REVIEWS IN ADVANCE

Natural Variability and Anthropogenic Trends in the Ocean Carbon Sink

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Abstract

Since preindustrial times, the ocean has removed from the atmosphere 41% of the carbon emitted by human industrial activities. Despite significant uncertainties, the balance of evidence indicates that the globally integrated rate of ocean carbon uptake is increasing in response to increasing atmospheric CO_2 concentrations. The El Niño–Southern Oscillation in the equatorial Pacific dominates interannual variability of the globally integrated sink. Modes of climate variability in high latitudes are correlated with variability in regional carbon sinks, but mechanistic understanding is incomplete. Regional sink variability, combined with sparse sampling, means that the growing oceanic sink cannot yet be directly detected from available surface data. Accurate and precise shipboard observations need to be continued and increasingly complemented with autonomous observations. These data, together with a variety of mechanistic and diagnostic models, are needed for better understanding, long-term monitoring, and future projections of this critical climate regulation service.

1. INTRODUCTION

The ocean carbon sink is the only cumulative net sink of anthropogenic carbon from the atmosphere, having absorbed 41% of all emissions resulting from fossil fuel use and cement manufacture (Sabine et al. 2004, Khatiwala et al. 2009, Ciais et al. 2013) (**Figure 1**). This sink is expected to grow and to substantially mitigate atmospheric carbon accumulation over the next several centuries (Randerson et al. 2015). On timescales of many thousands of years, the ocean should absorb approximately 85% of all anthropogenic carbon emissions (Archer et al. 2009).

The rate at which the ocean carbon sink grows, as well as the potential for it to be slowed by climate change feedbacks, is of critical interest. The rate of ocean carbon uptake largely determines the percentage of emissions that will remain in the atmosphere (**Figure 1**) and thus the warming that will occur. Over the last two decades, there is sufficient evidence to support the conclusion that the globally integrated ocean carbon sink has grown in response to increasing atmospheric carbon content (Ciais et al. 2013, Khatiwala et al. 2013). However, uncertainty remains large (~30% of the mean sink). To reduce this uncertainty, a deeper, mechanistic understanding at regional scales is needed (Lovenduski et al. 2016). Only with this knowledge will accurate assessment of the long-term strengths and potential vulnerabilities of this critical climate regulation service from the ocean become possible. Direct monitoring of the evolving sink will be continually necessary and will assist in verifying and improving future projections.

From global to regional scales, variability in ocean carbon uptake is significant (Long et al. 2013, Schuster et al. 2013, Wanninkhof et al. 2013). This variability obfuscates the detection of long-term

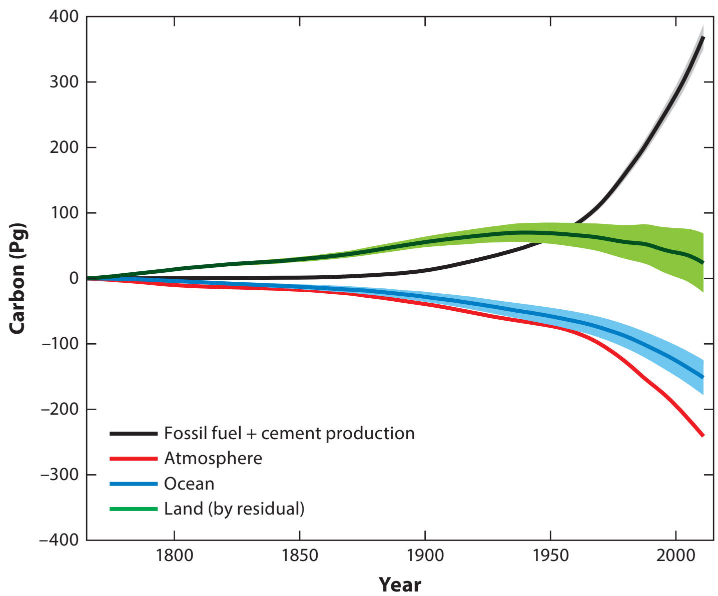


Figure 1

Cumulative sources and sinks of anthropogenic carbon for 1765–2011, using the reconstruction of ocean anthropogenic carbon accumulation by Khatiwala et al. (2009, 2013), fossil fuel and cement manufacture emissions from the Carbon Dioxide Information Analysis Center (**http://cdiac.ornl.gov/ftp/trends/co2_emis**), and atmospheric data from NOAA's Global Monitoring Division (**http://www.esrl.noaa.gov/gmd/ccgg/trends/data.html**). The net land source is calculated by difference [land = (fossil fuel + cement production) − atmosphere − ocean]. The light shading is uncertainty as estimated by the original sources, and from difference for land. Figure created following Khatiwala et al. (2009).

change in the carbon sink (McKinley et al. 2011, 2016; Fay & McKinley 2013; Landschützer et al. 2015). The sparsity of in situ data is a major additional complication in the detection of change (Fay et al. 2014, Lovenduski et al. 2015). Studies with both models and observations indicate that the primary driver of ocean carbon sink variability is ocean circulation and ventilation of the deep ocean. These processes are associated with variable winds and buoyancy fluxes that have some large-scale coherence in modes of climate variability, e.g., the El Niño–Southern Oscillation (ENSO), Southern Annular Mode (SAM), Pacific Decadal Oscillation (PDO), and North Atlantic Oscillation (NAO).

The ability of the ocean to sequester carbon from the atmosphere is also linked to biological productivity, with the biological pump creating a large surface-to-depth vertical gradient of carbon (Sarmiento & Gruber 2006). Although variability in productivity has been observed globally (Behrenfeld et al. 2006), the data are insufficient to determine whether anthropogenic trends are occurring (Henson et al. 2010). As yet, there is no evidence that this variability significantly modifies carbon uptake (Bennington et al. 2009).

In this review, we discuss the need for a deeper understanding of variability and trends in the ocean carbon sink (Section 2), review the basic processes driving ocean carbon uptake (Section 3), describe approaches used to quantify variability and assess its mechanisms (Section 4), summarize current understanding of variability and trends for the globe and four key regions (Section 5), and outline a new climate modeling approach that allows for holistic quantification of the relative magnitudes of variability and trends (Section 6).

2. MOTIVATION: WHY IS IT CRITICAL TO KNOW MORE THAN JUST THE GLOBALLY INTEGRATED SINK?

As summarized in the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report, the globally integrated ocean carbon sink over decadal timescales has changed from -2.0 ± 0.7 Pg C y^{-1} for 1980–1989 to -2.3 ± 0.7 Pg C y^{-1} for 2000–2009 (where negative values indicate ocean carbon uptake, with 90% confidence intervals on decadal mean flux estimates) (Ciais et al. 2013). That the growth of the partial pressure of CO_2 gas in the atmosphere ($p\mathrm{CO_2}^{\mathrm{atm}}$) drives a growing oceanic sink is consistent with our basic understanding that, as the globally averaged atmosphere-to-ocean $p\mathrm{CO_2}$ gradient increases, carbon accumulation in the ocean will occur at an increasing rate (Section 3). This behavior has been illustrated clearly with models forced with only historically observed increases in $p\mathrm{CO_2}^{\mathrm{atm}}$ and no climate variability or change (Graven et al. 2012, Ciais et al. 2013). Nonetheless, critical mysteries remain and weigh heavily on our ability to quantify relationships between the perturbed global carbon cycle and climate change.

First, substantial uncertainty remains on the mean sink (~30% of the total flux). Formally, the quantitative estimate of the 1980–1989 sink (-2.0 ± 0.7 Pg C y^{-1}) is not statistically distinguishable from that for 2000–2009 (-2.3 ± 0.7 Pg C y^{-1}). Reducing this uncertainty is absolutely critical to global partitioning of anthropogenic carbon sources and sinks. Each year, the Global Carbon Project (**http://www.globalcarbonproject.org**) estimates global sources and sinks of carbon, but because the heterogeneous and sparsely measured terrestrial biosphere cannot be directly measured, its flux is estimated by difference from estimated anthropogenic sources and the ocean sink (Le Quéré et al. 2015). In these budgets, land use change uncertainty is at least 50% of the mean flux, and uncertainty is growing for emissions from fossil fuel burning and cement manufacture (Ciais et al. 2013). Reduction in ocean sink uncertainty could therefore help to compensate from a global budgeting perspective.

Constraining regional carbon fluxes is of increasing importance as the international community considers moving toward national or subnational carbon accounting (Ciais et al. 2014, McKinley et al. 2015). Such accounting often depends on atmospheric inversions that lead to substantial difficulty in separating land and ocean fluxes, and where the a priori assumption for spatial distributions of air-sea CO_2 fluxes has a large uncertainty (Schuster et al. 2013), as illustrated here by the difference between two climatological estimates for the year 2000 (**Figure 2**). Rödenbeck et al. (2015) provided a detailed comparison of these and 12 additional CO_2 flux estimates.

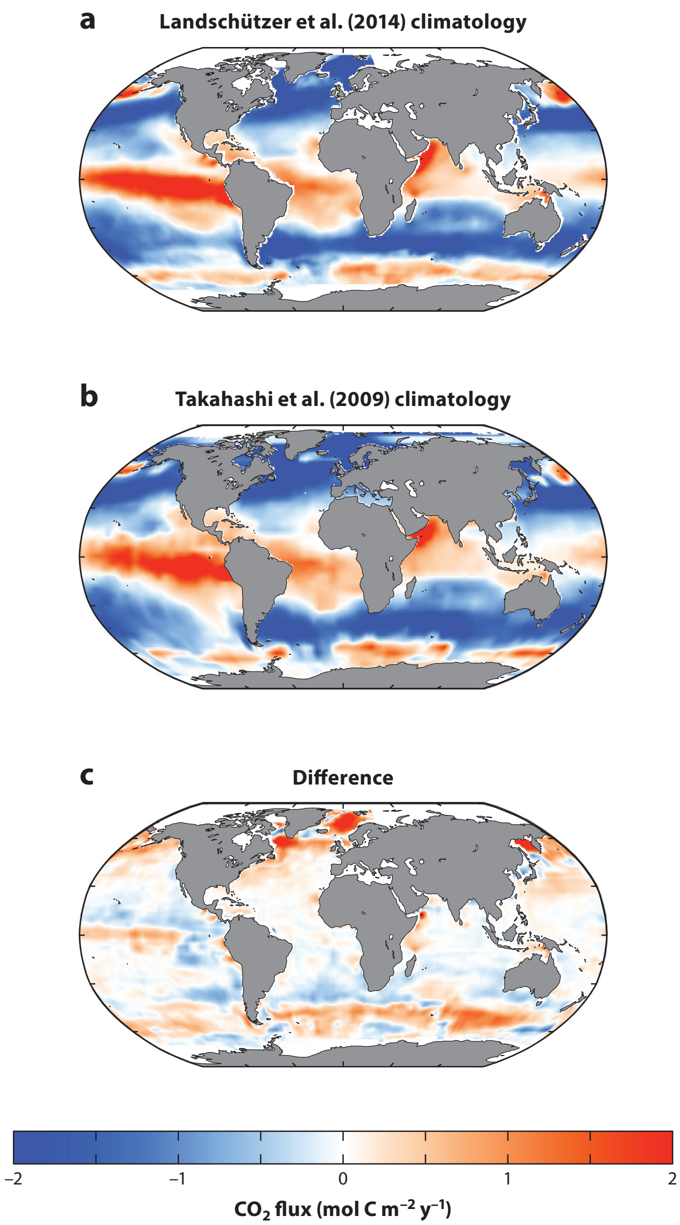


Figure 2

Annual mean CO_2 flux for the year 2000: (*a*) Landschützer et al. (2014) climatology, (*b*) Takahashi et al. (2009) climatology, and (*c*) the difference between the two (panel *a* minus panel *b*).

A further complication is the coastal zone, where carbon fluxes are large and highly variable but observations are few and research limited. The global coastal carbon flux is grossly constrained as a naturally occurring river-borne source of carbon from land to ocean of approximately 0.5 Pg C y^{-1} (Jacobson et al. 2007, Wanninkhof et al. 2013). This review addresses only carbon fluxes in the open ocean, as others have recently provided reviews for the coastal zone (e.g., Benway et al. 2016).

The agreement in December 2015 at the United Nations Framework Convention on Climate Change 21st Conference of the Parties (COP 21) indicates that the global community intends to limit global warming to no more than 2°C above preindustrial levels, and additionally aspires to only 1.5°C of warming. If this is to be achieved, there is little doubt that approaches to remove carbon from the atmosphere—so-called negative emissions—will be required (Fuss et al. 2014, McNutt et al. 2015). The expected need for huge carbon storage reservoirs indicates the oceans will likely need to be part of such efforts. Better knowledge of the regional patterns and temporal variability of ocean carbon uptake is thus needed for assessment and monitoring of such activities.

Finally, the current inability to accurately quantify the mean CO_2 sink regionally or locally (**Figure 2**) also suggests that present-day observational constraints are inadequate to support a detailed, quantitative, and mechanistic understanding of how the ocean carbon sink works and how it is responding to intensifying climate change. This lack of mechanistic understanding implies that our ability to model (Roy et al. 2011, Ciais et al. 2013, Frölicher et al. 2015, Randerson et al. 2015), and thus to project the future ocean carbon sink, including feedbacks caused by warming and other climate change, is seriously limited.

3. BASICS OF THE OCEAN CARBON SINK

The ocean is a sink for CO_2 because the gas is highly soluble in seawater, which in turn is due to its dissociation into ions. This solubility is critical to the ocean's ability to absorb atmospheric CO_2. However, the ocean carbon cycle has a complex natural cycle involving physical, chemical, and biological processes upon which this net uptake is imprinted. These basics are discussed in this section.

The dissociation of CO_2, and thus the partial pressure of CO_2 gas in the surface ocean ($p\mathrm{CO_2}^{\mathrm{ocean}}$), is a function of the total concentration of dissolved inorganic carbon (DIC), alkalinity, sea surface temperature (SST), and sea surface salinity. $p\mathrm{CO_2}^{\mathrm{ocean}}$ increases with increasing DIC, SST, or sea surface salinity and decreases with increasing alkalinity. For detailed discussions of carbon chemistry in seawater, we refer readers elsewhere (Sarmiento & Gruber 2006, Williams & Follows 2011).

The sea-to-air CO_2 flux is proportional to $\Delta p\mathrm{CO_2}$ ($p\mathrm{CO_2}^{\mathrm{ocean}} - p\mathrm{CO_2}^{\mathrm{atm}}$). The rate of flux is set by solubility and the gas-exchange coefficient (k_w), which is a function of turbulence and other boundary-layer complexities and is typically parameterized through a relationship to wind speed (Sarmiento & Gruber 2006, Williams & Follows 2011). However, it is only $\Delta p\mathrm{CO_2}$ that sets the direction and total potential for CO_2 flux: CO_2 flux $= k_w \times$ solubility $\times\ \Delta p\mathrm{CO_2}$. (Throughout this review, the sign convention used is positive fluxes to the atmosphere and negative fluxes to the ocean.) With $p\mathrm{CO_2}^{\mathrm{atm}}$ being relatively homogeneous across the globe, research on global ocean CO_2 flux focuses on understanding the dynamics of $p\mathrm{CO_2}^{\mathrm{ocean}}$.

$p\mathrm{CO_2}^{\mathrm{ocean}}$ is elevated above $p\mathrm{CO_2}^{\mathrm{atm}}$ in major upwelling regions, such as along the eastern boundaries of the basins and in the equatorial zones. Positive $\Delta p\mathrm{CO_2}$ indicates an outgassing CO_2 flux (**Figures 2** and **3*a***). In the western boundary currents and at high latitudes, $p\mathrm{CO_2}^{\mathrm{ocean}}$ is less than $p\mathrm{CO_2}^{\mathrm{atm}}$ on the mean, and thus $\Delta p\mathrm{CO_2}$ is negative and uptake of carbon by the ocean occurs (**Figures 2** and **3*b***).

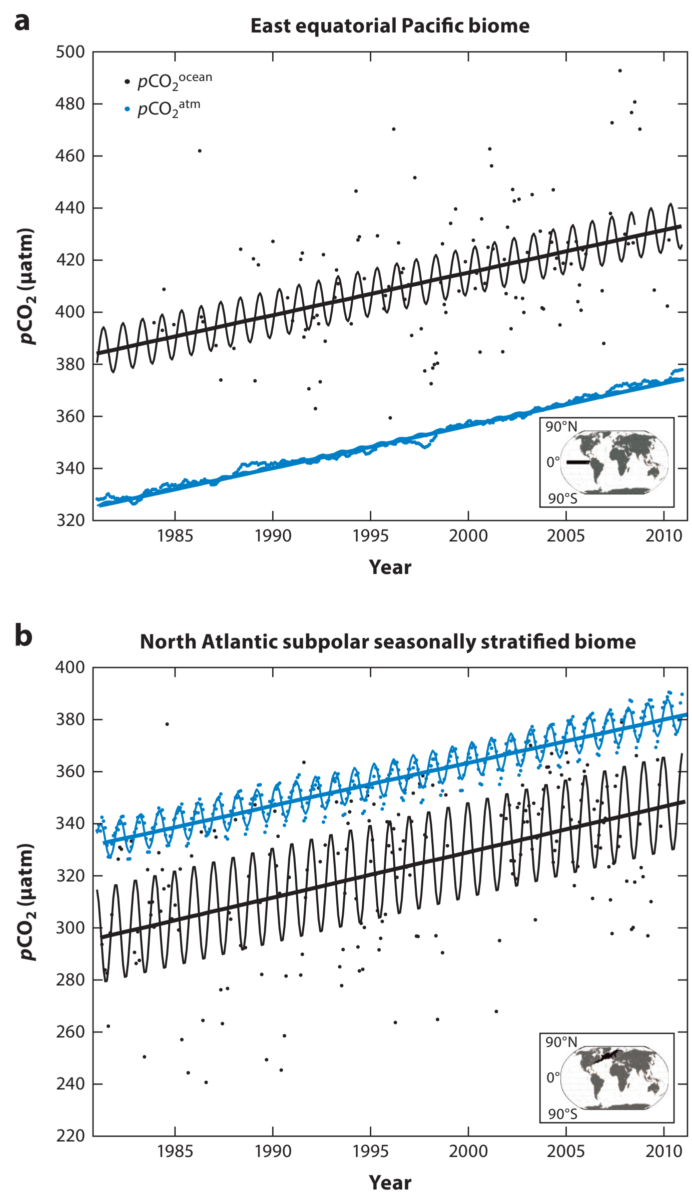


Figure 3

Biome-mean pCO_2^{ocean} data (*black dots*; data from Takahashi et al. 2014), with pCO_2^{atm} (*blue dots*; data from Coop. Glob. Atmos. Data Integr. Proj. 2013) for (*a*) the east equatorial Pacific biome and (*b*) the North Atlantic subpolar seasonally stratified biome. The insets show the location of each biome (*black shading*). The fits for the seasonal harmonics and trends are as in Fay & McKinley (2013).

What causes this pattern of flux to occur? If DIC were an inert tracer, ocean circulation would cause it to become homogeneously mixed from surface to deep water across the global ocean. However, global mean DIC exhibits a vertical gradient with increasing concentrations at depth (**Figure 4**) owing to several pumps that drive additional DIC to the deep ocean. If these pumps

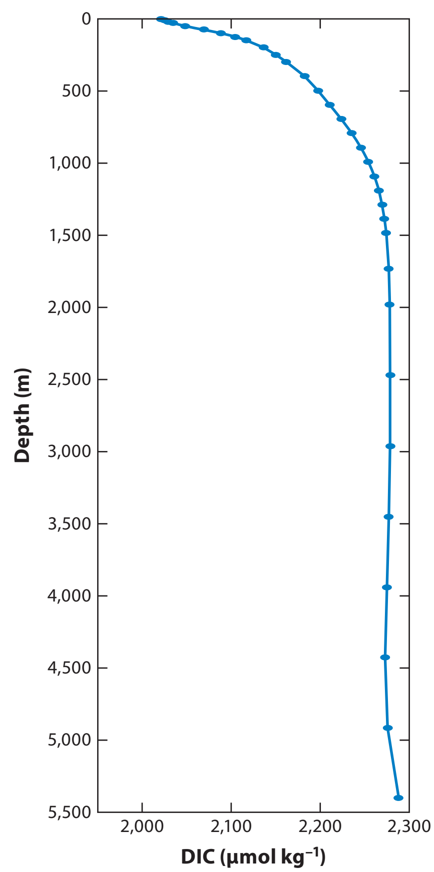


Figure 4

Global area-weighted mean dissolved inorganic carbon (DIC) depth profile from the Global Ocean Data Analysis Project, version 2 (GLODAPv2) (Key et al. 2015).

were not active, surface-ocean DIC would be higher, and pCO_2^{atm} would increase by 50% of the preindustrial value (i.e., +140 ppm) (Gruber & Sarmiento 2002).

If the only process affecting DIC in the ocean were gas exchange, such that saturation were achieved everywhere, DIC concentrations would be greater than 2,000 μmol kg^{-1} across the water column (Williams & Follows 2011). However, the differential solubility of carbon at different temperatures combined with the large-scale overturning circulation of the ocean—the solubility pump—also acts on DIC distribution (Volk & Hoffert 1985). As warm subtropical waters flow toward the poles in boundary currents, they are cooled and increase their capacity to hold CO_2—i.e., pCO_2^{ocean} is suppressed, ΔpCO_2 becomes more negative, and a flux to the ocean occurs (**Figure 2*a*,*b***). These cold, high-latitude waters with high carbon content are then subducted to the deep and intermediate ocean, increasing the mean deep-ocean DIC relative to the mean surface (**Figure 4**).

Biology also contributes to the vertical DIC gradient in the ocean. When nutrients and light are sufficient, phytoplankton in the surface ocean fix DIC into organic compounds via photosynthesis, i.e., the soft-tissue pump. A small portion of the fixed organic carbon escapes to the deep ocean through gravitational sinking of particulate detritus, fecal pellets, and other wastes. Of the total global ocean productivity of 52 Pg C y^{-1}, 10–15% is exported from the surface (Westberry et al. 2008, Siegel et al. 2014). This soft-tissue pump enhances the global mean vertical gradient in DIC by 70% (Sarmiento & Gruber 2006). In addition, the carbonate pump occurs because of organisms that make hard parts out of calcium carbonate ($CaCO_3$). As these organisms use DIC to make $CaCO_3$, they also use Ca^{2+} ions. The two positive charges on the calcium ions mean that for each mole removed, two moles of alkalinity are also used. This decrease of alkalinity results in a higher pCO_2^{ocean}, increased ΔpCO_2, and greater potential for outgassing. Despite this effect of raising pCO_2^{ocean}, the transfer of carbon to depth by the carbonate pump enhances the surface-to-depth DIC gradient by 20% (Sarmiento & Gruber 2006).

In the North Atlantic, atmospheric forcing drives large buoyancy losses, causing deep convective mixing that subducts cold, high-carbon waters to depth. In the equatorial Pacific, easterly winds cause divergence and upwelling that returns cool, high-DIC waters to the surface. In the Southern Ocean, westerly winds drive northward Ekman transport, creating the surface branch of a meridional overturning circulation that is associated with a divergence at approximately 60°S and a convergence at approximately 40°S. Upwelling in the Antarctic divergence exposes waters with high DIC to the atmosphere, elevating the local surface pCO_2^{ocean} relative to the overlying atmosphere, and a CO_2 flux to the atmosphere occurs (**Figure 2**). Mode and intermediate water formation in the convergence region subducts waters with relatively high carbon content into the interior of the ocean. Because these circulations both carry cold, high-carbon waters to depth and return deep waters to the surface after they have accumulated additional biologically sourced carbon, the integrated effect of solubility (Volk & Hoffert 1985), biology, circulation, and gas exchange—together constituting the gas-exchange pump—is to enhance the natural gradient of DIC from surface to depth by only 10% (Sarmiento & Gruber 2006).

Although it is an apparently small net influence on the natural surface-to-depth DIC gradient, the solubility pump component of the gas-exchange pump is the dominant effect driving oceanic uptake of anthropogenic carbon in the modern day. As anthropogenic CO_2 emissions cause pCO_2^{atm} to increase, ΔpCO_2 becomes more negative in regions where is it already negative, and positive ΔpCO_2 is brought closer to zero. Thus, as pCO_2^{atm} increases, regions of uptake absorb more carbon, and regions of net efflux outgas less carbon. In both cases, DIC accumulates in the ocean and pCO_2^{ocean} rises, as seen in the subpolar North Atlantic and equatorial Pacific biomes (Takahashi et al. 2009, Graven et al. 2012, Ciais et al. 2013) (**Figure 4**). On multicentennial timescales, the anthropogenic carbon absorbed in this way will also be returned to the surface by the ocean circulation, and uptake of carbon from the atmosphere will be slowed. However, currently the deep-ocean reservoir from which upwelled water is sourced has essentially no anthropogenic carbon, having come to equilibrium with the preindustrial atmosphere hundreds to thousands of years ago, and thus the uptake of anthropogenic carbon is maximized (Sabine et al. 2004, Khatiwala et al. 2009, DeVries 2014, McKinley et al. 2016).

That the dominant mechanisms of natural ocean carbon storage and anthropogenic carbon storage differ is one indication of the need to separate the natural carbon cycle from the anthropogenically modified carbon cycle in our studies. Although carbon molecules in the ocean are not distinguishable, a variety of approaches have been used to distinguish between carbon that is present in the ocean naturally and carbon that is present only because of human activities (Sabine & Tanhua 2010). The terms natural carbon and anthropogenic carbon, as well as the sum of these, referred to as contemporary carbon, are used throughout the rest of this review (Gruber et al. 2009).

4. METHODS TO CONSTRAIN VARIABILITY AND TRENDS

Surface-ocean pCO_2^{ocean} measurements are collected by dedicated research expeditions and via Volunteer Observing Ship (VOS) efforts. A variety of studies have focused on specific transects by individual research groups (e.g., Corbière et al. 2007). Two groups have undertaken compilation and synthesis efforts for these data, creating the Lamont-Doherty Earth Observatory data set (Takahashi et al. 2014) and the Surface Ocean CO_2 Atlas (Pfeil et al. 2013, Bakker et al. 2014). Although the two data sets differ in length, quality control efforts, and data processing, both offer more than 10 million historical observations of pCO_2^{ocean}.

If ocean circulation and biology change in ways that reduce ocean carbon uptake, pCO_2^{ocean} should increase more quickly than pCO_2^{atm}. In other words, if $dpCO_2^{ocean}/dt > dpCO_2^{atm}/dt$, then $d\Delta pCO_2 dt > 0$, indicating a decline in the sink. Conversely, if change in pCO_2^{ocean} is slower than change in pCO_2^{atm}, or $d\Delta pCO_2 dt < 0$, this indicates growth in the sink (Corbière et al. 2007, Schuster & Watson 2007, Le Quéré et al. 2009, Metzl 2009, Schuster et al. 2009, Takahashi et al. 2009, Metzl et al. 2010). A complication here is that warming reduces CO_2 solubility, which also leads to positive pCO_2^{ocean} trends. In practice, this means that if warming drives some of the pCO_2^{ocean} change, less carbon would be accumulated from the atmosphere.

It is important to note that analysis of trends in pCO_2^{ocean} as compared with trends in pCO_2^{atm} does not indicate quantitatively how much the CO_2 flux has grown ($d\Delta pCO_2 dt < 0$) or declined ($\Delta pCO_2 dt > 0$); it indicates only the direction of change. If the CO_2 flux is to be estimated, interpolation of pCO_2^{ocean} to at least monthly temporal resolution is needed, and in addition, a gas-exchange parameterization and wind-speed product must be selected. As discussed below, such efforts are now becoming more common (Landschützer et al. 2014, Rödenbeck et al. 2015), but they are not currently the primary basis for understanding of variability and trends in the ocean carbon sink. For consistency with the current state of the field, we focus our review on comparison of the direction of trends (growing or declining) rather than on quantitative estimates of flux change.

A major challenge across many pCO_2^{ocean} studies is the lack of temporal and spatial coherence as well as the different spatial scales they analyze. McKinley et al. (2011) and Fay & McKinley (2013) began to address these issues by (*a*) assessing carbon sink trends over gyre-scale ocean biomes, a scale relevant to global partitioning of carbon between the atmosphere and the ocean (e.g., Gruber et al. 2009, Canadell et al. 2011), and (*b*) avoiding the prescription of specific break points for changing trends by looking at all reasonable combinations of start and end years. Temperature and chemical influences on pCO_2^{ocean} change are also separated.

Available pCO_2^{ocean} data have been interpolated by various methods to gridded estimates of monthly pCO_2^{ocean} and contemporary CO_2 flux at every location across the globe, based on extrapolation of pCO_2^{ocean} with other complementary data. The Surface Ocean pCO_2 Mapping Intercomparison project compares multiple independently created estimates (Rödenbeck et al. 2015). All approaches inherently have uncertainties, as suggested by the mean flux comparison in **Figure 2**. For example, two different approaches based on the same Surface Ocean CO_2 Atlas pCO_2^{ocean} data result in substantially different dominant timescales of variability in the Southern Ocean (Landschützer et al. 2015).

The other primary approach to assess carbon sink variability and trends in the surface ocean is to use numerical models that simulate the regional or global circulation and associated carbon biogeochemistry. These have often been used in hindcast mode, where they are forced with observed fluxes of momentum, heat, and freshwater from the atmosphere for recent decades (Le Quéré et al. 2000, 2009, 2010; McKinley et al. 2004, 2006; Lovenduski et al. 2007, 2015; Bennington et al. 2009; Ullman et al. 2009; Long et al. 2013; Breeden & McKinley 2016). Earth system models—numerical representations of the coupled atmosphere, ocean, and biosphere system—are necessary to make projections of future change and variability in the ocean carbon sink (Roy et al. 2011, Frölicher et al. 2015, Randerson et al. 2015, Resplandy et al. 2015, McKinley et al. 2016).

The best long-term constraint on the cumulative ocean carbon sink is to observe the change of the carbon inventory of the interior ocean. The first global survey was the World Ocean Circulation Experiment/Joint Global Ocean Flux Study (WOCE/JGOFS) in the 1990s, and from these data, the first estimates of the cumulative uptake of anthropogenic carbon from the ocean were made (Sabine et al. 2004). Subsequently, these hydrographic surveys have been reoccupied on approximately decadal timescales, and empirical multiple linear regression approaches have been used to estimate anthropogenic carbon accumulation between section occupations (Sabine & Tanhua 2010, Wanninkhof et al. 2010, Ciais et al. 2013, Khatiwala et al. 2013). Although these studies have focused on the deep ocean, where the significant seasonal variability at the surface should have limited impact, significant variability attributable to circulation and biological variability remains (Wanninkhof et al. 2010). As these studies focus on multiyear change, the reasonable, though imperfect, assumption of water-mass equilibrium with CO_2 and other atmospheric gases at the time of subduction to the deep ocean is made.

As a complement to direct analyses of hydrographic data, a variety of inverse approaches have been developed that constrain steady-state ocean circulation estimates with hydrographic tracer data, and then use the resulting flow to estimate the propagation of anthropogenic carbon into the ocean interior (Waugh et al. 2006; Mikaloff Fletcher et al. 2006; Jacobson et al. 2007; Gruber et al. 2009; Khatiwala et al. 2009, 2013; DeVries 2014). As these inverse approaches evolve, they offer increasingly less dependence on uncertain circulation estimates from hindcast ocean models. However, a remaining, widely recognized concern with these studies is their assumption of steady-state ocean circulation since preindustrial times.

5. OCEAN CARBON SINK VARIABILITY AND TRENDS AT REGIONAL TO GLOBAL SCALES

The IPCC found that the ocean carbon sink has changed from -2.0 ± 0.7 Pg C y^{-1} for 1980–1989 to -2.3 ± 0.7 Pg C y^{-1} for 2000–2009, or -0.15 Pg C y^{-1} decade^{-1} in the last two decades (Ciais et al. 2013). By contrast, based only on in situ pCO_2^{ocean} data for 1972–2005, Takahashi et al. (2009) found that pCO_2^{ocean} increased at the same rate as pCO_2^{atm} (1.5 μatm y^{-1}) and thus that the global sink has remained constant. This difference can be resolved by the divergent time frames and methodologies and by the fact that uncertainty in the IPCC mean decadal estimate (± 0.7 Pg C y^{-1}) does not strictly preclude a steady sink from 1980 to 2009.

In their contribution to the Regional Carbon Cycle Assessment and Processes (RECCAP) synthesis effort, Wanninkhof et al. (2013) assessed global-scale surface CO_2 fluxes from a variety of methods. Their best estimate of mean ocean carbon uptake was -2.0 ± 0.6 Pg C y^{-1} for 1990–2009, with interannual variability of 0.2 Pg C y^{-1} based on ocean models and an empirical approach (Park et al. 2010). Although all approaches suggest an increasing sink since the 1960s, there is not detailed agreement on the magnitude of the trend from the 1990s to the 2000s (-0.13 to -0.50 Pg C y^{-1} decade^{-1}). Studies along individual hydrographic sections consistently illustrate a positive trend in carbon storage over the past several decades, but limited sampling and strong variability remain significant challenges (Wanninkhof et al. 2010, Ciais et al. 2013). Comparing across hydrographic estimates and inverse approaches, Khatiwala et al. (2013) found agreement on global mean uptake but substantial differences in spatial distributions and rates of anthropogenic carbon storage.

Although the dominant mechanism for a growing sink is increasing pCO_2^{atm}, climate change may be modifying ocean physics so as to limit growth of the sink. Modeling studies suggest that warming, increased stratification, and changes in circulation caused by anthropogenic climate change have modified the ocean carbon sink and will continue to do so (Le Quéré et al. 2010, Roy et al. 2011, Frölicher et al. 2015, Randerson et al. 2015, McKinley et al. 2016). However, these impacts to date have been relatively small (∼5% reduction of the potential growth of the sink) (Frölicher et al. 2015).

Observations and models indicate that the largest variability in CO_2 fluxes occurs in the North Atlantic, the North Pacific, the equatorial Pacific, and the Southern Ocean. These are the regions with greatest influence on globally integrated flux variability, and therefore we discuss them in detail below. For discussion of other regions, we refer readers to a paper by Fay & McKinley (2013).

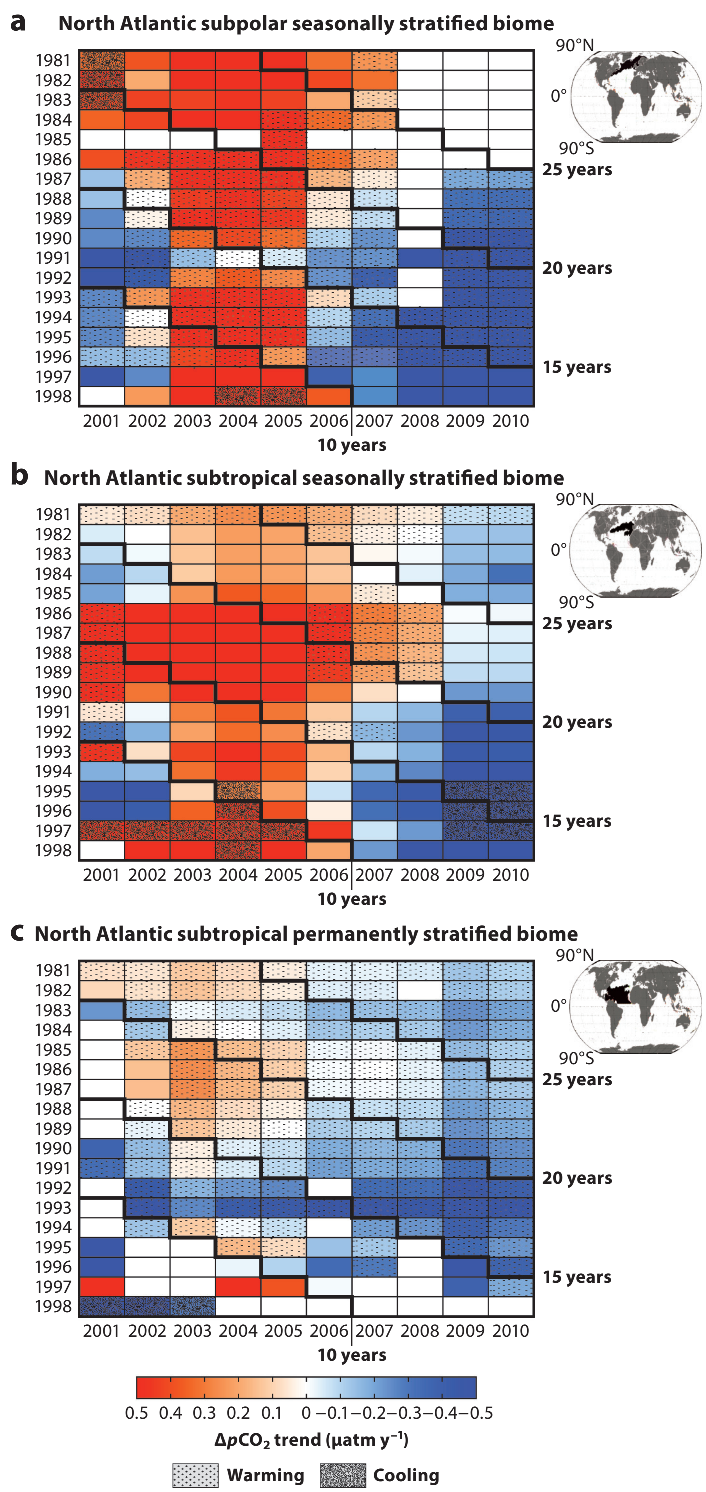
5.1. The North Atlantic

The North Atlantic is a particularly critical region for ocean carbon uptake, as it is the most intense (per unit area) anthropogenic CO_2 sink (Sabine et al. 2004; Khatiwala et al. 2009, 2013; DeVries 2014). This is due primarily to the large equator-to-pole heat transport that occurs here (Ganachaud & Wunsch 2000), which means that as warm waters move northward and lose heat, cooling occurs that lowers pCO_2^{ocean} and leads to an intense carbon uptake (**Figure 2**). As described below, understanding of CO_2 sink variability from both observations and models is complicated by complex geography, vigorous current systems, strong climactic forcing, sparse sampling, and the differing time frames used in independent analyses.

In their RECCAP contribution, Schuster et al. (2013) provided a best estimate of the 1990–2009 mean uptake of 0.47 ± 0.08 Pg C y^{-1} for 18–76°N. This best estimate is taken from a pCO_2^{ocean} flux climatology (Takahashi et al. 2009) and ocean-interior inversions (Gruber et al. 2009), with the mean across ocean models and atmospheric inversions providing substantially lower (0.30 ± 0.04 Pg C y^{-1}) and higher (0.59 ± 0.06 Pg C y^{-1}) uptake estimates, respectively. This region's complicated surface circulation affects carbon fluxes (Follows & Williams 2004, Thomas et al. 2008) and complicates comparisons among studies.

Corbière et al. (2007) and Metzl et al. (2010) considered the change in ocean carbon uptake in the subpolar gyre in winter using pCO_2^{ocean} observed along a VOS line between Iceland and Newfoundland. They found a reduced sink from 1993 to 2003 resulting from advection of warmer waters into the region, and a more rapid decline from 2001 to 2008 resulting from enhanced convection that increased DIC and reduced alkalinity. Using VOS pCO_2^{ocean} in the subpolar North Atlantic, several studies estimated large declines in the strength of the CO_2 sink for 1990–2006 (Schuster & Watson 2007, Schuster et al. 2009) and for 1997–2008 (Landschützer et al. 2013). By contrast, hindcast models have suggested an increasing or approximately neutral sink in this region from the mid-1990s to the mid-2000s (Thomas et al. 2008, Ullman et al. 2009) and connected these changes to a declining phase of the NAO. An SST-based pCO_2^{ocean} diagnostic also showed strong NAO-driven variability in this period (Park et al. 2010). With the basin-scale, multi-time-frame approach for analysis of pCO_2^{ocean} observations (McKinley et al. 2011, Fay & McKinley 2013) (**Figure 5**), opposing views about the recent directions of trends can largely be resolved. For time series starting throughout the 1980s and 1990s in the subpolar gyre and ending in the mid-2000s, data indicate temporary elevation of pCO_2^{ocean} trends above pCO_2^{atm} trends, but these signals are not sustained when end years in the late 2000s are considered. For the 1990s to years past 2008, the subpolar gyre of the North Atlantic has been a growing sink, indicated by negative trends in ΔpCO_2 (**Figure 5*a***). The data are insufficient to reveal biome-scale trends from the 1980s to the late 2000s.

A long-term growing sink in the subpolar gyre from the 1990s to the late 2000s is consistent with recently modeled multidecadal timescale variability in the natural carbon cycle. For 1948–2009, large-scale patterns of subpolar North Atlantic pCO_2^{ocean} variability are strongly influenced by the Atlantic Multidecadal Oscillation (AMO) in a regional hindcast model (Breeden


a North Atlantic subpolar seasonally stratified biome
b North Atlantic subtropical seasonally stratified biome
c North Atlantic subtropical permanently stratified biome
1981 1982 1983 1984 1985 1986 1987 1988 1989 1990 1991 1992 1993 1994 1995 1996 1997 1998
2001 2002 2003 2004 2005 2006 2007 2008 2009 2010
10 years
15 years
20 years
25 years
90°N
0°
90°S
0.5 0.4 0.3 0.2 0.1 0 −0.1 −0.2 −0.3 −0.4 −0.5
ΔpCO_2 trend (μatm y^{-1})
Warming
Cooling

& McKinley 2016). During a low AMO phase, such as the 1970s and 1980s, cool SSTs allow for deep convection that increases DIC supply to the surface and reduces carbon uptake even though cooling increases carbon solubility. In the high AMO phase of the 2000s, reduced solubility with warmer SSTs is similarly overwhelmed by reduced convection, and thus more carbon is absorbed. AMO connections to the Atlantic meridional overturning circulation have been postulated but not confirmed (Delworth & Mann 2000, Booth et al. 2012).

Moving to the southeast, in the northeastern Atlantic from approximately 35°N to 50°N, a declining sink between the 1980s or 1990s and the mid-2000s has been suggested (Schuster & Watson 2007; Le Quéré et al. 2009, 2010; Schuster et al. 2009; Watson et al. 2009). Modeling also suggests a declining carbon sink in this region for this period and connects this change to a shift of the NAO from strongly positive in the mid-1990s to relatively neutral in the mid-2000s. Thomas et al. (2008) found that the NAO shift drove a reduction in the transfer of low-DIC subtropical waters along the North Atlantic Current, enhancing the positive trend in pCO_2^{ocean} for 1980–2004. The basin-scale, multi-time-frame approach of McKinley et al. (2011) and Fay & McKinley (2013) also indicated a few years of reduced sink in the mid-2000s and then a return to a steady or slightly growing sink as time series extend from the 1980s to the late 2000s (**Figure 5*b***). Although ΔpCO_2 trends are in many cases slightly negative, they indicate a steady sink due to confidence intervals on pCO_2^{ocean} trends (see caption for **Figure 5*b***).

There has been a slightly increasing sink in the western subtropical gyre from the 1990s to the mid-2000s (Schuster & Watson 2007; Le Quéré et al. 2009, 2010; Schuster et al. 2009; Ullman et al. 2009). McKinley et al. (2011) and Fay & McKinley (2013) also found an increasing sink for the 1990s to the mid-to-late 2000s, but a steady sink given uncertainty for longer time series extending from the 1980s to the late 2000s (**Figure 5*c***). NAO-associated variability caused by changing convection has been found in the subtropical gyre, particularly at the Bermuda Atlantic Time-Series Study site (Gruber et al. 2002, McKinley et al. 2004).

Across the North Atlantic, from the subtropical to subpolar gyre, warming trends have a statistically significant impact on pCO_2^{ocean} trends for time series ending in 2001 or later (**Figure 5**). This is consistent with the influence of the increasing AMO index since the 1980s, combined with some influence of anthropogenic warming (Ting et al. 2009, Breeden & McKinley 2016). Because warming reduces carbon solubility, the anthropogenic component of this SST trend is consistent with a carbon-climate feedback whereby climate warming reduces ocean carbon uptake (Le Quéré et al. 2010, Roy et al. 2011). More data are needed for additional confirmation.

The weight of evidence from observational analyses and modeling is that climate variability on timescales from interannual to multidecadal strongly modulates ocean carbon uptake in the

Figure 5

Trends in ΔpCO_2 for (*a*) the North Atlantic subpolar seasonally stratified biome, (*b*) the North Atlantic subtropical seasonally stratified biome, and (*c*) the North Atlantic subtropical permanently stratified biome. The insets show the location of each biome (*black shading*). Red colors indicate $d\Delta pCO_2 dt > 0$ and a declining sink; blue colors indicate $d\Delta pCO_2 dt < 0$ and a growing sink; light shading indicates a sink change statistically indistinguishable from zero (intervals specified by biome below), i.e., a steady sink. White cells indicate that the data are insufficient to define the biome-mean trend (Fay & McKinley 2013). The patterns overlaid on the shading indicate statistically significant temperature trends as indicated in the legend. The average confidence intervals are 0.55 μatm y^{-1} (10–20 years) and 0.20 μatm y^{-1} (20–30 years) for panel *a*, 0.39 μatm y^{-1} (10–20 years) and 0.23 μatm y^{-1} (20–30 years) for panel *b*, and 0.34 μatm y^{-1} (10–20 years) and 0.21 μatm y^{-1} (20–30 years) for panel *c*. The average confidence intervals for atmospheric trends are 0.08 μatm y^{-1} for all time series longer than 10 years. pCO_2 data are from Takahashi et al. (2014).

North Atlantic. In the context of this variability, data are not yet sufficient to clearly reveal long-term trends.

5.2. The North Pacific

The North Pacific is a region strongly influenced by the PDO and has benefited from denser pCO_2^{ocean} sampling than in most other regions (Takahashi et al. 2006, 2009). Models have also been used to assess large-scale mechanisms of variability. The most comprehensive studies are summarized below.

In their contribution to RECCAP, Ishii et al. (2014) provided a best estimate of the 1990–2009 mean uptake of -0.47 ± 0.13 Pg C y^{-1} for 18–60°N in the North Pacific. This estimate was taken from pCO_2^{ocean} diagnostic models and ocean-interior inversions, while hindcast ocean models and atmospheric inversions are consistent within the uncertainty. In the North Pacific, carbon uptake is strongest in the subtropical-to-subarctic transition zone owing to the effect of cooling on pCO_2^{ocean} (Takahashi et al. 2002) and the lateral geostrophic advection of surface DIC (Ayers & Lozier 2012). The region is a strong sink in winter and a slight source in summer (Takahashi et al. 2009, Ayers & Lozier 2012).

Natural variability generates a range of up to 0.3 Pg C y^{-1} in the extratropical North Pacific carbon sink (Landschützer et al. 2014). Models illustrate that decadal timescale variability in air-sea CO_2 fluxes in the North Pacific is modulated by the PDO through influences on SST and DIC supply to the surface with winter convective mixing (McKinley et al. 2006, Valsala et al. 2012). This is consistent with observational evidence of an east–west seesaw pattern in surface DIC attributed to changes in SST and mixed-layer depth associated with the PDO (Yasunaka et al. 2014). ENSO also drives variability in the southwestern and northeastern regions of the North Pacific (Midorikawa et al. 2005, Wong et al. 2010), suggesting that the degree to which climate modes impact carbon cycle variability is dependent on the spatial and temporal scale (Xiu & Chai 2014).

Takahashi et al. (2006) assessed observed pCO_2^{ocean} trends from 1970 to 2004 in 32 10° × 10° boxes of the North Pacific. For the open ocean, they found that trends in pCO_2^{ocean} were indistinguishable from those in pCO_2^{atm} and that warming has not been a significant influence. They concluded that the carbon sink has been steady for 35 years. Fay & McKinley (2013) found no trend in ΔpCO_2 on similar time frames but negative ΔpCO_2 trends for time series ending after 2005. These negative ΔpCO_2 trends are linked to a shift from a positive phase of the PDO in the 1980s and 1990s to a negative phase in the 2000s. This shift brought cooler temperatures and lower pCO_2^{ocean} to the subpolar North Pacific and an increase in ocean carbon uptake.

5.3. The Equatorial Pacific

As discussed herein, the modulation of the circulation of the equatorial Pacific by ENSO leads to the dominant signal of globally integrated ocean CO_2 flux variability. As such, it has been the focus of multiple observational and modeling studies.

The contemporary mean flux out of the equatorial Pacific was 0.44 ± 0.41 Pg C y^{-1} for 1990–2009 over 18°S–18°N (Ishii et al. 2014). This best estimate is taken from pCO_2^{ocean} diagnostic models and ocean-interior inversions used in RECCAP, with hindcast ocean models and atmospheric inversions in agreement given uncertainties. In the equatorial Pacific, small mean CO_2 fluxes occur in the stratified western region, caused by low wind speeds and near equilibrium between ocean and atmosphere carbon. This contrasts with stronger outgassing fluxes in the eastern equatorial Pacific cold tongue region (Ishii et al. 2014), where wind-driven equatorial divergence brings DIC-rich waters to the surface (**Figures 2*a*** and **4**).

ENSO dominates the interannual variability of air-sea CO_2 flux in the equatorial Pacific (Feely et al. 2006, Ishii et al. 2009, Sutton et al. 2014). Model-based estimates of the regionally integrated variability range from 0.15 to 0.20 Pg C y^{-1} (1σ annual mean) (Le Quéré et al. 2000, McKinley et al. 2004, Long et al. 2013). During warm El Niño phases, reduced upwelling reduces the quantity of high-DIC waters reaching the surface, and less outgassing occurs; during cool La Niña phases, there is more upwelling and more carbon outgassing. Because these changes occur coherently over this vast ocean region, ENSO is the strongest coherent signal of variability in the globally integrated ocean CO_2 sink across a variety of methods, including hindcast models, Earth system models, diagnostic models, and atmospheric inversions (Le Quéré et al. 2000, 2010; McKinley et al. 2004; Park et al. 2010; Long et al. 2013; Wanninkhof et al. 2013; Landschützer et al. 2014; Resplandy et al. 2015; Rödenbeck et al. 2015).

Recently, moored observations have offered unique high-frequency pCO_2^{ocean} measurements between 170°W and 120°W (Niño 3.4 region) (Sutton et al. 2014). These data indicate positive ΔpCO_2 trends over the period 1997–2011, i.e., increased outgassing from the ocean, which is consistent with surface pCO_2^{ocean} trends (Fay & McKinley 2013). Sutton et al. (2014) attributed this trend to intensified wind-driven DIC and an increased frequency of La Niña events, both of which are associated with a PDO regime shift after the 1997–1998 El Niño.

5.4. The Southern Ocean

As summarized below, the Southern Ocean has been subject to intense study with respect to its variability and multidecadal change in CO_2 uptake. Studies in the early 2000s relied on numerical models, given that data were extremely sparse. The lack of data has been somewhat ameliorated since the mid-2000s, which is leading to further evolution of understanding with respect to Southern Ocean CO_2 sink variability and change.

Gruber et al. (2009) estimated a contemporary mean uptake of -0.34 ± 0.20 Pg C y^{-1} for the Southern Ocean region south of 44°S in 2000, which is consistent with the pCO_2^{ocean}-based climatological estimate of -0.30 ± 0.17 Pg C y^{-1} from Takahashi et al. (2009) when regionally integrated. However, the mean flux distributions within the Southern Ocean are quite different between these two approaches.

Studies using interior data indicate that the Southern Ocean is where the majority (~40%) of the ocean uptake of anthropogenic carbon has occurred to date (Sabine et al. 2004; Khatiwala et al. 2009, 2013; DeVries 2014). The Earth system models of the Coupled Model Intercomparison Project Phase 5 (CMIP5) suite also simulate this pattern and allow for the uptake to be attributed almost completely to increasing pCO_2^{atm} (Frölicher et al. 2015). Long-term increasing rates of anthropogenic carbon accumulation in the Southern Ocean are indicated by inverse studies (Khatiwala et al. 2013, DeVries 2014).

Interannual variability in Southern Ocean carbon uptake has been examined through the lens of multiple hindcast modeling studies (Wetzel et al. 2005; Lenton & Matear 2007; Lovenduski et al. 2007, 2013, 2015; Verdy et al. 2007; Wang & Moore 2012; Hauck et al. 2013; Lenton et al. 2013). Variability of CO_2 flux on these timescales has been linked to the SAM, a large-scale mode of climate variability in the Southern Hemisphere that drives variability in the position and intensity of the westerly winds (Thompson & Wallace 2000). In these modeling studies, positive phases of the SAM are associated with increased upwelling of DIC-rich water to the high-latitude surface, increasing pCO_2^{ocean} and temporarily reducing the carbon sink in this region (Lovenduski et al. 2007).

On multidecadal timescales, coarse-resolution ($\Delta x \approx 100$ km) hindcast modeling studies have found a decreasing efficiency of the Southern Ocean carbon sink over the past few decades, driven by a trend toward positive phases of the SAM (Le Quéré et al. 2007, Lovenduski et al. 2008). In these models, a positive trend in the SAM drives stronger westerly winds and increased upwelling of DIC-rich water to the surface over multidecadal timescales, reducing the efficacy of the Southern Ocean carbon sink. However, studies have questioned whether coarse-resolution models can adequately capture the response of sub-grid-scale eddies to increasing wind stress in the Southern Ocean (Böning et al. 2008, Hogg et al. 2008). These studies argue that the wind-driven increase in meridional overturning (and thus the upwelling of DIC-rich water) may be partially compensated by opposing trends in eddy-driven overturning, dampening the CO_2 flux response to trends in the SAM. Gent (2016) demonstrated that the effect of eddies can be reproduced in coarse-resolution models, provided that the eddy parameterization coefficient is variable. Using coarse-resolution models with variable eddy coefficients, Lovenduski et al. (2013) and Swart et al. (2014) found smaller changes in the Southern Ocean carbon sink in response to increasing wind stress, owing to increased eddy-induced advection of DIC out of the Southern Ocean surface.

Direct observations of pCO_2^{ocean} in the Southern Ocean are limited, and there is a lack of consensus in the literature about changes in the Southern Ocean carbon sink from these observations, with some studies pointing toward a decreasing sink over the past few decades (Metzl 2009; Takahashi et al. 2009, 2012) and others finding an unchanging carbon sink over this same time period (Fay & McKinley 2013). Fay et al. (2014) showed that carbon uptake slowed from 1990 to 2006 and subsequently strengthened. Landschützer et al. (2015) drew a similar conclusion from interpolated pCO_2^{ocean} data, and Munro et al. (2015) confirmed the strengthening of the carbon sink after 2002 in a study of year-round pCO_2^{ocean} data collected in the Drake Passage. Together, observationally based studies have suggested large and poorly constrained decadal variability in the Southern Ocean carbon sink. As the data sets have been extended, it has also become increasingly clear that not all of this variability can be linked to the SAM (Park et al. 2010, Fay & McKinley 2013, Landschützer et al. 2015). Lovenduski et al. (2015) used output from a hindcast model to investigate the length of the time series that would be necessary to detect changes in the Southern Ocean carbon sink from observations. They found that 13–17 years of monthly surface data would be required, which has not yet been achieved. These results argue for maintaining the nearly continuous under-way pCO_2 sampling in the Drake Passage region that began in 2002 (Sprintall et al. 2012).

5.5. Variability in the Biological Pump

The soft-tissue and carbonate pumps are critical to the natural gradient of DIC (see Section 3), and thus, change in ocean productivity is a potential mechanism to drive regional changes in ocean carbon uptake. Some analyses of satellite data have revealed large-scale declines in productivity from 1998 to the early 2000s (Gregg et al. 2005, Behrenfeld et al. 2006), with the proposed mechanism being increased stratification caused by climate warming. This mechanism remains under debate (Lozier et al. 2011, Behrenfeld et al. 2016). Ocean biogeochemical models indicate that the trends observed to date show decadal variability and cannot be attributed to global climate change (Henson et al. 2010). Projections of the future impacts of climate warming on ocean productivity generally suggest reduced subtropical and increased high-latitude productivity (Sarmiento et al. 2004, Bopp et al. 2013). Moreover, links between changes in surface productivity and carbon storage at depth have not been observationally established for the modern day (Siegel et al. 2014). In the North Atlantic, modeling and satellite data suggest a limited connection between interannual changes in productivity and carbon drawdown (Bennington et al. 2009). Better understanding productivity and carbon export to depth is a current research priority (Siegel et al. 2016).

5.6. Summary

The sum of the available evidence indicates that variability in the ocean carbon sink is significant and is driven primarily by physical processes of upwelling, convection, and advection. Despite evidence for a growing sink when globally integrated (Khatiwala et al. 2009, 2013; Ciais et al. 2013; DeVries 2014), this variability, combined with sparse sampling, means that it is not yet possible to directly confirm from surface observations that long-term growth in the oceanic sink is occurring. Because the approaches used to infer the growing sink are forced to assume a steady-state ocean circulation, it is critical to confirm that increases in oceanic uptake are actually occurring in the context of the variable ocean circulation.

At regional scales, modes of climate variability create coherent large-scale patterns of variability in CO_2 fluxes. Globally integrated variability fluctuates with ENSO. Yet, at regional scales outside the equatorial Pacific, these modes tend to explain less than 20% of the large-scale variance in $p\mathrm{CO_2}^{\mathrm{ocean}}$ and CO_2 flux (McKinley et al. 2004, 2006; Breeden & McKinley 2016), indicating that much variance remains undescribed. Consistent with the limited amount of variance explained, the mechanistic connections of these modes are not well understood, except in the equatorial Pacific with ENSO. In the North Atlantic, a variety of studies have suggested a connection of the NAO and AMO to $p\mathrm{CO_2}^{\mathrm{ocean}}$ and CO_2 fluxes, but whether these changes occur through convection or advection remains an open question. In the Southern Ocean, the SAM has been linked to $p\mathrm{CO_2}^{\mathrm{ocean}}$ and CO_2 fluxes through impacts on wind-driven ventilation and subduction; however, since the mid-2000s, the clear relationship to SAM has substantially weakened (Fay & McKinley 2013, Landschützer et al. 2015). In the North Pacific, the relative influence of the PDO as opposed to ENSO requires further study. Particularly as observations in the high latitudes have become more abundant, evidence has grown that climate modes do not adequately explain carbon cycle variability and that mechanistic understanding of carbon sink variability requires substantial additional elucidation.

6. DISTINGUISHING VARIABILITY FROM TRENDS

Direct detection of the growing ocean carbon sink at regional scales is critical so that we can validate inverse estimates that assume a steady-state ocean circulation and improve the mechanisms represented in numerical models. Furthermore, although climate modes are associated with some of the large-scale variance in the carbon sink, they are incomplete descriptors. Thus, a quantitative assessment of the magnitude of variability versus the growing trend is needed. Here, we use a large ensemble of a single Earth system model that allows study of internal variability without confounding effects from structural differences across different Earth system models (Lovenduski et al. 2016, McKinley et al. 2016).

The large ensemble of the Community Earth System Model (CESM-LE) (Kay et al. 2015) simulates variability and change in the ocean carbon cycle in recent decades and through 2100. CESM is a state-of-the-art coupled climate model consisting of atmosphere, ocean, land surface, and sea ice components. The CESM-LE experiment described here included 32 members with ocean biogeochemistry output (Long et al. 2013) and a control integration of more than 2,000 years. A transient integration (ensemble member 1) started at year 402 and was integrated for 251 years under historical forcing (1850–2005) and then the IPCC representative concentration pathway 8.5 scenario for 2006–2100. Additional ensemble members were initialized from ensemble member 1 at January 1, 1920, with round-off level perturbations applied to the air temperature field. For $p\mathrm{CO_2}^{\mathrm{ocean}}$ and CO_2 flux, the mean fluxes and variability were consistent with observations (McKinley et al. 2016).

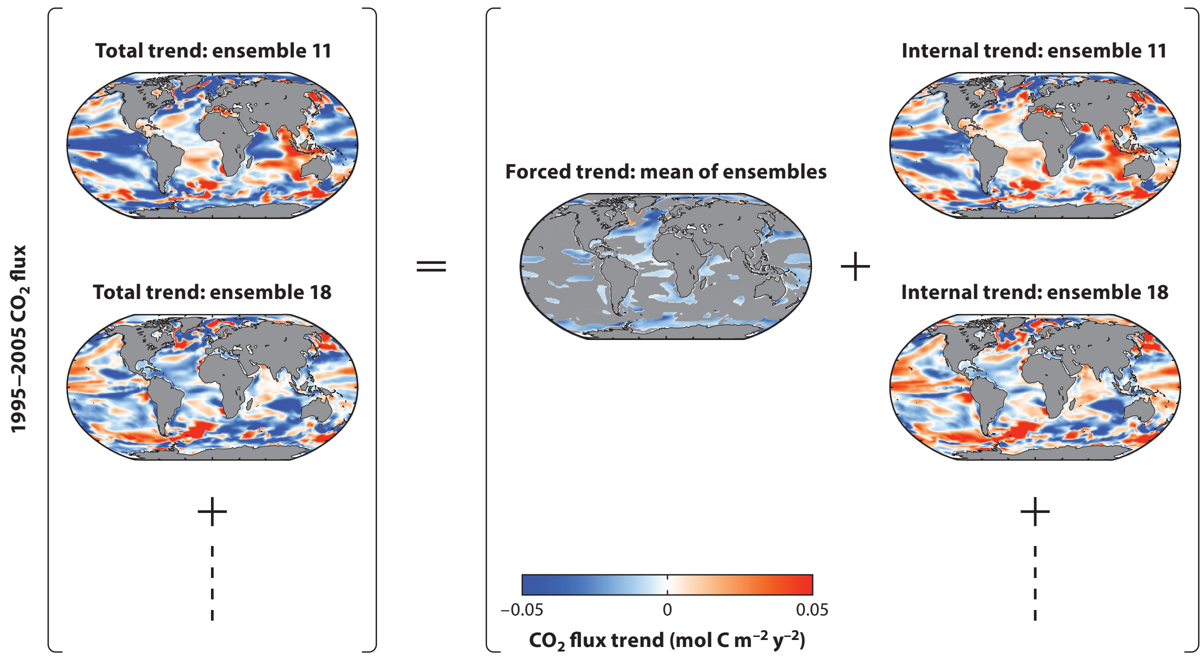


Figure 6

Decomposition of CO_2 flux trends from the large ensemble of the Community Earth System Model (CESM-LE) into internal, forced, and total trends for 1995–2005. The left column shows the trend of two representative ensemble members (11 and 18), the middle column shows the forced trend (mean of all ensemble members), and the right column shows the internal trend for ensemble members 11 and 18 (total ensemble trend minus forced trend).

Each ensemble member responds to the anthropogenic and natural forcing during the historical period and future anthropogenic forcing. Forcing leads to changes that are common across all ensemble members; this is the forced trend. Each ensemble member also develops a unique pattern of variability resulting from the inherent internal variability of the modeled climate system; these are the internal trends unique to each ensemble member. The forced trend is calculated as the mean trend across the ensemble, and the internal trend for each ensemble is the difference between the forced trend and the total trend for each ensemble member.

The change in CO_2 flux over 10 years (1995–2005) gives the total trend (left column of **Figure 6**) and is due almost entirely to the internal variability of each ensemble (right column of **Figure 6**), with only limited contributions from the forced trend (middle column of **Figure 6**). In most regions, the forced trends in CO_2 flux are too small to be statistically significant. For the 30-year change (1985–2015) however, internal variability has a lesser impact, and the forced trend begins to emerge from the variability (**Figure 7**). The forced 30-year trend indicates a growing carbon sink, consistent with the IPCC synthesis (Ciais et al. 2013).

Consistent with the discussion of the prior section, this CESM-LE analysis further illustrates that variability in CO_2 flux is large and sufficient to prevent detection of anthropogenic trends in ocean carbon uptake on decadal timescales (**Figure 6**). McKinley et al. (2016) used the same approach to illustrate where and when detection would be possible if data were sufficient to constrain annual mean fluxes since 1990 for each 1 × 1 grid point across the globe. The earliest

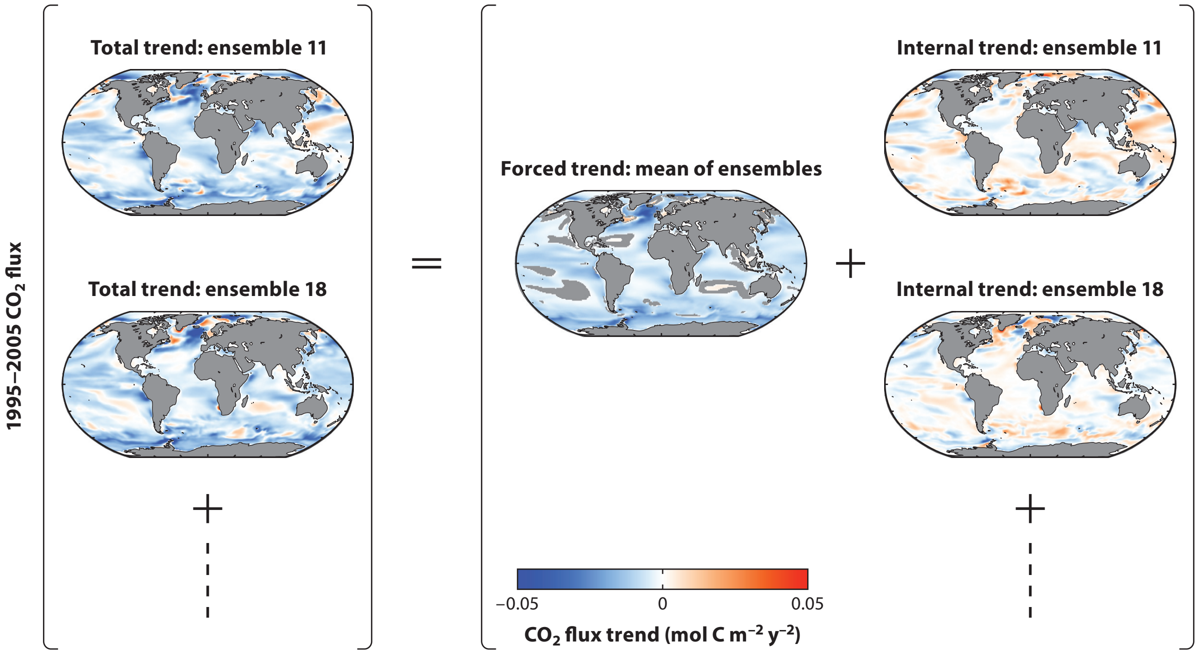


Figure 7

Decomposition of CO_2 flux trends from the large ensemble of the Community Earth System Model (CESM-LE) into internal, forced, and total trends for 1985–2015. The left column shows the trend of two representative ensemble members (11 and 18), the middle column shows the forced trend (mean of all ensemble members), and the right column shows the internal trend for ensemble members 11 and 18 (total ensemble trend minus forced trend).

detection of trends would occur in regions where they are largest: in the Southern Ocean and in parts of the North Atlantic and North Pacific (2010–2020) (middle column of **Figure 7**). In the equatorial Pacific, anthropogenic trends are large, but variability is also large, making for intermediate detection timescales (2030–2050). The longest detection timescales are in the subtropics (2050+) owing to small or negligible increases in the CO_2 sink in these regions. Because annual mean fluxes since 1990 are poorly constrained, except at the few ocean time-series stations (Bates et al. 2014), detection of forced trends will likely be later than summarized here at most locations across the globe. These results emphasize that long-term observations are critical for monitoring the changing ocean carbon cycle.

SUMMARY POINTS

1. The underlying chemistry and basic mechanisms of the ocean carbon sink are well understood. Biological mechanisms are critical to the ocean's natural ability to sequester carbon in the deep ocean, but the ocean carbon cycle's response to the anthropogenic perturbation of rapid growth in pCO_2^{atm} is dominated by the solubility effect as ΔpCO_2 becomes more negative.

2. The El Niño–Southern Oscillation is the dominant mode for global CO_2 flux interannual variability. Extratropical variability is associated with modes of climate variability (the Southern Annular Mode, North Atlantic Oscillation, Atlantic Multidecadal Oscillation, and Pacific Decadal Oscillation), but there is significant additional unexplained variability. The mechanisms of extratropical variability, whether associated with climate modes or not, are incompletely elucidated.
3. The globally integrated ocean CO_2 sink appears to have grown over the past several decades. Inverse approaches using interior data and an assumed steady-state ocean circulation have found that the most rapid growth of the sink occurs in the Southern Ocean. However, data from the surface ocean have not yet confirmed growth in the carbon sink. This lack of direct detection is attributable to both large physical variability and sparse sampling.
4. A novel approach to Earth system modeling, the large ensemble, indicates that because the Southern Ocean and North Atlantic forced trends in carbon uptake are large, they could be detectable above the noise of variability in the current decade.

FUTURE ISSUES

1. The most accurate approach to constraining the long-term evolution of the ocean carbon sink is the use of high-quality, full-depth observations of dissolved inorganic carbon and alkalinity. The World Ocean Circulation Experiment/Joint Global Ocean Flux Study (WOCE/JGOFS) program of the 1990s was a groundbreaking effort that first allowed for closure of the global carbon budget (Sabine et al. 2004). Without quantification of carbon uptake by the ocean, net fluxes from the terrestrial biosphere would remain unconstrained (Le Quéré et al. 2015). This work is continuing with the current Global Ocean Ship-Based Hydrographic Investigations Program (GO-SHIP) (Talley et al. 2016; **http://www.go-ship.org**) and will need to be sustained for long-term monitoring of the response of the global carbon cycle to the anthropogenic perturbation.
2. pCO_2^{ocean} observations from research and Volunteer Observing Ship programs offer additional invaluable information with respect to CO_2 fluxes and their temporal variability. Current efforts to continually quality control and synthesize these data are critical for a growing mechanistic understanding (Pfeil et al. 2013, Bakker et al. 2014). These data can be combined with other data, particularly from satellites, to better understand global space-time variability (Rödenbeck et al. 2015).
3. Autonomous (Martz et al. 2015, Johnson et al. 2016) and moored (Sutton et al. 2014) platforms offer rapidly expanding observational opportunities with unprecedented temporal and spatial coverage. Shipboard calibration of autonomous sensors will continue to be an absolute necessity. Continued technological development is needed to fully constrain the carbonate system from these platforms.
4. Processes critical to understanding and predicting the ocean carbon sink remain poorly understood. For example, the most intense (per unit area) sink for anthropogenic carbon occurs in the subpolar North Atlantic, and how the sink is modified by variability in the meridional overturning circulation (Lozier 2012) is not observationally constrained.

5. The land-ocean carbon continuum also remains largely unconstrained (Regnier et al. 2013, Laruelle et al. 2014). For a full understanding of the ocean carbon cycle, fluxes in the coastal zone will need substantially more research and monitoring attention (Benway et al. 2016).
6. Large ensembles of climate models offer a new opportunity to clearly distinguish anthropogenic trends from internal variability (McKinley et al. 2016). As additional large ensembles become available, comparisons across their estimates for forced trends and internal variability will be illuminating.

DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

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LITERATURE CITED

Archer D, Eby M, Brovkin V, Ridgwell A, Cao L, et al. 2009. Atmospheric lifetime of fossil fuel carbon dioxide. *Annu. Rev. Earth Planet. Sci.* 37:117–34

Ayers JM, Lozier MS. 2012. Unraveling dynamical controls on the North Pacific carbon sink. *J. Geophys. Res. Oceans* 117:C01017

Bakker DCE, Pfeil B, Smith K, Hankin S, Olsen A, et al. 2014. An update to the Surface Ocean CO_2 Atlas (SOCAT version 2). *Earth Syst. Sci. Data* 6:69–90

Bates N, Astor Y, Church M, Currie K, Dore J, et al. 2014. A time-series view of changing ocean chemistry due to ocean uptake of anthropogenic CO_2 and ocean acidification. *Oceanography* 27(1):126–41

Behrenfeld MJ, O'Malley RT, Boss ES, Westberry TK, Graff JR, et al. 2016. Revaluating ocean warming impacts on global phytoplankton. *Nat. Clim. Change* 6:323–30

Behrenfeld MJ, O'Malley RT, Siegel DA, McClain CR, Sarmiento JL, et al. 2006. Climate-driven trends in contemporary ocean productivity. *Nature* 444:752–55

Bennington V, McKinley GA, Dutkiewicz S, Ullman D. 2009. What does chlorophyll variability tell us about export and air-sea CO_2 flux variability in the North Atlantic? *Glob. Biogeochem. Cycles* 23:GB3002

Benway HM, Alin SR, Boyer E, Cai WJ, Coble PG, et al. 2016. *A science plan for carbon cycle research in North American coastal waters*. Rep. Coast. Carbon Synth. (CCARS) Community Workshop, Woods Hole, MA, Aug. 19–21. doi: 10.1575/1912/7777

Böning C, Dispert A, Visbeck M, Rintoul S, Schwarzkopf F. 2008. The response of the Antarctic Circumpolar Current to recent climate change. *Nat. Geosci.* 1:864–69

Booth BBB, Dunstone NJ, Halloran PR, Andrews T, Bellouin N. 2012. Aerosols implicated as a prime driver of twentieth-century North Atlantic climate variability. *Nature* 484:228–32

Bopp L, Resplandy L, Orr JC, Doney SC, Dunne JP, et al. 2013. Multiple stressors of ocean ecosystems in the 21st century: projections with CMIP5 models. *Biogeosciences* 10:6225–45

Breeden M, McKinley GA. 2016. Climate impacts on multidecadal $p$$CO_2$ variability in the North Atlantic: 1948–2009. *Biogeosciences* 13:3387–96

Canadell JG, Ciais P, Gurney K, Le Quéré C, Piao S, et al. 2011. An international effort to quantify regional carbon fluxes. *Eos Trans. AGU* 92:81–82

Ciais P, Dolman AJ, Bombelli A, Duren R, Peregon A, et al. 2014. Current systematic carbon-cycle observations and the need for implementing a policy-relevant carbon observing system. *Biogeosciences* 11:3547–3602

Ciais P, Sabine C, Bala G, Bopp L, Brovkin V, et al. 2013. Carbon and other biogeochemical cycles. In *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, ed. TF Stocker, D Qin, G-K Plattner, M Tignor, SK Allen, et al., pp. 465–570. Cambridge, UK: Cambridge Univ. Press

Coop. Glob. Atmos. Data Integr. Proj. 2013. *Multi-laboratory compilation of synchronized and gap-filled atmospheric carbon dioxide records for the period 1979–2012 (obspack_co2_1_GLOBALVIEW-CO2_2013_v1.0.4_2013–12–23)*. Compiled by NOAA Glob. Monit. Div., Boulder, CO. Data product accessed at **http://dx.doi.org/10.3334/OBSPACK/1002**

Corbière A, Metzl N, Reverdin G, Brunet C, Takahashi T. 2007. Interannual and decadal variability of the oceanic carbon sink in the North Atlantic subpolar gyre. *Tellus B* 59:168–78

Delworth TL, Mann ME. 2000. Observed and simulated multidecadal variability in the Northern Hemisphere. *Clim. Dyn.* 16:661–76

DeVries T. 2014. The oceanic anthropogenic CO_2 sink: storage, air-sea fluxes, and transports over the industrial era. *Glob. Biogeochem. Cycles* 28:631–47

Fay AR, McKinley GA. 2013. Global trends in surface ocean $p$$CO_2$ from in situ data. *Glob. Biogeochem. Cycles* 27:541–57

Fay AR, McKinley GA, Lovenduski NS. 2014. Southern Ocean carbon trends: sensitivity to methods. *Geophys. Res. Lett.* 41:6833–40

Feely RA, Takahashi T, Wanninkhof R, McPhaden MJ, Cosca CE, et al. 2006. Decadal variability of the air-sea CO_2 fluxes in the equatorial Pacific Ocean. *J. Geophys. Res. Oceans* 111:C08S90

Follows M, Williams RG. 2004. Mechanisms controlling the air-sea flux of CO_2 in the North Atlantic. In *The Ocean Carbon Cycle and Climate*, ed. M Follows, T Oguz, pp. 217–49. Dordrecht, Neth.: Springer

Frölicher TL, Sarmiento JL, Paynter DJ, Dunne JP, Krasting JP, et al. 2015. Dominance of the Southern Ocean in anthropogenic carbon and heat uptake in CMIP5 models. *J. Clim.* 28:862–86

Fuss S, Canadell JG, Peters GP, Tavoni M, Andrew RM. 2014. Betting on negative emissions. *Nat. Clim. Change* 4:850–53

Ganachaud A, Wunsch C. 2000. Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. *Nature* 408:453–57

Gent PR. 2016. Effects of Southern Hemisphere wind changes on the meridional overturning circulation in ocean models. *Annu. Rev. Mar. Sci.* 8:79–94

Graven HD, Gruber N, Key R, Khatiwala S, Giraud X. 2012. Changing controls on oceanic radiocarbon: new insights on shallow-to-deep ocean exchange and anthropogenic CO_2 uptake. *J. Geophys. Res.* 117:C10005

Gregg WW, Casey NW, McClain CR. 2005. Recent trends in global ocean chlorophyll. *Geophys. Res. Lett.* 32:L03606

Gruber N, Gloor M, Mikaloff Fletcher SE, Doney SC, Dutkiewicz S, et al. 2009. Oceanic sources, sinks, and transport of atmospheric CO_2. *Glob. Biogeochem. Cycles* 23:GB1005

Gruber N, Keeling CD, Bates NR. 2002. Interannual variability in the North Atlantic Ocean carbon sink. *Science* 298:2374–78

Gruber N, Sarmiento JL. 2002. Large-scale biogeochemical-physical interactions in elemental cycles. In *The Sea*, Vol. 12: *Biological-Physical Interactions in the Sea*, ed. AR Robinson, JJ McCarthy, BJ Rothschild, pp. 337–99. New York: Wiley & Sons

Hauck J, Völker C, Wang T, Hoppema M, Losch M, Wolf-Gladrow DA. 2013. Seasonally different carbon flux changes in the Southern Ocean in response to the southern annular mode. *Glob. Biogeochem. Cycles* 27:1236–45

Henson SA, Sarmiento JL, Dunne JP, Bopp L, Lima ID, et al. 2010. Detection of anthropogenic climate change in satellite records of ocean chlorophyll and productivity. *Biogeosciences* 7:621–40

Hogg AMC, Meredith MP, Blundell JR, Wilson C. 2008. Eddy heat flux in the Southern Ocean: response to variable wind forcing. *J. Clim.* 21:608–20

Ishii M, Feely RA, Rodgers KB, Park GH, Wanninkhof R, et al. 2014. Air-sea CO_2 flux in the Pacific Ocean for the period 1990–2009. *Biogeosciences* 11:709–34

Ishii M, Inoue HY, Midorikawa T, Saito S, Tokieda T, et al. 2009. Spatial variability and decadal trend of the oceanic CO_2 in the western equatorial Pacific warm/fresh water. *Deep-Sea Res. II* 56:591–606

Jacobson AR, Mikaloff Fletcher SE, Gruber N, Sarmiento JL, Gloor M. 2007. A joint atmosphere-ocean inversion for surface fluxes of carbon dioxide: 1. Methods and global-scale fluxes. *Glob. Biogeochem. Cycles* 21:GB1019

Johnson KS, Jannasch HW, Coletti LJ, Elrod VA, Martz TR, et al. 2016. Deep-sea DuraFET: a pressure tolerant pH sensor designed for global sensor networks. *Anal. Chem.* 88:3249–56

Kay JE, Deser C, Phillips A, Mai A, Hannay C, et al. 2015. The Community Earth System Model (CESM) large ensemble project: a community resource for studying climate change in the presence of internal climate variability. *Bull. Am. Meteorol. Soc.* 96:1333–49

Key RM, Olsen A, van Heuven S, Lauvset SK, Velo A, et al. 2015. *Global Ocean Data Analysis Project, version 2 (GLODAPv2)*. ORNL/CDIAC-162, NDP-P093, Carbon Dioxide Inf. Anal. Cent., Oak Ridge Natl. Lab., US Dep. Energy, Oak Ridge, TN. **http://cdiac.ornl.gov/oceans/GLODAPv2**

Khatiwala S, Primeau F, Hall T. 2009. Reconstruction of the history of anthropogenic CO_2 concentrations in the ocean. *Nature* 462:346–49

Khatiwala S, Tanhua T, Mikaloff Fletcher SE, Gerber M, Doney SC, et al. 2013. Global ocean storage of anthropogenic carbon. *Biogeosciences* 10:2169–91

Landschützer P, Gruber N, Bakker DCE, Schuster U. 2014. Recent variability of the global ocean carbon sink. *Glob. Biogeochem. Cycles* 28:927–49

Landschützer P, Gruber N, Bakker DCE, Schuster U, Nakaoka S, et al. 2013. A neural network-based estimate of the seasonal to inter-annual variability of the Atlantic Ocean carbon sink. *Biogeosciences* 10:7793–815

Landschützer P, Gruber N, Haumann FA, Rödenbeck C, Bakker DC, et al. 2015. The reinvigoration of the Southern Ocean carbon sink. *Science* 349:1221–24

Laruelle GG, Lauerwald R, Pfeil B, Regnier P. 2014. Regionalized global budget of the CO_2 exchange at the air-water interface in continental shelf seas. *Glob. Biogeochem. Cycles* 28:1199–214

Le Quéré C, Moriarty R, Andrew RM, Peters GP, Ciais P, et al. 2015. Global carbon budget 2014. *Earth Syst. Sci. Data* 7:47–85

Le Quéré C, Orr JC, Monfray P, Aumont O, Madec G. 2000. Interannual variability of the oceanic sink of CO_2 from 1979 through 1997. *Glob. Biogeochem. Cycles* 14:1247–65

Le Quéré C, Raupach MR, Canadell JG, Marland G, Bopp L, et al. 2009. Trends in the sources and sinks of carbon dioxide. *Nat. Geosci.* 2:831–36

Le Quéré C, Rodenbeck C, Buitenhuis ET, Conway TJ, Langenfelds R, et al. 2007. Saturation of the Southern Ocean CO_2 sink due to recent climate change. *Science* 316:1735–38

Le Quéré C, Takahashi T, Buitenhuis ET, Rödenbeck C, Sutherland SC. 2010. Impact of climate change and variability on the global oceanic sink of CO_2. *Glob. Biogeochem. Cycles* 24:GB4007

Lenton A, Matear RJ. 2007. Role of the Southern Annular Mode (SAM) in Southern Ocean CO_2 uptake. *Glob. Biogeochem. Cycles* 21:GB2016

Lenton A, Tilbrook B, Law R, Bakker D, Doney SC, et al. 2013. Sea-air CO_2 fluxes in the Southern Ocean for the period 1990–2009. *Biogeosci. Discuss.* 10:285–333

Long MC, Lindsay K, Peacock S, Moore JK, Doney SC. 2013. Twentieth-century oceanic carbon uptake and storage in CESM1(BGC). *J. Clim.* 26:6775–800

Lovenduski NS, Fay AR, McKinley GA. 2015. Observing multidecadal trends in Southern Ocean CO_2 uptake: What can we learn from an ocean model? *Glob. Biogeochem. Cycles* 29:416–26

Lovenduski NS, Gruber N, Doney SC. 2008. Towards a mechanistic understanding of the decadal trends in the Southern Ocean carbon sink. *Glob. Biogeochem. Cycles* 22:GB3016

Lovenduski NS, Gruber N, Doney SC, Lima ID. 2007. Enhanced CO_2 outgassing in the Southern Ocean from a positive phase of the Southern Annular Mode. *Glob. Biogeochem. Cycles* 21:GB2026

Lovenduski NS, Long MC, Gent PR, Lindsay K. 2013. Multi-decadal trends in the advection and mixing of natural carbon in the Southern Ocean. *Geophys. Res. Lett.* 40:139–42

Lovenduski NS, McKinley GA, Fay AR, Lindsay K, Long MC. 2016. Partitioning uncertainty in ocean carbon uptake projections: internal variability, emission scenario, and model structure. *Glob. Biogeochem. Cycles*. In revision. doi: 10.1002/2016GB005426

Lozier MS. 2012. Overturning in the North Atlantic. *Annu. Rev. Mar. Sci.* 4:291–315

Lozier MS, Dave AC, Palter JB, Gerber LM, Barber RT. 2011. On the relationship between stratification and primary productivity in the North Atlantic. *Geophys. Res. Lett.* 38:L18609

Martz TR, Daly KL, Byrne RH, Stillman JH, Turk D. 2015. Technology for ocean acidification research: needs and availability. *Oceanography* 28(2):40–47

McKinley GA, Carlson CA, Andrews A, Brown DJ, Romero-Lankao P, Shrestha G. 2015. Managing the carbon cycle requires strong science. *Eos*, Nov. 25. **http://eos.org/opinions/managing-the-carbon-cycle-requires-strong-science**

McKinley GA, Fay AR, Takahashi T, Metzl N. 2011. Convergence of atmospheric and North Atlantic carbon dioxide trends on multidecadal timescales. *Nat. Geosci.* 4:606–10

McKinley GA, Pilcher DJ, Fay AR, Lindsay K, Long MC, Lovenduski NS. 2016. Timescales for detection of trends in the ocean carbon sink. *Nature* 530:469–72

McKinley GA, Rödenbeck C, Gloor M, Houweling S, Heimann M. 2004. Pacific dominance to global air-sea CO_2 flux variability: a novel atmospheric inversion agrees with ocean models. *Geophys. Res. Let.* 31:GL22308

McKinley GA, Takahashi T, Buitenhuis E, Chai F, Christian JR, et al. 2006. North Pacific carbon cycle response to climate variability on seasonal to decadal timescales. *J. Geophys. Res.* 111:C07S06

McNutt MK, Abdalati W, Caldeira K, Doney SC, Falkowski PG, et al. 2015. *Climate Intervention: Carbon Dioxide Removal and Reliable Sequestration*. Washington, DC: Natl. Res. Counc.

Metzl N. 2009. Decadal increase of oceanic carbon dioxide in Southern Indian Ocean surface waters (1991–2007). *Deep-Sea Res. II* 56:607–19

Metzl N, Corbière A, Reverdin G, Lenton A, Takahashi T, et al. 2010. Recent acceleration of the sea surface fCO_2 growth rate in the North Atlantic subpolar gyre (1993–2008) revealed by winter observations. *Glob. Biogeochem. Cycles* 24:GB4004

Midorikawa T, Nemoto K, Kamiya H, Ishii M, Inoue HY. 2005. Persistently strong oceanic CO_2 sink in the western subtropical North Pacific. *Geophys. Res. Lett.* 32:L05612

Mikaloff Fletcher SE, Gruber N, Jacobson AJ, Doney SC, Dutkiewicz S, et al. 2006. Inverse estimates of anthropogenic CO_2 uptake, transport, and storage by the ocean. *Glob. Biogeochem. Cycles* 20:GB2002

Munro DR, Lovenduski NS, Takahashi T, Stephens BB, Newberger T, Sweeney C. 2015. Recent evidence for a strengthening CO_2 sink in the Southern Ocean from carbonate system measurements in the Drake Passage (2002–2015). *Geophys. Res. Lett.* 42:7623–30

Park GH, Wanninkhof R, Doney SC, Takahashi T, Lee K, et al. 2010. Variability of global net sea-air CO_2 fluxes over the last three decades using empirical relationships. *Tellus B* 62:352–68

Pfeil B, Olsen A, Bakker DCE, Hankin S, Koyuk H, et al. 2013. A uniform, quality controlled Surface Ocean CO_2 Atlas (SOCAT). *Earth Syst. Sci. Data* 5:125–43

Randerson JT, Lindsay K, Munoz E, Fu W, Moore JK, et al. 2015. Multicentury changes in ocean and land contributions to the climate-carbon feedback. *Glob. Biogeochem. Cycles* 29:744–59

Regnier P, Friedlingstein P, Ciais P, Mackenzie FT, Gruber N, et al. 2013. Anthropogenic perturbation of the carbon fluxes from land to ocean. *Nat. Geosci.* 6:597–607

Resplandy L, Séférian R, Bopp L. 2015. Natural variability of CO_2 and O_2 fluxes: What can we learn from centuries-long climate models simulations? *J. Geophys. Res. Oceans* 120:384–404

Rödenbeck C, Bakker DC, Gruber N, Iida Y, Jacobson AR, et al. 2015. Data-based estimates of the ocean carbon sink variability—first results of the Surface Ocean pCO_2 Mapping intercomparison (SOCOM). *Biogeosciences* 12:7251–78

Roy T, Bopp L, Gehlen M, Schneider B, Cadule P, et al. 2011. Regional impacts of climate change and atmospheric CO_2 on future ocean carbon uptake: a multimodel linear feedback analysis. *J. Clim.* 24:2300–18

Sabine CL, Feely R, Gruber N, Key R, Lee K, et al. 2004. The oceanic sink for anthropogenic CO_2. *Science* 305:367–71

Sabine CL, Tanhua T. 2010. Estimation of anthropogenic CO_2 inventories in the ocean. *Annu. Rev. Mar. Sci.* 2:175–98

Sarmiento JL, Gruber N. 2006. Carbon cycle, CO_2, and climate. In *Ocean Biogeochemical Dynamics*, pp. 392–457. Princeton, NJ: Princeton Univ. Press

Sarmiento JL, Slater R, Barber R, Bopp L, Doney SC, et al. 2004. Response of ocean ecosystems to climate warming. *Glob. Biogeochem. Cycles* 18:GB3003

Schuster U, McKinley GA, Bates N, Chevallier F, Doney SC, et al. 2013. An assessment of the Atlantic and Arctic sea-air CO_2 fluxes, 1990–2009. *Biogeosciences* 10:607–27

Schuster U, Watson AJ. 2007. A variable and decreasing sink for atmospheric CO_2 in the North Atlantic. *J. Geophys. Res. Oceans* 112:C11006

Schuster U, Watson AJ, Bates NR, Corbière A, Gonzalez-Davila M, et al. 2009. Trends in North Atlantic sea-surface fCO_2 from 1990 to 2006. *Deep-Sea Res. II* 56:620–29

Siegel DA, Buesseler KO, Behrenfeld MJ, Benitez-Nelson CR, Boss E, et al. 2016. Prediction of the export and fate of global ocean net primary production: the EXPORTS science plan. *Front. Mar. Sci.* 3:22

Siegel DA, Buesseler KO, Doney SC, Sailley SF, Behrenfeld MJ, Boyd PW. 2014. Global assessment of ocean carbon export by combining satellite observations and food-web models. *Glob. Biogeochem. Cycles* 28:181–96

Sprintall J, Chereskin TK, Sweeney C. 2012. High-resolution underway upper ocean and surface atmospheric observations in Drake Passage: synergistic measurements for climate science. *Oceanography* 25(3):70–81

Sutton AJ, Feely RA, Sabine CL, McPhaden MJ, Takahashi T, et al. 2014. Natural variability and anthropogenic change in equatorial Pacific surface ocean pCO_2 and pH. *Glob. Biogeochem. Cycles* 28:131–45

Swart NC, Fyfe JC, Saenko OA, Eby M. 2014. Wind-driven changes in the ocean carbon sink. *Biogeosciences* 11:6107–17

Takahashi T, Sutherland SC, Feely RA, Wanninkhof R. 2006. Decadal change of the surface water pCO_2 in the North Pacific: a synthesis of 35 years of observations. *J. Geophys. Res. Oceans* 111:C07S05

Takahashi T, Sutherland SC, Kozyr A. 2014. *Global Surface pCO2 (LDEO) Database V2013*. ORNL/CDIAC-159, NDP-088, Carbon Dioxide Inf. Anal. Cent., Oak Ridge Natl. Lab., US Dep. Energy, Oak Ridge, TN. **http://cdiac.ornl.gov/oceans/LDEO_Underway_Database**

Takahashi T, Sutherland SC, Sweeney C, Poisson A, Metzl N, et al. 2002. Global sea–air CO_2 flux based on climatological surface ocean pCO_2, and seasonal biological and temperature effects. *Deep-Sea Res. II* 49:1601–22

Takahashi T, Sutherland SC, Wanninkhof R, Sweeney C, Feely RA, et al. 2009. Climatological mean and decadal changes in surface ocean pCO_2, and net sea–air CO_2 flux over the global oceans. *Deep-Sea Res. II* 56:554–77

Takahashi T, Sweeney C, Hales B, Chipman D, Newberger T, et al. 2012. The changing carbon cycle in the Southern Ocean. *Oceanography* 25(3):26–37

Talley LD, Feely RA, Sloyan BM, Wanninkhof R, Baringer MO, et al. 2016. Changes in ocean heat, carbon content, and ventilation: a review of the first decade of GO-SHIP global repeat hydrography. *Annu. Rev. Mar. Sci.* 8:185–215

Thomas H, Friederike Prowe AE, Lima ID, Doney SC, Wanninkhof R, et al. 2008. Changes in the North Atlantic Oscillation influence CO_2 uptake in the North Atlantic over the past 2 decades. *Glob. Biogeochem. Cycles* 22:GB4027

Thompson DW, Wallace JM. 2000. Annular modes in the extratropical circulation. Part I: month-to-month variability. *J. Clim.* 13:1000–16

Ting M, Kushnir Y, Seager R, Li C. 2009. Forced and internal twentieth-century SST trends in the North Atlantic. *J. Clim.* 22:1469–81

Ullman DJ, McKinley GA, Bennington V, Dutkiewicz S. 2009. Trends in the North Atlantic carbon sink: 1992–2006. *Glob. Biogeochem. Cycles* 23:GB4011

Valsala V, Maksyutov S, Telszewski M, Nakaoka S, Nojiri Y, et al. 2012. Climate impacts on the structures of the North Pacific air-sea CO_2 flux variability. *Biogeosciences* 9:477–92

Verdy A, Dutkiewicz S, Follows MJ, Marshall J, Czaja A. 2007. Carbon dioxide and oxygen fluxes in the Southern Ocean: mechanisms of interannual variability. *Glob. Biogeochem. Cycles* 21:GB2020

Volk T, Hoffert MI. 1985. Ocean carbon pumps: analysis of relative strengths and efficiencies in ocean-driven atmospheric CO_2 changes. In *The Carbon Cycle and Atmospheric CO: Natural Variations Archean to Present*, ed. ET Sundquist, WS Broecker, pp. 99–110. Washington, DC: Am. Geophys. Union

Wang S, Moore JK. 2012. Variability of primary production and air-sea CO_2 flux in the Southern Ocean. *Glob. Biogeochem. Cycles* 26:GB1008

Wanninkhof R, Doney SC, Bullister JL, Levine NM, Warner M, Gruber N. 2010. Detecting anthropogenic CO_2 changes in the interior Atlantic Ocean between 1989 and 2005. *J. Geophys. Res. Oceans* 115:C11028

Wanninkhof R, Park GH, Takahashi T, Sweeney C, Feely R, et al. 2013. Global ocean carbon uptake: magnitude, variability and trends. *Biogeosciences* 10:1983–2000

Watson AJ, Schuster U, Bakker DC, Bates NR, Corbière A, et al. 2009. Tracking the variable North Atlantic sink for atmospheric CO_2. *Science* 326:1391–93

Waugh DW, Hall TM, McNeil BI, Key RM, Matear R. 2006. Anthropogenic CO_2 in the oceans estimated using transit-time distributions. *Tellus B* 58:376–79

Westberry T, Behrenfeld MJ, Siegel DA, Boss E. 2008. Carbon-based primary productivity modeling with vertically resolved photoacclimation. *Glob. Biogeochem. Cycles* 22:GB2024

Wetzel P, Winguth A, Maier-Reimer E. 2005. Sea-to-air CO_2 flux from 1948 to 2003: a model study. *Glob. Biogeochem. Cycles* 19:GB2005

Williams RG, Follows MJ. 2011. *Ocean Dynamics and the Carbon Cycle: Principles and Mechanisms*. Cambridge, UK: Cambridge Univ. Press

Wong CS, Christian JR, Wong SKE, Page J, Xie L, Johannessen S. 2010. Carbon dioxide in surface seawater of the eastern North Pacific Ocean (Line P), 1973–2005. *Deep-Sea Res. I* 57:687–95

Xiu P, Chai F. 2014. Connections between physical, optical and biogeochemical processes in the Pacific Ocean. *Prog. Oceanogr.* 122:30–53

Yasunaka S, Nojiri Y, Nakaoka SI, Ono T, Mukai H, Usui N. 2014. North Pacific dissolved inorganic carbon variations related to the Pacific Decadal Oscillation. *Geophys. Res. Lett.* 41:1005–11