Simulation of anthropogenic CO₂ uptake in the CCSM3.1 ocean circulation-biogeochemical model: comparison with data-based estimates

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Abstract. The global ocean has taken up a large fraction of the CO₂ released by human activities since the industrial revolution. Quantifying the oceanic anthropogenic carbon (Cₐₐₙt) inventory and its variability is important for predicting the future global carbon cycle. The detailed comparison of data-based and model-based estimates is essential for the validation and continued improvement of our prediction capabilities. So far, three global estimates of oceanic Cₐₐₙt inventory that are “data-based” and independent of global ocean circulation models have been produced: one based on the ΔC* method, and two that are based on constraining surface-to-interior transport of tracers, the TTD method and a maximum entropy inversion method (GF). The GF method, in particular, is capable of reconstructing the history of Cₐₐₙt inventory through the industrial era. In the present study we use forward model simulations of the Community Climate System Model (CCSM3.1) to estimate the Cₐₐₙt inventory and compare the results with the data-based estimates. We also use the simulations to test several assumptions of the GF method, including the assumption of constant climate and circulation, which is common to all the data-based estimates. Though the integrated estimates of global Cₐₐₙt inventories are consistent with each other, the regional estimates show discrepancies up to 50 %. The CCSM3 model underestimates the total Cₐₐₙt inventory, in part due to weak mixing and ventilation in the North Atlantic and Southern Ocean. Analyses of different simulation results suggest that key assumptions about ocean circulation and air-sea disequilibrium in the GF method are generally valid on the global scale, but may introduce errors in Cₐₐₙt estimates on regional scales. The GF method should also be used with caution when predicting future oceanic anthropogenic carbon uptake.

1 Introduction

Since the industrial revolution, a large amount of carbon dioxide (CO₂) has been emitted into the atmosphere due to human activities. Increased atmospheric CO₂ is the largest contributor to the anthropogenic Greenhouse effect (Solomon et al., 2007). Given the importance of CO₂ to climate, it is crucial to understand the global carbon cycle. The ocean plays an important role in the global carbon cycle, modulating atmospheric CO₂ concentrations and climate. The global ocean has taken up 20 to 35 percent of CO₂ released by human activities since the industrial revolution (Khatiwala et al., 2009; Sabine et al., 2004; Houghton, 2007). Some studies have suggested that the oceanic carbon sink may have changed during the past few decades (Wang and Moore, 2012; Lovenduski et al., 2007; Le Quéré et al., 2007; Wetzel et al., 2005; Perez et al., 2010b), though significant uncertainties remain (e.g. McKinley et al., 2011). Quantifying the oceanic carbon inventory and its variability is therefore important for understanding the global carbon cycle and how it might change over time.

Anthropogenic carbon concentrations in the ocean cannot be measured directly and the magnitude of the anthropogenic carbon signal is only a few percent of the large natural background. Substantial effort has been made to estimate the oceanic anthropogenic CO₂ inventory and variability based on observational data and model simulations in recent years (Hall et al., 2004; Khatiwala et al., 2009; Álvarez et al., 2009; Mikaloff Fletcher et al., 2006; Sabine et al., 2004; Waugh et al., 2006). Generally, assessments of the oceanic CO₂ uptake based on observations can be made in two ways. One strategy is to estimate the uptake rates by repeated hydrographic measurements of the carbon
This approach, known as the TTD method, assumes that the surface to the interior (or transit time distribution (TTD)). The distribution of times needed for water to be transported from the surface mixed layer via a Green function, the concentration in the interior can be related to its time evolving concentration as a passive, inert tracer. As a consequence, its concentration in the ocean starting at the beginning of the industrial revolution, unlike previous estimates which provided only a snapshot of the C\textsubscript{ant} distribution. Another advantage of the GF method is that it can, in principal, be used to predict future anthropogenic carbon inventories in the ocean, given a projection of the future atmospheric CO\textsubscript{2} concentrations. The validity of such an estimate would however depend on the not only the projected atmospheric CO\textsubscript{2} concentration, but also changes in the ocean’s circulation and biological pump remaining small. Previous observations suggest ongoing changes in the ocean. The heat content of the world ocean has increased substantially over the past few decades (Levitus et al., 2000, 2005) along with the global averaged sea surface temperature (Hansen et al., 2006, 2010). Dissolved oxygen concentrations have also decreased in several ocean basins (Stramma et al., 2008).
CO₂ flux and the carbon pumps, and therefore impact carbon storage in the ocean. For example, warmer sea-surface temperature causes increases in the partial pressure of CO₂ in the surface ocean. Elevated temperature can also increase seawater buffer capacity. These two effects of temperature affect ocean carbon uptake, yet oppose each other. Temperature changes also have significant impact on the growth rate of phytoplankton, which are key players in transporting carbon to the deep ocean. Changing oxygen concentrations and circulation can affect nutrient cycling and distributions, which influence the efficiency of the biological pump. Moreover, some previous research has found that enriched CO₂ could affect ocean biological productivity (Palacios and Zimmerman, 2007; Riebesell et al., 2007). It is also likely that ocean circulation will change in the future as the climate continues to change. With a reduced meridional overturning circulation and convective mixing, the transport of CO₂ into the ocean interior may slow down and oceanic carbon uptake decreases. Currently, all the data-based estimates of anthropogenic carbon uptake assume constant climate and ocean circulation. Also, the biological pump is assumed to be constant and known (the ΔC* method) (Gruber et al., 1996) or unimportant in the sequestration of anthropogenic carbon (the TTD and GF methods) (Waugh et al., 2006; Khatiwala et al., 2009). These assumptions may introduce significant errors in studies of future anthropogenic carbon uptake.

Coupled carbon cycle-ocean circulation models have also been used to estimate the ocean anthropogenic CO₂ inventory and assist the understanding of the global carbon cycle (e.g. Orr et al., 2001; Xu and Li, 2009). The strength of global carbon cycle models embedded within climate models is that they can take into account the impact of changes in future climate. Such models can also provide us with the opportunity to evaluate and interpret data-based estimates. However, given limited resolution and other model imperfections, carbon-cycle models do not necessarily agree with constraints provided by hydrographic tracer observations for the current era. It is necessary to evaluate the performance of carbon cycle models against data-based estimates of the anthropogenic carbon uptake. Several models have been evaluated in this way (Álvarez et al., 2009; Xu and Li, 2009; Orr et al., 2001). Discrepancies exist among different data-based estimates and uncertainties are caused by various assumptions. In a recent review, Sabine and Tanhua (2010) summarized advantages and disadvantages of several different data-based methods, including the TTD method and the ΔC* method, but without a full assessment of each method. Some earlier data-based methods have been evaluated using carbon cycle model simulated data (Matsumoto and Gruber, 2005; Yool et al., 2010; Levine et al., 2008).

Different estimates of oceanic anthropogenic carbon uptake have different strengths and weaknesses. The first step of improvement in estimates is detailed comparisons of different widely-used methods and understanding biases and uncertainties in these methods. One objective of this study is to compare anthropogenic carbon inventory estimated using the GF method and previous global estimates from the ΔC* method and the TTD method. Another objective is to evaluate the ability of a coupled carbon cycle-ocean circulation model to estimate oceanic anthropogenic carbon uptake and compare it with different data-based estimates of anthropogenic carbon inventory in detail. Lastly, we use model simulations to assess assumptions made in the GF method that have not previously been evaluated.

2 Methods and model description

2.1 Coupled carbon cycle – ocean circulation model

The ocean circulation model used in this study is the coarse resolution, ocean circulation component of the National Center for Atmospheric Research (NCAR) Community Climate System Model 3.1 (CCSM3.1) (Collins et al., 2006; Yeager et al., 2006). The carbon cycle was simulated with an ocean biogeochemical/ecosystem model, the Biogeochemical Elemental Cycling (BEC) model (Moore et al., 2002, 2004). The model includes 25 vertical levels with 8 levels in the upper 103 m. It has 100 × 116 horizontal grid points. The longitudinal resolution is 3.6° and the latitudinal resolution varies in the range of 0.9°–2.0°, with finer resolution near the equator. The wind speed-mixing relation in the model was adjusted to better match the observed mixed layer depths in the Southern Ocean (de Boyer Montégut et al., 2004). This modification significantly reduces the model’s mixed layer depth bias, from 18 m too shallow in the standard CCSM 3.1 model to a bias of 2 m too deep in the Southern Ocean (Wang and Moore, 2011).

The BEC model includes five phytoplankton functional groups, one zooplankton group and the biogeochemical cycling of multiple growth limiting nutrients (nitrate, ammonium, phosphate, iron and silicate) (Wang and Moore, 2011; Moore et al., 2004). The five phytoplankton groups are diatoms, diazotrophs, small phytoplankton, coccolithophores and Phaeocystis (Wang and Moore, 2011). The light-, nutrient-, and temperature-dependencies of phytoplankton growth rate are modeled multiplicatively. Phytoplankton growth rates decrease under nutrient stress according to Michaelis-Menten nutrient uptake kinetics. Phytoplankton photoadaptation is described by varying chlorophyll to nitrogen ratios based on the model of Geider et al. (1998). Ecosystem parameters were chosen based on field and laboratory data and were described in detail by Moore et al. (2004, 2002) and Wang and Moore (2011). The ecosystem module is coupled with an ocean biogeochemistry module, which includes full carbonate system thermodynamics and air-sea CO₂ and O₂ fluxes. The atmosphere component is not coupled in this study. Atmospheric conditions, including atmospheric CO₂ concentrations, are prescribed boundary conditions, so that there is no oceanic feedback to the atmosphere.
The model was spun up for 600 years with repeating National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) meteorological reanalysis climate data and satellite-based estimates of climatological sea ice cover (Large and Yeager, 2004) to allow tracer fields in the model to approach an approximate steady state. The initial distributions of nutrients, inorganic carbon and alkalinity were based on the World Ocean Atlas (WOA) 2001 database (Conkright et al., 2002) and the GLODAP database (Key et al., 2004). Dissolved iron initialization was based on simulations from Moore and Braucher (2008). Dust deposition was the repeating annual climatology by Luo et al. (2003). Atmospheric CO$_2$ concentration was set to be 278 ppm for the model initialization. Model year 600 corresponded to year 1764.

Four simulations (summarized in Table 1) were conducted following the initial spin-up of 600 years. The control simulation (Ctrl) was a continuation of the spin-up run, forced with the same repeating climatology data. Atmospheric CO$_2$ concentration remained fixed at 278 ppm. The control run represents the pre-industrial scenario. The second simulation (Cexp) was also forced with repeating climatology data, but the atmospheric CO$_2$ concentrations was prescribed according to the reconstructed historical record starting in 1765. In the third and fourth simulations, labeled Fexp and Texp, the coupled model was forced with a repeating climatological surface forcing until year 1947, at which point the NCEP/NCAR 6-hourly data for momentum, heat, and freshwater fluxes (Large and Yeager, 2009) were then used instead of the climatological boundary forcing. The Ice fraction satellite data were from the Scanning Multichannel Microwave Radiometer (SMMR) from 1978 to 1988 and Special Sensor Microwave/Imager after 1988 (Large and Yeager, 2004). For the period prior to 1978 where satellite ice fraction data were unavailable, we used climatological sea ice data. In Fexp, the atmospheric CO$_2$ concentration was kept fixed at 278 ppm, while in Texp we used the reconstructed atmospheric CO$_2$ concentrations for the years from 1765 onwards. Both Fexp and Texp represent scenarios where changing climate is taken into account. Model performance was previously evaluated against biogeochemical and physical observations (Wang and Moore, 2011, 2012).

2.2 Definitions of anthropogenic carbon (C$_{ant}$)

There are several different definitions for the anthropogenic carbon. In data-based methods, the ocean circulation over the industrial period is generally assumed to be constant (although the GF method allows for a seasonal cycle). Furthermore, while the GF method allows a time-varying air-sea disequilibrium of CO$_2$, both the ΔC* and TTD methods assume it to be constant. Given these assumptions, oceanic anthropogenic carbon should be defined as the uptake of CO$_2$ emitted by human activity, given the premise of approximately constant natural carbon system and constant climate. It is equivalent to the difference between DIC concentrations in the Cexp simulation and in the Ctrl simulation, i.e. $C_{ant, cnst} = [DIC]_{Cexp} - [DIC]_{Ctrl}$.

Oceanic carbon uptake can be affected by climate, ocean circulation, and chemical properties of seawater. Changing temperatures can influence both the uptake capacity for C$_{ant}$ and the potential strength of the solubility pump (as reviewed by Friis, 2006). Changes in sea-surface temperature can also influence the distribution of phytoplankton and the efficiency of the biological pump. Elevated CO$_2$ level is altering the seawater carbon chemistry and causing ocean acidification, which also influence sea-surface $p$CO$_2$ and ocean carbon uptake. Compared with C$_{ant, cnst}$, a better definition of oceanic anthropogenic carbon should include the effects of CO$_2$ induced climate change on C$_{ant}$ uptake. This effect is neglected in data-based methods, but can be estimated in coupled circulation-carbon cycle models. To calculate the anthropogenic carbon inventory based on this definition, the formula should be $C_{ant, var} = [DIC]_{Texp} - [DIC]_{Fexp}$.

As suggested by Keeling (2005), the definition of the C$_{ant}$ used in data-based methods may be an incomplete measure of the change in the ocean carbon content, since the natural carbon system may also change under anthropogenic perturbations. There is another definition, which is transferred from the definition of atmospheric anthropogenic CO$_2$. The anthropogenic portion of CO$_2$ in the atmosphere is commonly known as the CO$_2$ concentration difference between modern time and pre-industrial era, i.e. the increased part of CO$_2$. This definition represents both gross CO$_2$ emissions by human activities and effects on atmospheric CO$_2$ due to changes in uptakes of CO$_2$ by land and ocean as a result of changing climate. A similar definition for the oceanic anthropogenic carbon is usually used in modeling studies (Sarmiento et al., 1992; Le Quéré et al., 2010; Tjiputra et al., 2010), which represents the oceanic uptake of increased atmospheric CO$_2$ and changes in the biological pump and the natural DIC storage due to changing climate. The formula used to calculate the complete C$_{ant}$ is $C_{ant, all} = [DIC]_{Texp} - [DIC]_{Ctrl}$. To evaluate the complete feedback between the ocean carbon system and climate, it is necessary to consider both changes in the natural carbon cycle and emitted anthropogenic CO$_2$.

2.3 Reviewing the assumptions made in the Green function (GF) method

The GF method is a generalization of the TTD method. Like the latter, it makes several assumptions. First, the anthropogenic carbon perturbation is assumed to be a conservative tracer and its distribution in the ocean is only related to physical processes, i.e. the GF method neglects possible changes in the biological pump. The GF estimate of C$_{ant}$ uptake makes use of a Green function, G, which is obtained by deconvolving steady and transient tracer data under a constant climate assumption. The Green function is then used
to propagate the surface history of the anthropogenic carbon, $C_{\text{ant}}$, into the ocean interior.

In order to estimate the (unknown) surface history $C_{\text{ant}}^s$ of anthropogenic carbon, the GF method makes use of mass conservation, namely, the rate of change of inventory of $C_{\text{ant}}$ is equal to the instantaneous air-sea flux of $C_{\text{ant}}$. The rate of change of inventory is given in terms of the convolution of the Green function with $C_{\text{ant}}^s$, while the flux is proportional to the change in surface disequilibrium of CO$_2$. The GF method assumes that this change in air-sea disequilibrium ($\delta p_{\text{CO}_2}$) at any given point is proportional to the anthropogenic CO$_2$ perturbation in the atmosphere, ($\delta p_{\text{CO}_2}^\text{atm}$).

$$\delta p_{\text{CO}_2} \approx \varepsilon \delta p_{\text{CO}_2}^\text{atm},$$

where the proportionality constant, $\varepsilon$, is a function of space. This assumption, justified on the basis of simulations in a carbon cycle model, together with the CO$_2$ system equilibrium chemistry in seawater allows the known atmospheric CO$_2$ history to be related to the surface $C_{\text{ant}}^s$ concentration. The unknown proportionality constant is then constrained by imposing mass conservation and requiring that the computed surface $p_{\text{CO}_2}$ value for the year 2000 matches the observed value in the Takahashi database (Takahashi et al., 2009) (see Khatiwala et al., 2009) for details).

### 2.4 Estimates of the $\Delta C^*$ method and the TTD method

Estimates of global anthropogenic carbon distributions by the $\Delta C^*$ method ($C_{\text{ant},\Delta C^*}$) and the TTD method ($C_{\text{ant},\text{TTD}}$) were obtained from the GLODAP data website (http://cdiac.ornl.gov/oceans/glodap/GlopDV.html). Both estimates were made on subsets of the GLODAP dataset (Waugh et al., 2006; Key et al., 2004; Sabine et al., 2004). The calculations using the $\Delta C^*$ method were made in different studies (Lee et al., 2003; Sabine et al., 2002, 1999) with slight differences in the detailed techniques and synthesized by Sabine et al. (2004, 2005). There are occurrences of negative values of anthropogenic carbon concentrations, mainly due to mapping errors and uncertainties in the dataset, which were set to zero by Sabine et al. (2004). These negative values are not physically possible, and are set to zero in comparisons of column inventories (Fig. 1) and total inventories (Table 2). However, the uncorrected regional inventories are also included in Table 2. In comparisons of regional anthropogenic carbon vertical distributions (Fig. 3), these negative values are kept to show the uncertainties in the method.

Global estimates by the TTD method were made by Waugh et al. (2006) with a pointwise version of this method applied to temperature, salinity and CFC-12 from the GLODAP dataset. The TTD method overestimated anthropogenic carbon concentrations in deep waters when CFC-12 is near or below the measurement detection limit of $\sim 0.005$ pmol kg$^{-1}$ (Waugh et al., 2006). This bias yields an anthropogenic carbon concentration of $2.5 \mu$mol kg$^{-1}$. In calculations of regional and global inventories estimated using the TTD method, all values of anthropogenic carbon concentrations less than $2.5 \mu$mol kg$^{-1}$ are set to zero (Waugh et al., 2006).

Based on model evaluations of the TTD method, it was found that the TTD method overestimated the global anthropogenic carbon inventory by $\sim 20 \%$, with the largest biases in the Southern Ocean. Waugh et al. (2006) corrected this positive bias by reducing the global anthropogenic carbon inventory by $20 \%$. The positive bias was not corrected in data comparisons in this study, since the non-uniform pattern of the bias make the correction difficult.

### Table 1. Simulation descriptions and definition of $C_{\text{ant}}$

| Simulation | AtmCO$_2$ | Forcing | Description |
|------------|-----------|---------|-------------|
| Ctrl       | 278 ppm   | constant climate | provides information of oceanic DIC distributions under the pre-industrial scenario |
| Cexp       | ramp up after 1765 | constant climate | provides information of anthropogenic carbon storage in the ocean under constant climate |
| Fexp       | 278 ppm   | varying climate after 1948 | provides information of changes in natural oceanic carbon system under the changing climate |
| Texp       | ramp up after 1765 | varying climate after 1948 | represents the combined effects of rising atmospheric CO$_2$ and the changing climate on oceanic carbon storage |

| Definitions of $C_{\text{ant}}$ |
|---------------------------------|
| $C_{\text{ant,cnst}}$ | $[\text{DIC}]_{\text{Cexp}} - [\text{DIC}]_{\text{Ctrl}}$ |
| $C_{\text{ant,var}}$  | $[\text{DIC}]_{\text{Fexp}} - [\text{DIC}]_{\text{Fexp}}$ |
| $C_{\text{ant,all}}$  | $[\text{DIC}]_{\text{Fexp}} - [\text{DIC}]_{\text{Ctrl}}$ |
Fig. 1. Column inventories of anthropogenic DIC in year 1994 by (a) the ΔC* method; (b) the TTD method; (c) the GF method; (d) the CCSM (C_{ant,cnst}).

Table 2. Regional and global distributions of C_{ant} inventories in 1994 (in PgC)

| Region          | ΔC* method | Uncorrected ΔC* | TTD method | GF method | CCSM C_{ant,cnst} | CCSM C_{ant,var} | CCSM C_{ant,all} |
|-----------------|------------|-----------------|------------|-----------|--------------------|------------------|------------------|
| N. Pacific      | 16         | 14              | 17         | 18        | 18                 | 18               | 18               |
| S. Pacific      | 13         | 12              | 18         | 16        | 15                 | 14               | 15               |
| N. Indian       | 3          | 3               | 2          | 2         | 2                  | 2                | 2                |
| S. Indian       | 11         | 9               | 11         | 11        | 10                 | 10               | 10               |
| N. Atlantic     | 22         | 21              | 24         | 22        | 18                 | 18               | 18               |
| S. Atlantic     | 10         | 8               | 11         | 8         | 6                  | 6                | 6                |
| Southern Ocean  | 30         | 27              | 49         | 36        | 22                 | 23               | 23               |
| Global          | 106        | 94              | 133        | 114       | 92                 | 91               | 93               |

1 The values are obtained when negative values are kept in calculations
2 The Southern Ocean is defined as south of 35° S.
3 The value calculated by Waugh et al. (2006) is 134 PgC. The difference may occur due to rounding error.

3 Results and discussion

3.1 Oceanic anthropogenic carbon inventory

3.1.1 Global and regional inventories

Since estimates based on the ΔC* and TTD methods are only available for 1994, we first compare the oceanic anthropogenic carbon inventories from the data based estimates and the model simulations for that year. The global inventory estimated using the ΔC* method is 106 ± 17 PgC (Sabine et al., 2004), while that from the TTD method was in the range of 94–121 PgC. The TTD-based inventory has been corrected for a positive bias of ~20% as per model-based evaluations (Waugh et al., 2006). The uncorrected C_{ant,TTD} is 134 PgC, as shown in Fig. 1 and Table 2. The C_{ant,ΔC} is corrected by setting negative values to zero. The uncorrected C_{ant,ΔC} is 94 PgC (Table 2). The inventory estimated using the GF method (C_{ant,GF}) was 114 ± 22 PgC (Khatiwala et al., 2009). These estimates, which are based on the GLODAP dataset, only include the open ocean, i.e. marginal seas and the Arctic Ocean are excluded. The same mask is applied to the CCSM output to make the results comparable.
The global inventory of anthropogenic carbon simulated with CCSM is at the lower end of the data-based estimates, but falls within their uncertainty. The CCSM \( C_{\text{ant,\text{const}}} \) is 92 PgC in the constant climate scenario (Cexp). This is about 20 % lower than \( C_{\text{ant,GF}} \) and about 14 % lower than \( C_{\text{ant,TTD}} \) and \( C_{\text{ant,C*}} \). Despite the nominal agreement with the previously quoted error bars, we believe that the CCSM estimate is likely low for reasons to be discussed later. The CCSM simulated \( C_{\text{ant}} \) under varying climate is similar to that under constant climate on both global and regional scales.

The column inventories from data-based estimates and CCSM simulations show a similar pattern, with high column inventories in the North Atlantic and the Southern Ocean (Fig. 1). Regional inventories of anthropogenic carbon from different methods are compared in Table 2. The various estimates agree best in the Indian Ocean. The total \( C_{\text{ant}} \) in the Indian Ocean is \( \sim 13 \) PgC, with \( \sim 2 \) PgC in the North Indian Ocean and \( \sim 11 \) PgC in the South Indian Ocean (\( > 35^\circ \text{S} \)). When the area south of \( 35^\circ \text{S} \) is included, the values of \( C_{\text{ant}} \) are larger than 20 PgC, which is greater than the estimate of \( C_{\text{ant}} \) in the Indian Ocean in the year 2000 by Hall and Primeau (2004) and Hall et al. (2004). Except the \( C_{\text{ant,C*}} \), all other \( C_{\text{ant}} \) estimates in the deep South Indian Ocean reveal penetration of anthropogenic carbon, consistent with results by Álvarez et al. (2009). Estimates of \( C_{\text{ant}} \) in the Pacific Ocean are also consistent, in the range of \( 29 \) PgC–35 PgC with the lower bound set by \( C_{\text{ant,\text{const}}} \) and the upper bound set by \( C_{\text{ant,TTD}} \) (Table 2).

Estimated inventories of \( C_{\text{ant}} \) have less agreement in the Atlantic Ocean and the Southern Ocean. The \( C_{\text{ant,TTD}} \) in the Southern Ocean is much larger than other estimates (Fig. 2 and Table 2). Model evaluations of the TTD method suggested that the TTD method overestimated the \( C_{\text{ant}} \) inventory by 60 % in the Southern Ocean, due to a positive bias in the deep convection region primarily caused by the assumption of constant disequilibrium (Waugh et al., 2006). \( C_{\text{ant}} \) inventories estimated using the \( \Delta C^* \) and GF methods are the same in the North Atlantic Ocean, but the regional patterns of column inventories are different (Fig. 2). The column inventories of \( C_{\text{ant,GF}} \) are higher in deep water formation region of the North Atlantic Ocean and lower in the low to mid-latitude region of the North Atlantic Ocean. These regional differences are averaged out over the entire North Atlantic Ocean. The estimated \( C_{\text{ant,\Delta C*}} \) in the Southern Ocean is also significantly smaller than the \( C_{\text{ant,GF}} \), mainly because of the assumption of constant disequilibrium made in the \( \Delta C^* \) method (Khatiwala et al., 2009). Previous model evaluation suggested that the \( \Delta C^* \)-based estimate showed significant deviations from the "true" value in the high-latitude North Atlantic Ocean and the Southern Ocean and indicated that the estimate from this method may be erroneous in these regions (Levine et al., 2008). The estimated \( C_{\text{ant}} \) from the CCSM is smaller than estimates using data-based methods in both the Atlantic Ocean and the Southern Ocean (Table 2 and Fig. 2). Compared with the GF method, the CCSM is 18 % and 33 % lower in the North Atlantic Ocean and the Southern Ocean, respectively. It is likely that the CCSM underestimates the \( C_{\text{ant}} \) in these two regions. The mixing in the CCSM is relatively weak at high latitudes, especially in the Southern Ocean. A previous study has suggested that the weak formation and ventilation of Antarctic Intermediate Water and mode waters in the Southern Ocean in the CCSM may cause a weak oceanic carbon sink (Thornton et al., 2009). If we include the Atlantic sector of the Southern Ocean, \( C_{\text{ant}} \) from the \( \Delta C^* \) method, the GF method, and from the CCSM are smaller than the estimate of Vázquez-Rodríguez et al. (2009), who suggested that the \( C_{\text{ant}} \) inventory in the South Atlantic Ocean was \( 22 \pm 5 \) PgC in 1994. Estimates for the North Atlantic Ocean cannot be compared directly with Vázquez-Rodríguez et al. (2009)'s study because
of the different latitude ranges used in each estimate. However, the estimate for the North Atlantic Ocean using the TTD method is on the low end of all estimates in the study by Vázquez-Rodríguez et al. (2009), but nevertheless is the largest of all estimates in our comparisons.

### 3.1.2 Vertical profiles of \( \text{C}_{\text{ant}} \)

As shown in Fig. 2 and Table 2, regional distributions of anthropogenic carbon estimated using different methods can be very different, although the total regional inventories are similar. Figure 3 shows vertical profiles of the regionally-averaged \( \text{C}_{\text{ant}} \) from data-based methods and the CCSM. The Southern Ocean is defined as areas south of 35\(^\circ\) S. Regionally averaged \( \text{C}_{\text{ant}} \) vertical profiles estimated using different methods show a similar pattern. Surface \( \text{C}_{\text{ant}} \) concentrations are in the range of 45–57 mmol m\(^{-3}\) in most regions. The lowest surface \( \text{C}_{\text{ant}} \) values are in the Southern Ocean, in the range of 38–41 mmol m\(^{-3}\). In most regions, \( \text{C}_{\text{ant}} \) values decrease rapidly with depth down to about 1000 m to 1500 m, and then remain approximately constant and low in deeper waters. The concentrations of \( \text{C}_{\text{ant}} \) in deep waters are highest in the North Atlantic Ocean, where there is strong deep convection and mixing. The upper-ocean \( \text{C}_{\text{ant}} \) estimates from different methods are wide-ranging in most regions, especially in the Southern Hemisphere (Fig. 3).

As mentioned above, there are some problematic values in estimates using the \( \Delta C^\text{*} \) method and the TTD method. The \( \Delta C^\text{*} \) method produces negative values in deep waters in all regions. This is probably due to technical errors and uncertainties of the method, but the cause is unclear (Waugh et al., 2005). The negative values indicate certain issues in the \( \Delta C^\text{*} \) method, which suggests clearer problems in the estimates of \( \text{C}_{\text{ant}} \). Compared with the other methods, values of \( \text{C}_{\text{ant}} \) based on the TTD method are highest in deep waters in most regions, especially in the Atlantic Ocean and the Southern Ocean. One major source of uncertainties in the TTD is the assumption of constant disequilibrium, used for estimating the surface \( \text{C}_{\text{ant}} \) history. A previous study showed that estimated \( \text{C}_{\text{ant}} \) in the TTD method overestimated the \( \text{C}_{\text{ant}} \) inventory by 5–10\% (Hall et al., 2004). Waugh et al. (2004) suggested that the estimated \( \text{C}_{\text{ant}} \) with constant disequilibrium assumption most likely represented the upper bounds of \( \text{C}_{\text{ant}} \). Thus, high values of \( \text{C}_{\text{ant}} \) from the TTD shown in Fig. 3 likely overestimate the \( \text{C}_{\text{ant}} \) inventory.

The estimated \( \text{C}_{\text{ant}} \) in the GF method is likely more realistic than the other two estimates. The \( \text{C}_{\text{ant}} \) estimated using the GF method usually falls in the middle of the range of all estimated \( \text{C}_{\text{ant}} \) in these comparisons. The values of \( \text{C}_{\text{ant}} \) from the TTD method by Waugh et al. (2006) is under the assumption of constant disequilibrium in the industrial era. The estimate should be lowered if varying disequilibrium is allowed in calculations, especially in younger waters (Waugh et al., 2004; Hall et al., 2004). The GF method, which is a generalization of the TTD method, is also based on the assumption that anthropogenic CO\(_2\) can be treated as a conservative tracer. But it relaxes some assumptions in the TTD method. The GF method includes the mixing of waters with different ages and types and allows the disequilibrium to evolve through time. The allowance for an increasing air-sea disequilibrium explains why the \( \text{C}_{\text{ant}} \) inventory estimated using the GF method is generally smaller than that from the TTD method. The differences in assumptions and considerations between the TTD method and the GF method also explain why the \( \text{C}_{\text{ant}} \) needs to be corrected for the positive bias, while the GF method does not require such a correction.

The regionally-averaged \( \text{C}_{\text{ant}} \) estimated using the CCSM exhibits roughly similar vertical profiles with other estimated \( \text{C}_{\text{ant}} \), but with some differences in the details. The \( \text{C}_{\text{ant,cnst}} \) from the CCSM falls on the upper end of all estimates in the upper 200 m in the Indian Ocean and the Pacific Ocean. There is a good agreement between the \( \text{C}_{\text{ant,cnst}} \) and the \( \text{C}_{\text{ant,TTD}} \) in the upper 200 m in the Atlantic Ocean, but both are \( \sim \)5 mmol m\(^{-3}\) higher than the \( \text{C}_{\text{ant,GF}} \). The \( \text{C}_{\text{ant}} \) estimated using the CCSM turns out to be lower than other estimates of \( \text{C}_{\text{ant}} \) below 400 m. There are clear differences between the regional averaged \( \text{C}_{\text{ant,cnst}} \) and the \( \text{C}_{\text{ant,all}} \), whereas the later one includes impacts of varying climate. The differences between the \( \text{C}_{\text{ant,cnst}} \) and the \( \text{C}_{\text{ant,all}} \) are in the range of \( \sim 4 \text{ mmol m}^{-3} \)–\( 6 \text{ mmol m}^{-3} \) in the upper 1000 m. The largest difference between the \( \text{C}_{\text{ant,cnst}} \) and other data-based \( \text{C}_{\text{ant}} \) are located along ventilation pathways, between 1000 m and 2500 m in the North Atlantic Ocean and between 500 m and 1500 m in the Southern Ocean. These differences are mainly due to weak mixing and ventilation in the Southern Ocean and the North Atlantic Ocean in the CCSM (Thorton et al., 2009). The weak downward transport of CO\(_2\) in these two regions causes the CCSM to underestimate the global \( \text{C}_{\text{ant}} \) inventory. The differences between the modeled \( \text{C}_{\text{ant}} \) and the \( \text{C}_{\text{ant,GF}} \) during 1948 to 2003 are also shown in Fig. 4. The global \( \text{C}_{\text{ant}} \) inventories estimated using the CCSM increase from \( \sim 45 \text{ PgC} \) to \( \sim 107 \text{ PgC} \) from 1948 to 2003, while the \( \text{C}_{\text{ant,GF}} \) inventories increase from 57 PgC to 132 PgC. Through the study period, the \( \text{C}_{\text{ant}} \) estimated using the CCSM changes proportionally to the \( \text{C}_{\text{ant,GF}} \), a value about 80\% of the \( \text{C}_{\text{ant,GF}} \). The bias in circulation causes a bias in the uptake of anthropogenic carbon. The accumulation of this bias leads to increasing differences between the model simulated \( \text{C}_{\text{ant}} \) and the GF estimate.

### 3.2 Assessment of assumptions of the GF method

There are three major assumptions in the GF method as reviewed in Sect. 2.3. With these assumptions, the GF method can estimate the time-varying distribution of \( \text{C}_{\text{ant}} \) in the ocean, given tracer observations and the atmospheric CO\(_2\).

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The coupled BEC/CCSM is therefore not capable of testing the first assumption in the GF method, because the same assumption is made in the BEC model. The other two

**Fig. 3.** Comparison of regional averaged $C_{\text{ant}}$ from the $\Delta C^*$ method (blue asterisks), the TTD method (black circles), the GF method (red crosses), the $C_{\text{ant,cnst}}$ from the CCSM (green triangles), and the $C_{\text{ant,all}}$ from the CCSM (magenta diamonds). The Southern Ocean is south of 35° S. Data are vertically interpolated to compare values at the same depth.

Some earlier research found increased CO$_2$ uptake by phytoplankton and seagrasses under higher CO$_2$ conditions (Palacios and Zimmerman, 2007; Riebesell et al., 2007). These results suggest that enriched CO$_2$ may stimulate ocean biological productivity and modify the cycling of nutrients, indicating possible biological effects on anthropogenic carbon uptake. Under such conditions, $C_{\text{ant}}$ may not behave as a conservative tracer, although the strength of these effects is poorly constrained and likely smaller than the quoted uncertainty of the GF method. It would be useful to evaluate such effects, but the BEC model does not include influences of changes in CO$_2$ concentrations on phytoplankton growth or calcification, i.e., the biology does not change with rising CO$_2$. The coupled BEC/CCSM is therefore not capable of testing the first assumption in the GF method, because the same assumption is made in the BEC model. The other two

**Fig. 4.** Changes in the estimated ocean $C_{\text{ant}}$ inventories, estimated using the GF method and the CCSM, through time (the left axis) and the differences in total DIC inventory from two simulations forced with rising atmospheric CO$_2$ (the right axis). Shown are: $C_{\text{ant,GF}}$ (red crosses); $C_{\text{ant,cnst}}$ (green triangles); $C_{\text{ant,all}}$ (blue crosses); $C_{\text{ant,all}}$ (magenta diamonds). The black squares show the differences between total DIC in the Texp and in the Cexp.
assumptions can, however, be examined with the CCSM simulations. The GF method is not applied to the coupled ocean carbon cycle model in this study as in previous assessments of the ΔC* method by Matsumoto and Gruber (2005) and of the TTD method by Waugh et al. (2006). Instead, simulation results from different scenarios are analyzed and compared to determine the applicability of the assumptions made in the GF method.

### 3.2.1 Assumption of constant circulation

The second assumption made by the GF method is that ocean circulation is in a cyclo-stationary state, i.e. a seasonally repeating steady state. This assumption is made out of necessity, as there are insufficient observations available to constrain the Green function for a time-varying circulation. The GF method therefore cannot capture the effect of climate change on ocean circulation and hence CO₂ uptake. To evaluate the potential errors, we compare changes in global C<sub>ant</sub> inventories estimated using different CCSM simulations and the GF method from 1948 to 2003 (Fig. 4).

As defined in Sect. 2.2, C<sub>ant, cnst</sub> represents the oceanic anthropogenic carbon inventory under constant climate and constant circulation, while C<sub>ant, var</sub> is the net increase in DIC due to rising atmospheric CO₂ if changing climate is taken into account. The difference between C<sub>ant, cnst</sub> and C<sub>ant, var</sub> is the difference in CO₂ uptake due to changing climate, and thus a measure of the error introduced due to the assumption of constant circulation in the GF method. The globally integrated C<sub>ant, cnst</sub> inventories and C<sub>ant, var</sub> inventories match well in the period from 1948 to 2003 (Fig. 4). The difference between C<sub>ant, cnst</sub> and C<sub>ant, var</sub> is less than 1% of the C<sub>ant</sub> inventory estimate. To the extent that the variability in the CCSM circulation is representative of the true variability, our comparison implies that possible circulation changes due to climate change during the period of 1948 to 2003 have had little impact on the C<sub>ant</sub> inventory. On a global scale, the assumption of constant circulation for the last few decades that is made in the GF method and the other data-based estimates appears to be a reasonable one.

If the total anthropogenic perturbation on the ocean carbon system is considered, the anthropogenic carbon should be defined as C<sub>ant, all</sub>, which includes changes in DIC storage due to both rising atmospheric CO₂ and changing climate. The difference between C<sub>ant, all</sub> and C<sub>ant, cnst</sub> is shown as ΔT<sub>exp</sub>−C<sub>exp</sub> in Fig. 4. ΔT<sub>exp</sub>−C<sub>exp</sub> values vary from −2 PgC to 2 PgC, which are greater than the difference between C<sub>ant, cnst</sub> and C<sub>ant, var</sub>. Overall difference between C<sub>ant, all</sub> and C<sub>ant, cnst</sub> are less than ±4% of the total inventory, which is well within the roughly 20% uncertainty of the GF method (as well as other data-based techniques). Differences between C<sub>ant, all</sub> and C<sub>ant, cnst</sub> largely arise from changes in the natural carbon cycle, particularly from the solubility pump. While the GF method can take into account variations in sea surface temperature, salinity, and winds as it impacts the solubility pump, this was not done in their 2009 study.) The small impact of changing climate on both the natural carbon system and the uptake of C<sub>ant</sub> suggests that the assumption of constant circulation and fixed Green functions are reasonable in the study period for the total anthropogenic perturbation estimate.

Values of ΔT<sub>exp</sub>−C<sub>exp</sub> are larger than 4 PgC in the upper 500 m, and account for more than 10% of the total anthropogenic carbon signal. Vertical profiles of C<sub>ant, all</sub> and C<sub>ant, cnst</sub> also show substantial differences in the upper ocean (Fig. 3). This suggests that the upper-ocean carbon cycle is sensitive to changes in physical fields caused by changing climate, but these differences tend to average out in the integrated carbon inventories. Climate change has had limited impacts on the natural carbon cycle. But it seems likely that both the solubility pump and the biological pump will be increasingly affected in future decades as climate continues to change. A modeling study suggested changes in the natural carbon content may be equivalent to a 5% reduction in the C<sub>ant</sub> inventory in 1994 (Matsumoto et al., 2010). Thus, the assumption of constant circulation may potentially cause a bias in predictions of the future ocean carbon inventory, although the magnitude of this bias is likely to be smaller than the uncertainty in the GF and other data-based methods for C<sub>ant</sub>.

### 3.2.2 Assumption about disequilibrium

The carbon system and CO₂ flux are calculated explicitly in the CCSM, as is the air-sea CO₂ disequilibrium (δpCO₂). Since ocean uptake of anthropogenic carbon is driven by the change in disequilibrium, it is necessary to have an estimate of the space- and time-varying disequilibrium to accurately estimate C<sub>ant</sub> uptake from observations. To address this problem, Khatiwala et al. (2009) assumed that the change in disequilibrium (δΔpCO₂) is proportional to the anthropogenic perturbation on the atmospheric CO₂ (δpCO₂<sub>atm</sub>). This assumption of a linear relationship between δΔpCO₂ and δpCO₂<sub>atm</sub> was justified on the basis of simulations in an ocean carbon cycle model which showed a similar relationship (Fig. S1 in Khatiwala et al. (2009)). By combining this relationship with surface ocean pCO₂ measurements and nonlinear CO₂ system chemistry, the GF method can arrive at an estimate of the surface ocean time history of anthropogenic carbon. Here, we assess this assumption with our CCSM simulations, focusing in particular on the effect of climate change. The δΔpCO₂ is calculated using differences of δpCO₂ between two simulations with reconstructed CO₂ concentrations and the pre-industrial CO₂ concentration, respectively. Changes in the disequilibrium (δΔpCO₂<sub>var</sub>) for C<sub>ant</sub> uptake under changing climate are calculated as the difference between ΔpCO₂ in Texp and Fexp. Similarly, in the case including both changes in the natural carbon system and the C<sub>ant</sub> uptake, δΔpCO₂<sub>all</sub> is the difference between ΔpCO₂ in Texp and Ctrl. The δpCO₂<sub>atm</sub> is calculated based...
on the historical atmospheric CO\textsubscript{2} records used in the simulations. An approximate linear relationship can be found in simulations under constant climate (Ctrl and Cexp). However, there is a much weaker linear relationship between \( \delta \Delta p\text{CO}_2 \) and \( \delta p\text{CO}_2^\text{atm} \) in the cases that include the effects of changing climate (not shown). There are significant interannual and decadal variations in \( \delta \Delta p\text{CO}_2 \), which do not follow changes in atmospheric CO\textsubscript{2} very closely. These variations are mainly due to the effects of varying climate on the natural carbon system and/or the biological pump.

Figure 5 shows the standard errors (SSE) in the linear relationship between \( \delta \Delta p\text{CO}_2 \text{var} \) and \( \delta p\text{CO}_2^\text{atm} \) and between \( \delta \Delta p\text{CO}_2 \text{all} \) and \( \delta p\text{CO}_2^\text{atm} \), where \( \epsilon \) in the linear relationship of \( \delta \Delta p\text{CO}_2 = \epsilon \* \delta p\text{CO}_2^\text{atm} \) is determined using ordinary least squares with data from year 1765 to year 2003. Therefore the errors shown in Fig. 5 are the minimum that could be achieved given the linear relationship assumption. The magnitude of errors in the assumption of linear relationship between \( \delta \Delta p\text{CO}_2 \text{var} \) and \( \delta p\text{CO}_2^\text{atm} \) are usually 10 \% \sim 25 \% of the averaged \( \delta \Delta p\text{CO}_2 \text{var} \). The SSE is generally small in regions where changes in the disequilibrium are small. The errors are larger in the Arctic Ocean, the North Atlantic Ocean, the Equatorial Pacific and the Southern Ocean. The deviations from the assumed linear relationship in the Arctic Ocean are likely due to the presence of sea ice that inhibits the CO\textsubscript{2} exchange. The magnitude of SSE in this region is generally 10 \% \sim 30 \% of the \( \delta \Delta p\text{CO}_2 \). However, the Arctic Ocean is not included in the GF estimate. There is a significant cancellation of errors when the \( \delta \Delta p\text{CO}_2 \text{var} \) is averaged over a 10-year period, but on shorter time scales substantial errors remain. The SSE is larger than 50 \% of the averaged \( \delta \Delta p\text{CO}_2 \text{all} \) in most of the ocean and more than 100 \% in many areas. The errors are also larger in the Arctic Ocean, the North Atlantic Ocean, and increased significantly in the Equatorial Pacific and areas along the Antarctic continental shelf. These errors are mainly due to variations of the natural carbon system under changing climate. The values of SSE are still larger than 2 ppm in many areas after the \( \delta \Delta p\text{CO}_2 \text{var} \) is smoothed over 10 years. Calculating the history of CO\textsubscript{2} flux using the linear relationship may cause significant biases on regional scales, especially in large-SSE regions such as the southernmost Southern Ocean, the North Atlantic Ocean, the Equatorial Pacific Ocean, and areas with high biological productivity. However, it is likely that those errors are smoothed out when averaging over large spatial scales (as done in the GF method).

4 Summary

In this study, three widely used data-based global estimates of oceanic anthropogenic carbon inventories were compared with the \( C_{\text{ant}} \) inventories simulated using BEC/CCSM coupled model under different forcing scenarios. The largest differences among \( C_{\text{ant}} \) estimates occur in the Southern Ocean. Though the integrated \( C_{\text{ant}} \) inventories are similar in the North Atlantic Ocean, this is the region of second largest spatial discrepancies between the estimates. The oceanic anthropogenic carbon estimated using different data-based methods show quite different spatial patterns, though the globally integrated \( C_{\text{ant}} \) inventories are consistent. The coupled model can generally reproduce the oceanic anthropogenic carbon inventory and distribution patterns, but it is likely to underestimate the global \( C_{\text{ant}} \) inventory. The modeled global inventory of anthropogenic carbon is 92 PgC in simulations with constant climate and the estimates are similar in simulations with changing climate. The model estimates are roughly within the error bars of data-based estimates, but on the low end of the range. The CCSM mainly underestimates the \( C_{\text{ant}} \) inventory in the Southern Ocean and the North Atlantic Ocean. This is primarily due to weak deep mixing and ventilation in those regions in the CCSM. The result is consistent with a previous study by Thornton et al. (2009). The weak Southern Ocean ventilation is partly due to a persistent shallow bias in mixed layer depth. Along the Antarctic Circumpolar Current, mixed layer depths are too shallow by 20–40 m during summer months, and by more than 100 m in some regions during winter months. This is thought to be a result of missing processes in the model (near inertial wave and Langmuir mixing). Ongoing research is attempting to reduce
Fig. 5. Maps of the standard error (top two panels) in the linear relationship between $\delta \Delta pCO_2$ and $\delta pCO_2^{\text{atm}}$ with varying climate on annual time scale; maps of 10 yr smoothed standard errors (middle two panels); and maps of mean $\delta \Delta pCO_2$ from 1948–2003 (bottom two panels). Maps on the left column are calculated using the Fexp and the Texp, i.e. $\delta \Delta pCO_2^{\text{var}} = \Delta pCO_2^{\text{Texp}} - \Delta pCO_2^{\text{Fexp}}$. Maps on the right column are calculated using the Ctrl and the Texp, i.e. $\delta \Delta pCO_2^{\text{all}} = \Delta pCO_2^{\text{Texp}} - \Delta pCO_2^{\text{Ctrl}}$.

Our simulations suggest that the assumption of constant ocean circulation during the last few decades made by all the data-based estimates is reasonable. There is less than 1% difference between estimates of the $C_{\text{ant}}$ inventories under constant climate (constant circulation) and varying climate. If changes in the natural cycle system caused by varying climate are taken into account, the uncertainty due to this assumption is less than 4% of the $C_{\text{ant}}$ inventory, which falls within the uncertainty of data-based estimates. The accuracy of calculations of the carbon history in the surface ocean is closely related to the linear-relationship assumption. Our analysis suggests that the linear relationship between changes in air-sea disequilibrium and changes in atmospheric $CO_2$ is a good approximation under constant climate. There are substantial errors in the assumption of linear relationship in simulations with changing climate, especially if the complete anthropogenic perturbation on the ocean carbon system is considered. While these errors may cause significant biases in estimates on regional scales and interannual time scales, they would not significantly affect the global carbon estimates in the past.
scales, they largely average out on large spatial scales and on time scales of decades or longer. (It is important to note that these errors refer to the surface boundary condition for $\text{CO}_2$, and not the interior concentrations, which depend on the convolution of the boundary condition with the GF.) The GF method depends critically on the assumption that changes in the ocean circulation do not affect the uptake of anthropogenic CO$_2$. Uncertainties in its regional estimates may be larger, especially for the Southern Ocean, North Atlantic and Equatorial Pacific. Recent research on decadal trends shows that the capacity of the ocean to take up CO$_2$ may have decreased in these regions (Sabine and Gruber, 2006; Corbière et al., 2007; Schuster and Watson, 2007; Metzl, 2009), where the linear relationship assumption may be problematic. The GF method should be used with caution in predicting future oceanic anthropogenic carbon uptake, with careful consideration of uncertainties due to changing climate. These uncertainties are, however, likely to remain smaller than the intrinsic uncertainty of the GF and other data-based methods.

In this study, we have also tried to clarify the definition of anthropogenic carbon in the ocean. For the purpose of closing the carbon budget of the atmosphere, the land and the ocean, the anthropogenic carbon should be defined as the ocean uptake of increased atmospheric CO$_2$ under changing climate. A more broad and complete definition of the anthropogenic perturbation on the ocean carbon system should include the uptake of anthropogenic CO$_2$ and changes in the natural carbon system. Some studies have shown that changes in the natural carbon cycle contributed to observed changes in the DIC, and that transport and distributions of anthropogenic CO$_2$ may be affected by climate variability (Thomas et al., 2008; Wanninkhof et al., 2010). With the knowledge that the global ocean has been undergoing many changes (Hansen et al., 2010; Levitus et al., 2005; Thompson and Solomon, 2002), the broader definition of anthropogenic carbon perturbation should be adopted in global carbon cycle research. So far, model-based assessments have shown that many data-based methods have substantial potential errors (Yool et al., 2010; Waugh et al., 2006; Levine et al., 2008; Matsumoto and Gruber, 2005). It seems that what is needed is reduction and better quantification of errors on regional scales in data-based methods so that model errors can be better pinned down. It is also necessary to further develop the data-based methods and models and to characterize related uncertainties based on sustained observations in order to understand the variability of the ocean carbon system.

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