $\Delta C^*$  method of Gruber *et al.* [1996]. (b) Distribution of the gas exchange component of natural CO<sub>2</sub>,  $\Delta C_{\text{gas ex}}$ , which is estimated from measurements of dissolved inorganic carbon by subtracting from it a component that accounts for the oceanic cycling of carbon induced by the biological pump, and the data-based estimates of  $C_{\text{ant}}$  shown in (a). The two tracers  $C_{\text{ant}}$  and  $\Delta C_{\text{gas ex}}$  are quasi-conservative in the sense that their distribution can only be changed by air-sea fluxes of anthropogenic and natural CO<sub>2</sub>, respectively, and ocean mixing. The inversion interprets these distributions by determining, given ocean circulation and mixing, a set of surface ocean fluxes that most closely matches these observations. Also shown are isolines of potential density anomalies,  $\sigma_\theta$  (density referenced to the ocean surface minus 1000 kg m<sup>-3</sup>), along which most of the oceanic flow occurs. Major ocean circulation features are indicated by schematic arrows. Based on data from GLODAP [Key *et al.*, 2004]. NADW: North Atlantic Deep Water, CDW: Circumpolar Deep Water; SAMW: Sub-Antarctic Mode Water; AAIW: Antarctic Intermediate Water.

the deeper parts of the Atlantic, which is the primary cause for the existence of a substantial interhemispheric transport of natural carbon [Mikaloff Fletcher *et al.*, 2006b]. However, the magnitude of this inversion-inferred southward transport across the equator is only about  $0.34 \pm 0.02$  Pg C yr<sup>-1</sup>, much smaller than originally proposed by Keeling *et al.* [1989] and estimated by Broecker and Peng [1992]. What drives this exchange of natural CO<sub>2</sub> across the air-sea interface? A net air-sea flux of natural CO<sub>2</sub> occurs whenever the surface ocean  $p\text{CO}_2$  differs from that of the atmosphere. Thus, any mechanism elevating oceanic  $p\text{CO}_2$  tends to drive out-gassing from the oceans and vice-versa. The dominant mechanisms that can alter surface ocean  $p\text{CO}_2$  are heat fluxes that affect the solubility of CO<sub>2</sub> (solubility pump) and the interaction of biological transformations in the ocean with ocean circulation and mixing, which affect the concentrations of DIC and alkalinity (*Alk*). Freshwater fluxes can also alter  $p\text{CO}_2$ , mainly due to their dilution effect on DIC and *Alk*, but are generally less important (see Keeling *et al.* [2004] and Dore *et al.* [2003] for exceptions). For the biologically-driven mechanism (biological pump) it is the balance between the physical supply of DIC and *Alk* and the biological removal of these tracers that sets the final  $p\text{CO}_2$ , and hence the direction of the air-sea CO<sub>2</sub> flux. A good measure of this balance is the efficiency of the biolog-

ical pump [Sarmiento and Gruber, 2006], i.e., the ability of phytoplankton to utilize all of the available surface nutrients in the presence of a physical resupply of nutrients and DIC from below. The CO<sub>2</sub> outgassing in the tropics and in the Southern Ocean is driven in large part by the transport of DIC-laden interior ocean waters to the surface, where DIC is then only partially depleted by the biological pump as a result of it being inefficient (see also Murnane *et al.* [1999]). This biological-pump-driven outgassing of CO<sub>2</sub> is particularly strong in the Southern Ocean, as deep waters that have accumulated a large amount of remineralized DIC are being upwelled there. In the tropics, heating of the surface waters contributes to the outgassing. In contrast, the temperate latitudes and the North Atlantic are regions of net cooling, causing an uptake of CO<sub>2</sub> from the atmosphere by the solubility pump. These regions also tend to have a very efficient biological pump, which enhances the solubility-pump-induced uptake of natural CO<sub>2</sub> (see also discussion in Broecker and Peng [1992] and Gruber and Sarmiento [2002]).

#### 4. Implications

Our estimates of the anthropogenic and contemporary air-sea CO<sub>2</sub> fluxes have several important consequences for

our understanding of the global carbon cycle and its anthropogenic perturbation. First, we are able now, with more confidence than previously possible, to quantify the global oceanic uptake of anthropogenic CO<sub>2</sub> and its spatial distribution. This provides new and tighter constraints on the global budget of anthropogenic CO<sub>2</sub>. Second, better quantitative estimates of the contemporary air-sea CO<sub>2</sub> fluxes can support the interpretation of atmospheric CO<sub>2</sub> data, particularly with regard to the estimation of the net fluxes with the terrestrial biosphere [Jacobson *et al.*, 2006a, b]. Third, a better understanding of the air-sea CO<sub>2</sub> fluxes helps to improve our ability to project the impact of future climate change on the air-sea CO<sub>2</sub> fluxes, as well as to improve our ability to explain past changes. We discuss these three implications in turn.

#### 4.1. Oceanic uptake of anthropogenic CO<sub>2</sub>

The inversion of the oceanic distribution of the reconstructed anthropogenic CO<sub>2</sub> yields a global uptake flux of anthropogenic CO<sub>2</sub> of  $2.20 \pm 0.25$  Pg C yr<sup>-1</sup> ( $\pm 1$  standard deviation) for a nominal year of 1995 [Mikaloff Fletcher *et al.*, 2006a]. This estimate compares well with the most recent set of independent estimates of the global ocean uptake of anthropogenic CO<sub>2</sub> (Table 2), after adjustments have been applied to account for the river outgassing. We adopted the same global adjustment of  $0.45$  Pg C yr<sup>-1</sup> we employed above for “raw” inverse estimates, and adjusted the estimates based on *p*CO<sub>2</sub> observations as well as those based on the inversion of atmospheric CO<sub>2</sub> observations. The magnitude of this adjustment is rather uncertain, as both the actual magnitude of the river outgassing as well as the fraction of that outgassing actually included in the *p*CO<sub>2</sub> climatology is not well known. This adds a substantial uncertainty to the *p*CO<sub>2</sub> and atmospheric CO<sub>2</sub>-based estimates of the oceanic uptake of anthropogenic CO<sub>2</sub>.

Although not significantly different from the other estimates in Table 2, our ocean inverse estimate of  $2.2 \pm 0.25$  Pg C yr<sup>-1</sup> stands out due to its small uncertainty. It compares exceptionally well with the most recent uptake estimate based on the simultaneous measurements of CO<sub>2</sub> and of the O<sub>2</sub>/N<sub>2</sub> ratio in the atmosphere [Manning and Keeling, 2006]. This latter estimate includes a  $0.5$  Pg C yr<sup>-1</sup> correction for the effect of the warming-induced outgassing of O<sub>2</sub> and N<sub>2</sub>

on the atmospheric O<sub>2</sub>/N<sub>2</sub> ratio, whose magnitude represents a major source of uncertainty in this method [Plattner *et al.*, 2002; Bopp *et al.*, 2002; LeQuéré *et al.*, 2003]. This agreement is very encouraging, as the atmospheric O<sub>2</sub>/N<sub>2</sub> method is perhaps currently the best method to constrain the oceanic sink for anthropogenic CO<sub>2</sub> on a continuing basis, permitting scientists to detect changes in this sink fast and reliably.

Our inverse estimate of the oceanic uptake of anthropogenic CO<sub>2</sub> confirms also an earlier finding by Matsumoto *et al.* [2004] that the majority of the models that participated in OCMIP-2 tend to overestimate the oceanic uptake of anthropogenic CO<sub>2</sub>. In fact, when Matsumoto *et al.* [2004] selected only those 4 models that are in agreement with a number of basic oceanic chlorofluorocarbon and bomb radiocarbon constraints, they found an oceanic uptake of anthropogenic CO<sub>2</sub> that is indistinguishable from ours (Table 2).

#### 4.2. Global budget and land biosphere

The convergence of several independent estimates of the global oceanic uptake flux of anthropogenic CO<sub>2</sub> evident in Table 2 provides new constraints on the global redistribution of anthropogenic CO<sub>2</sub> during the decade of the 1990s. When our decadal ocean uptake estimate for the 1990s of  $22 \pm 3$  Pg C is combined with the accumulation of CO<sub>2</sub> in the atmosphere ( $32 \pm 1$  Pg C) and with the total emissions from fossil fuel and cement production ( $63 \pm 4$  Pg C), a net terrestrial sink of  $9 \pm 5$  Pg C emerges for the decade of the 1990s (Table 1). This is not significantly different than earlier budgets (e.g., Sabine *et al.* [2004],  $15 \pm 9$  Pg C), but the uncertainty has been nearly halved.

An important caveat in the discussion of these budgets is the assumption that the global air-sea balance of natural CO<sub>2</sub> has remained constant over this decade, i.e., that no anomalous net flux of natural CO<sub>2</sub> has occurred in response to global climate change or other anthropogenic perturbations (see Keeling [2005], Sabine and Gruber [2005], and Matsumoto and Gruber [2005] for a discussion). In addition, because of our use of OGCMs with climatological mean circulation, the inverse calculations also make the implicit assumption that the anthropogenic CO<sub>2</sub> uptake only varies in response to the change in atmospheric CO<sub>2</sub> and is not affected by interannual to decadal fluctuation in ocean circulation. By considering decadal mean budgets, we assume that we have removed much of the year-to-year variability that may exist in the oceanic uptake of anthropogenic CO<sub>2</sub>, but the impact of long-term changes, such as the observed ocean warming [Levitus *et al.*, 2000, 2005], on the air-sea balance of natural CO<sub>2</sub> is currently not well established. Keeling [2005] suggested that ocean warming and the resulting increase in oceanic stratification has led to an outgassing of natural CO<sub>2</sub> of about  $7 \pm 10$  Pg C since 1860. It is reasonable to expect that a substantial fraction of this outgassing occurred since the early 1970s when global surface temperature began to rise quickly [Jones and Moberg, 2003]. Even so, the expected outgassing of natural CO<sub>2</sub> for the 1990s may amount to not more than perhaps 2 Pg C, which is less than 10% of the total anthropogenic CO<sub>2</sub> uptake over this period (Table 1). If we accounted for this outgassing in our budget, we would have to increase the net terrestrial balance for the 1990s accordingly to  $-11 \pm 5$  Pg C.

This net terrestrial sink consists of two opposing fluxes: A loss (outgassing) of CO<sub>2</sub> due to land-use change and a residual flux (uptake) that is assumed to occur outside the regions impacted by land-use change. For the land-use change emissions, we adopt here the mean estimate of the recent studies by Houghton [2003b], DeFries *et al.* [2002], and Achard *et al.* [2002] of 16 Pg C and use their range (8 to 28 Pg C) as an

**Table 1.** Anthropogenic CO<sub>2</sub> budget for the 1990s.

Processes	1990-1999 (Pg C)
<i>Constrained sources and sinks</i>	
(1) Emissions from fossil fuel and cement production <sup>a</sup>	$63 \pm 4$
(2) Storage in atmosphere <sup>b</sup>	$-32 \pm 1$
(3) Oceanic uptake <sup>c</sup>	$-22 \pm 3$
<i>Inferred net terrestrial balance</i>	
(4) Net terrestrial balance = [-(1)-(2)-(3)]	$-9 \pm 5$
<i>Terrestrial balance</i>	
(5) Emissions from land use change <sup>d</sup>	16 (8 to 28)
(6) Terrestrial biosphere sink = [-(1)-(2)-(3)]-(5)	$-25$ (-17 to -37)

<sup>a</sup> after Marland *et al.* [2006]

<sup>b</sup> after Keeling and Whorf [2006]

<sup>c</sup> this study and Mikaloff Fletcher *et al.* [2006a], scaled to the entire decade of the 1990s.

<sup>d</sup> average of estimates by Houghton [2003b], DeFries *et al.* [2002], and Achard *et al.* [2002].

**Table 2.** Summary of recent estimates of the oceanic uptake rate of anthropogenic CO<sub>2</sub> for the period of the 1990s and early 2000s.

Method	Estimate (Pg C yr <sup>-1</sup> )	Time-period	Authors
<i>Estimates based on oceanic observations</i>			
Ocean Inversion (10 models)	2.20 ± 0.25	Nominal 1995	this study and Mikaloff Fletcher <i>et al.</i> [2006a]
Ocean Inversion (3 models)	1.80 ± 0.40	Nominal 1990	Gloor <i>et al.</i> [2003]
Air-sea pCO <sub>2</sub> difference (adjusted) <sup>a</sup>	2.0 ± 50%	Nominal 2000 <sup>b</sup>	Takahashi <i>et al.</i> [2006]
Air-sea pCO <sub>2</sub> difference (adjusted) <sup>a</sup>	2.0 ± 60%	Nominal 1995	Takahashi <i>et al.</i> [2002]
<i>Estimates based on atmospheric observations</i>			
Atmospheric O <sub>2</sub> /N <sub>2</sub> ratio	1.9 ± 0.6	1990-1999	Manning and Keeling [2006]
Atmospheric O <sub>2</sub> /N <sub>2</sub> ratio	2.2 ± 0.6	1993-2003	Manning and Keeling [2006]
Atmospheric CO <sub>2</sub> inversions (adjusted) <sup>a</sup>	1.8 ± 1.0	1992-1996	Gurney <i>et al.</i> [2004]
<i>Estimates based on oceanic and atmospheric observations</i>			
Air-sea <sup>13</sup> C disequilibrium	1.5 ± 0.9	1985-1995	Gruber and Keeling [2001]
Joint atmosphere-ocean inversion	2.1 ± 0.2	1992-1996	Jacobson <i>et al.</i> [2006b]
<i>Estimates based on ocean biogeochemistry models</i>			
OCMIP-2 (13 models)	2.4 ± 0.5	1990-1999	Watson and Orr [2003]
OCMIP-2 (4 “best” models) <sup>d</sup>	2.2 ± 0.2	1990-1999	Matsumoto <i>et al.</i> [2004]

<sup>a</sup> Adjusted by 0.45 Pg C yr<sup>-1</sup> to account for the outgassing of natural CO<sub>2</sub> that is driven by the carbon input by rivers.

<sup>b</sup> The estimate for a nominal year of 1995 would be less than 0.1 Pg C yr<sup>-1</sup> smaller.

<sup>c</sup> Corrected for wrong windspeeds used in published version, see

[http://www.ideo.columbia.edu/res/pi/CO2/carbondioxide/pages/air\\_sea\\_flux\\_rev1.html](http://www.ideo.columbia.edu/res/pi/CO2/carbondioxide/pages/air_sea_flux_rev1.html).

<sup>d</sup> These models were selected on the basis of their ability to simulate correctly, within the uncertainty of the data, the observed oceanic inventories and regional distributions of chlorofluorocarbon and bomb radiocarbon.

uncertainty estimate. This results in an inferred biospheric carbon sink of 25 Pg C for the decade of the 1990s (range 17 to 37 Pg C), or 2.5 Pg C yr<sup>-1</sup>. The mechanisms driving this inferred uptake are not well understood and are currently the subject of intense scientific debate (e.g., Schimel *et al.* [2001]).

Inversions of atmospheric CO<sub>2</sub> have played a major role in pinpointing the possible sink regions, thereby providing information about possible mechanisms [Tans *et al.*, 1990; Schimel *et al.*, 2001; Houghton, 2003a]. However, the rapid mixing of the atmosphere together with a relatively limited number of atmospheric observing stations makes the spatial attribution of source and sink regions at the Earth’s surface on the basis of such atmospheric inversions difficult and subject to methodological discrepancies [Gloor *et al.*, 2000]. Nevertheless, there is strong consensus that the northern hemisphere land biosphere acted as a major sink region in the 1990s [Gurney *et al.*, 2002, 2003, 2004; Jacobson *et al.*, 2006b]. Less clear is the role of the tropics and of the southern hemisphere land regions, where the atmospheric CO<sub>2</sub> constraints are relatively weak, providing an opportunity for oceanic constraints to improve the atmospheric CO<sub>2</sub> inversion solutions.

Jacobson *et al.* [2006a] followed this path and developed a joint atmosphere-ocean inversion approach, in which information about the air-sea CO<sub>2</sub> fluxes inferred from an ocean inversion very similar to that reported here is used simultaneously with an atmospheric CO<sub>2</sub> inversion to improve surface flux estimates, particularly over land. Relative to inversions of atmospheric CO<sub>2</sub> alone, Jacobson *et al.* [2006b] identify the tropical and southern hemisphere land regions to be a much larger net source of CO<sub>2</sub>, with a magnitude as large as the most recent emission estimates from land-use change in those regions. This implies little CO<sub>2</sub> uptake in unperturbed tropical land regions.

### 4.3. Processes

An improved quantitative description of the spatial distribution of the oceanic sources and sinks of atmospheric CO<sub>2</sub> as well as its attribution to the exchange of natural and anthropogenic CO<sub>2</sub> helps to identify regions in the ocean that are critical in determining the past, present, and future net air-sea balance of CO<sub>2</sub>. The Southern Ocean stands out

with its large uptake of anthropogenic CO<sub>2</sub> and with its substantial outgassing of natural CO<sub>2</sub>. Simulations show that this region reacts with relatively high sensitivity to climate variations over the last 50 years [Lovenduski *et al.*, in prep.], as well as to future climate change, due to the interaction of sea-ice, upwelling, and convection with changes in heat and freshwater fluxes [Sarmiento *et al.*, 1998]. Given the large exchange fluxes of the two CO<sub>2</sub> components, relatively small changes in the Southern Ocean can lead to large changes in the net atmosphere-ocean balance of CO<sub>2</sub>, providing for substantial feedbacks in the climate system. Feedbacks between the physical climate system and the Southern Ocean carbon cycle have also been suggested to explain the much lower atmospheric CO<sub>2</sub> levels during the last few periods of maximum glacial extents [Toggweiler, 1999; Sigman and Boyle, 2000; Gildor *et al.*, 2002].

## 5. Outlook

While our results here present a major step forward in our ability to constrain the oceanic sources and sinks of atmospheric CO<sub>2</sub>, our analysis is by design limited to the long-term mean state and, by itself, does not provide any insights on how strongly and how fast the ocean carbon cycle will respond to changes in external forcing. This represents the next big challenge of ocean carbon cycle research and requires a continuing commitment to observe the time-evolving ocean carbon cycle as well as detailed modeling and process studies.

### Acknowledgments.

We are deeply indebted to the numerous scientists and technicians involved with the sampling, processing, and quality controlling of the oceanic observations underlying this research. We are also thankful to the funding agencies, both within the U.S. as well as in many other countries who supported the collection, processing, and archiving of these data. Core financial support for this study came from the National Aeronautics and Space Administration under grant NAG5-12528 to NG and SMF, with additional support by the U.S. National Science Foundation. MG, SM, FJ, and AM thank the European Commission for support through CarboOcean (511176-2) and the NOCES project (EVK2-CT-2001-00134). TT has been supported by NOAA grant NAO30AR4320179P27.

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# Oceanic sources and sinks of atmospheric CO<sub>2</sub>: Supplementary material

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## Appendix A: River fluxes

In steady-state, that part of the input of inorganic and organic carbon into the ocean by rivers that escapes burial is released back into the atmosphere as a flux of CO<sub>2</sub> across the air-sea interface [Sarmiento and Sundquist, 1992]. While the burial rate of carbon on the seafloor of the deeper ocean is reasonably well established ( $\sim 0.1$  Pg C yr<sup>-1</sup> as organic carbon and  $\sim 0.1$  Pg C yr<sup>-1</sup> as CaCO<sub>3</sub>), the burial in shallow sediments as well as the net input of carbon by rivers beyond the river mouth is poorly established. One reason is that most transport estimates of riverine carbon pertain to a location far upstream of the river mouth, and therefore do not include the myriad of transformation processes that occur in the estuaries and in the very nearshore environments. The most commonly adopted estimate for the river input of carbon is 0.4 Pg C yr<sup>-1</sup> in the form of organic carbon, and 0.4 Pg C yr<sup>-1</sup> as inorganic carbon [Sarmiento and Sundquist, 1992]. Assuming a burial flux of 0.2 Pg C yr<sup>-1</sup>, requires a CO<sub>2</sub> outgassing of 0.6 Pg C yr<sup>-1</sup> in steady-state. More recent analyses suggest a much larger input of organic carbon, perhaps as large as 1 Pg C yr<sup>-1</sup> [Richey, 2004], although it is unclear how much of this flux makes it past the estuary. Nor is it known to which extent current river input estimates reflect anthropogenically perturbed systems or pre-industrial conditions.

As the ultimate aim of the inversion is to estimate the contemporary CO<sub>2</sub> flux across the air-sea interface, we have to evaluate whether our inversely based estimates include this steady-state outgassing of river-derived carbon,

or whether we need to apply corrections. The answer to this question depends on two independent aspects: First, in what respect does the  $\Delta C_{\text{gas ex}}$  tracer reflect the gain of (inorganic) carbon from rivers and its subsequent loss by air-sea exchange? Second, what fraction of the global outgassing of river-derived carbon is actually reflected in the data set employed in our study, whose sampling locations are predominantly in the open ocean?

The answer to the first question depends critically on the carbon-to-phosphorus ratio in both the inorganic and organic fractions of the river supply (see Eq. (2)). For clarity, we discuss separately how the inversion attributes the organic and inorganic carbon that is added to the ocean by rivers.

With regard to the input of dissolved inorganic matter, rivers tend to have a very low ratio of phosphate to DIC [Meybeck, 1993]. As a result, DIC added to the ocean via rivers is reflected on a nearly mol to mol basis as a gain in  $\Delta C_{\text{gas ex}}$ . When that river derived CO<sub>2</sub> is eventually outgassed to the atmosphere, there is a corresponding decrease in  $\Delta C_{\text{gas ex}}$ . In a perfectly sampled ocean, a perfect inversion, i.e. an inversion free from systematic biases, will infer from this increase of  $\Delta C_{\text{gas ex}}$  in coastal regions an uptake of CO<sub>2</sub> from the atmosphere, and will infer from the decrease in  $\Delta C_{\text{gas ex}}$  in the open ocean an outgassing of CO<sub>2</sub>, with the two fluxes balancing each other.

With regard to organic carbon, there are similarities and differences. As is the case for the river input of dissolved inorganic matter, the phosphorus-to-carbon ratio in dissolved organic matter is much smaller than the canonical stoichiometric ratio of marine organic matter (Redfield ratio). Thus, the remineralization of this organic carbon leads to an increase in  $\Delta C_{\text{gas ex}}$  on a nearly mol to mol basis, which is then interpreted by the inversion as an uptake of CO<sub>2</sub> from the atmosphere. This organic carbon derived CO<sub>2</sub> will eventually outgas, which will be reflected in  $\Delta C_{\text{gas ex}}$  as a decrease, and hence correctly attributed in the inversion to a sea-to-air flux. Thus, analogous to inorganic carbon, the air-sea fluxes estimated by a perfect inversion of a perfectly sampled ocean will balance globally. There is an important difference to the river input of inorganic carbon, however, as organic carbon changes  $\Delta C_{\text{gas ex}}$  only at the location where it is remineralized. Thus, the riverine organic carbon signal is not attributed to an air-to-sea flux in the region where the rivers enter the oceans, but rather attributed to the ocean region where the organic carbon is remineralized.

In summary, the oceanic inversion of  $\Delta C_{\text{gas ex}}$  tends to find a globally balanced flux even in the presence of a steady-state outgassing of riverine carbon. This balance emerges because the inversion incorrectly interprets the addition of carbon by rivers as an air-to-sea flux, while it correctly determines the sea-to-air flux associated with the outgassing of the riverine carbon. Therefore, we need to subtract from the “raw” inversion estimates the riverine carbon signal that was incorrectly attributed by the ocean inversion to an air-to-sea flux.

To achieve this we add to our “raw” inverse fluxes a regionally specific estimate of the net riverine carbon input (the total input of river carbon minus the carbon that gets

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buried on the seafloor [Jacobson *et al.*, 2006]). Specifically, we adopt an estimate of  $0.45 \text{ Pg C yr}^{-1}$  for the global total outgassing of riverine carbon based on the analysis of Jacobson *et al.* [2006] and distribute this flux regionally on the basis of the spatially resolved GEM-CO<sub>2</sub> product, which is based on the work of Ludwig *et al.* [1996] (see Jacobson *et al.* [2006] for further details). The magnitude of the global adjustment as well as its regionalization is uncertain, so that we assign an uncertainty of  $\pm 50\%$  to these riverine carbon fluxes (see Table 1).

The answer to the second question, i.e. whether our network is actually reflecting the input of riverine carbon and its subsequent outgassing, turns out to be less important, despite the fact that we are using primarily an open ocean network. This is because the inversion misses not only the added inorganic river carbon signal in the case of lacking data in coastal regions, but also the resulting ocean outgassing signal. As a result, the lack of coastal data causes no imbalances in the inverted signals. However, there is a reduction of the river flux subtraction we need to apply to obtain net-air sea fluxes (which is taken into account by Jacobson *et al.* [2006]).

## Appendix B: $p\text{CO}_2$ Climatology

The methods used to construct the new  $p\text{CO}_2$  climatology of Takahashi *et al.* [2006] as well as those used to compute the net air-sea CO<sub>2</sub> flux are the same as described in Takahashi *et al.* [1995] and Takahashi *et al.* [2002] with a few exceptions. First, the total number of  $p\text{CO}_2$  observations used in the 2006 climatology is about 2.88 million, which is 3 times larger than the 0.94 million used in the 2002 climatology. The largest improvement occurred in the southern hemisphere, in particular south of  $50^\circ\text{S}$ , where very few winter time data were available for the 2002 climatology. Second, all oceanic  $p\text{CO}_2$  observations were corrected to a reference year of 2000 using the same rate of increase of  $1.5 \mu\text{atm yr}^{-1}$ . This change contributed to the significant reduction in the magnitude of the sea-air  $p\text{CO}_2$  differences and hence the magnitude of the CO<sub>2</sub> sink flux over the Southern Ocean. Third, a new parameterization of wintertime  $p\text{CO}_2$  in ice-fields was used. Forth, all observations are included in the particular version used here, while the 2002 climatology excluded  $p\text{CO}_2$  observations in the tropical Pacific that were collected during El Niño episodes.

The net air-sea flux was computed from the monthly  $p\text{CO}_2$  climatology using the climatological mean monthly values for wind speed (10 meters above surface) and ice cover percentage estimated on the basis of the NCEP Reanalysis 2 data (2006) (taken from <http://www.cdc.noaa.gov>). The areas where less than 20% of the ocean is covered with ice are assumed to be ice-free; and those with greater than 70% ice cover are assumed to be closed for sea-air exchange. The flux over areas with 20% to 70% ice cover is reduced proportional to ice cover. As Takahashi *et al.* [2002], a square

dependence of the gas transfer velocity on the windspeed was assumed for the central flux estimate. To assess the uncertainty stemming from the ill known gas transfer velocity, we additionally used the linear windspeed dependence of Krakauer *et al.* [2006] and the cubic relationship of Wanninkhof and McGillis [1999].

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**Table 1.** Inverse estimates of the air-sea fluxes of natural, anthropogenic, and river derived CO<sub>2</sub> for the 23 regions resolved by the ocean inversion (see Figure 1 in main text). Also listed is the sum of the three components, i.e., the contemporary CO<sub>2</sub> flux, in comparison with the estimates based on the pCO<sub>2</sub> climatology of *Takahashi et al.* [2006]. Positive fluxes indicate outgassing.

Region Name	Region Number <sup>a</sup>	Area (m <sup>2</sup> )	Inv. Natural CO <sub>2</sub> flux weighted mean <sup>b</sup> (Pg C yr <sup>-1</sup> )	Riverine CO <sub>2</sub> flux based on J06 <sup>c</sup> (Pg C yr <sup>-1</sup> )	Inv. Anthro. CO <sub>2</sub> flux weighted mean <sup>b</sup> (Pg C yr <sup>-1</sup> )	Inv. Contemp. CO <sub>2</sub> flux weighted mean <sup>b</sup> (Pg C yr <sup>-1</sup> )	T06 Contemp. CO <sub>2</sub> flux square wind <sup>d</sup> (Pg C yr <sup>-1</sup> )
Arctic	1	1.03×10 <sup>13</sup>	-0.02 ± 0.02	0.04 ± 0.02	-0.01 ± 0.03	0.00 ± 0.04	-0.04
High-Lat. North Atlantic	2	9.88×10 <sup>12</sup>	-0.11 ± 0.04	0.03 ± 0.01	-0.09 ± 0.04	-0.17 ± 0.06	-0.30
Mid-Lat. North Atlantic	3	1.03×10 <sup>13</sup>	-0.12 ± 0.04	0.01 ± 0.01	-0.13 ± 0.04	-0.24 ± 0.06	-0.17
Subtropical North Atlantic	4	1.51×10 <sup>13</sup>	-0.08 ± 0.04	0.04 ± 0.02	-0.04 ± 0.04	-0.08 ± 0.06	-0.06
Tropical North Atlantic	5	1.26×10 <sup>13</sup>	0.03 ± 0.01	0.09 ± 0.05	-0.04 ± 0.02	0.08 ± 0.05	0.03
Tropical South Atlantic	6	1.06×10 <sup>13</sup>	0.14 ± 0.02	0.02 ± 0.01	-0.09 ± 0.02	0.06 ± 0.03	0.11
Subtropical South Atlantic	7	8.96×10 <sup>12</sup>	0.02 ± 0.01	0.00 ± 0.00	-0.02 ± 0.01	-0.01 ± 0.01	0.03
Mid-Lat. South Atlantic	8	9.72×10 <sup>12</sup>	-0.11 ± 0.05	0.01 ± 0.00	-0.05 ± 0.02	-0.16 ± 0.05	-0.16
Subpolar South Atlantic	9	8.77×10 <sup>12</sup>	0.11 ± 0.05	0.00 ± 0.00	-0.11 ± 0.07	0.00 ± 0.08	-0.13
Southern Ocean	10	2.59×10 <sup>13</sup>	0.04 ± 0.04	0.00 ± 0.00	-0.24 ± 0.10	-0.20 ± 0.11	0.01
High-Lat. Western N. Pacific	11	4.10×10 <sup>12</sup>	0.04 ± 0.02	0.01 ± 0.01	-0.02 ± 0.01	0.03 ± 0.03	0.02
Mid to High-Lat. East. N. Pac.	12	7.34×10 <sup>12</sup>	-0.02 ± 0.03	0.02 ± 0.01	-0.02 ± 0.01	-0.03 ± 0.03	-0.14
Western Subtrop. North Pacific	13+14	2.22×10 <sup>13</sup>	-0.29 ± 0.05	0.05 ± 0.02	-0.14 ± 0.04	-0.39 ± 0.07	-0.35
Eastern Subtrop. North Pacific	15	1.07×10 <sup>13</sup>	-0.02 ± 0.01	0.00 ± 0.00	-0.01 ± 0.02	-0.03 ± 0.02	-0.06
Western Tropical North Pacific	16	1.95×10 <sup>13</sup>	0.07 ± 0.04	0.02 ± 0.01	-0.05 ± 0.02	0.04 ± 0.04	0.01
Eastern Tropical North Pacific	17	1.67×10 <sup>13</sup>	0.15 ± 0.04	0.01 ± 0.00	-0.13 ± 0.04	0.02 ± 0.06	0.11
Western Tropical South Pacific	18	1.20×10 <sup>13</sup>	0.05 ± 0.02	0.01 ± 0.01	-0.03 ± 0.01	0.03 ± 0.02	0.02
Eastern Tropical South Pacific	19	1.83×10 <sup>13</sup>	0.36 ± 0.08	0.00 ± 0.00	-0.08 ± 0.02	0.28 ± 0.09	0.40
Western South Pacific	20+22	2.50×10 <sup>13</sup>	-0.35 ± 0.09	0.00 ± 0.00	-0.09 ± 0.04	-0.44 ± 0.10	-0.36
Eastern South Pacific	21+23+24	1.38×10 <sup>13</sup>	-0.01 ± 0.03	0.00 ± 0.00	-0.02 ± 0.01	-0.02 ± 0.03	0.04
Subpolar S. Pac. and S. Indian	25+30	2.84×10 <sup>13</sup>	0.25 ± 0.09	0.01 ± 0.00	-0.39 ± 0.12	-0.14 ± 0.15	-0.31
Tropical Indian	26+27	3.24×10 <sup>13</sup>	0.14 ± 0.02	0.09 ± 0.04	-0.11 ± 0.03	0.12 ± 0.06	0.15
South Indian	28+29	2.65×10 <sup>13</sup>	-0.22 ± 0.06	0.01 ± 0.00	-0.24 ± 0.07	-0.46 ± 0.09	-0.46
Total		3.587×10 <sup>14</sup>	0.03 ± 0.07	0.45 ± 0.23	-2.18 ± 0.25	-1.70 ± 0.35	-1.61
Northern Extratropics (>18°N)		8.99×10 <sup>13</sup>	-0.63 ± 0.10	0.19 ± 0.04	-0.47 ± 0.08	-0.90 ± 0.13	-1.11
Tropics (18°S - 18°N)		1.22×10 <sup>14</sup>	0.93 ± 0.11	0.23 ± 0.06	-0.54 ± 0.07	0.62 ± 0.14	0.84
Southern Extratropics (<18°S)		1.47×10 <sup>14</sup>	-0.27 ± 0.17	0.03 ± 0.01	-1.18 ± 0.19	-1.42 ± 0.25	-1.34

<sup>a</sup> Region numbers refer to the 30 regions used for computing the basis functions (see *Mikaloff Fletcher et al.* [2003]).

<sup>b</sup> Weighted mean fluxes estimated by the inversion. For details and estimates of the 10 individual models, see *Mikaloff Fletcher et al.* [2006a] (anthropogenic CO<sub>2</sub> fluxes) and *Mikaloff Fletcher et al.* [2006b] (natural CO<sub>2</sub> fluxes).

<sup>c</sup> The riverine carbon outgassing was estimated on the basis of the spatial distribution of the input of organic and inorganic carbon into the ocean by rivers. See *Jacobson et al.* [2006] for details.

<sup>d</sup> Flux estimates based on the pCO<sub>2</sub> climatology of *Takahashi et al.* [2006]. The fluxes listed were computed assuming a square dependence of the gas transfer velocity on the wind speed (the latter taken from the NCEP reanalysis: [www.cdc.noaa.gov](http://www.cdc.noaa.gov)).