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Key Points:
- Observations and model simulations of ocean anthropogenic carbon assume different start dates
- Once referenced to the same period, 1971–1995, models and observations of ocean anthropogenic carbon agree to within 4%
- A model bias in the mean position of Southern Hemisphere westerlies results in a bias in the pattern of Southern Hemisphere carbon uptake

Supporting Information:
- Supporting Information S1

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Agreement of CMIP5 Simulated and Observed Ocean Anthropogenic CO_2 Uptake

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Abstract Previous studies found large biases between individual observational and model estimates of historical ocean anthropogenic carbon uptake. We show that the largest bias between the Coupled Model Intercomparison Project phase 5 (CMIP5) ensemble mean and between two observational estimates of ocean anthropogenic carbon is due to a difference in start date. After adjusting the CMIP5 and observational estimates to the 1791–1995 period, all three carbon uptake estimates agree to within 3 Pg of C, about 4% of the total. The CMIP5 ensemble mean spatial bias compared to the observations is generally smaller than the observational error, apart from a negative bias in the Southern Ocean and a positive bias in the Southern Indian and Pacific Oceans compensating each other in the global mean. This dipole pattern is likely due to an equatorward and weak bias in the position of Southern Hemisphere westerlies and lack of mode and intermediate water ventilation.

1. Introduction

Since the start of the industrial revolution, human activity has caused atmospheric CO_2 levels to rise. During this time, the ocean has absorbed roughly one third of emitted anthropogenic carbon (C_{ant}) (Khatiwala et al., 2013). Ocean carbon uptake therefore influences how much CO_2 remains in the atmosphere, driving global warming. Accurately measuring and simulating ocean carbon storage is important for assessing the current climate and projecting future climates, and efforts such as the Global Carbon project have been made to constrain fluxes in the carbon cycle (Le Quere et al., 2016), yet uncertainties remain. The reliability of models and observations should be thoroughly examined to understand both the shortcomings and strengths of these tools.

Uncertainties in observational estimates of ocean anthropogenic carbon storage are of the order of 25% (Frölicher et al., 2015). We use the Coupled Model Intercomparison Project phase 5 (CMIP5) Earth System models (ESMs) to simulate carbon uptake and climate change. Frölicher et al. (2015) showed that ESMs from CMIP5 disagree in their estimates of ocean anthropogenic carbon in the year 1995 by as much as 20%. There are also large discrepancies between observational estimates of ocean anthropogenic carbon uptake from different methods: Sabine et al. (2004), hereafter referred to as Sabine et al. (2004), estimate the ocean absorbed 106 ± 17 PgC by 1995, while Khatiwala et al. (2009), hereafter referred to as Khatiwala et al. (2009), estimate 114 ± 22 PgC. The CMIP5 mean for 1995 C_{ant} is 90 ± 7 PgC, underestimating oceanic C_{ant} storage compared to both the Sabine et al. (2004) and Khatiwala et al. (2009) observations. Observational and model mean estimates therefore disagree by as much as 25%. Since carbon is the primary agent driving global change, so it is important to constrain both observational and models estimates, as well as model spread.

In this paper, we argue that the largest biases are due to the difference in time period covered by each estimate of C_{ant}: CMIP5 model historical simulations collectively begin in the year 1850, while Sabine et al. (2004) measure C_{ant} from 1791 and Khatiwala et al. (2009) from 1765. While fossil fuel emissions prior to 1850 are negligible, changes in land use resulted in increased atmospheric CO_2 levels since 1765 (Khatiwala et al., 2009). These pre-1850 rises in atmospheric pCO_2 will increase ocean carbon long after that change in atmospheric pCO_2 has ended, since ocean uptake of atmospheric carbon occurs over multiple time scales (Joos et al., 2013; Revelle & Suess, 1957). We present two methods for adjusting ocean C_{ant} uptake due to prior rises in

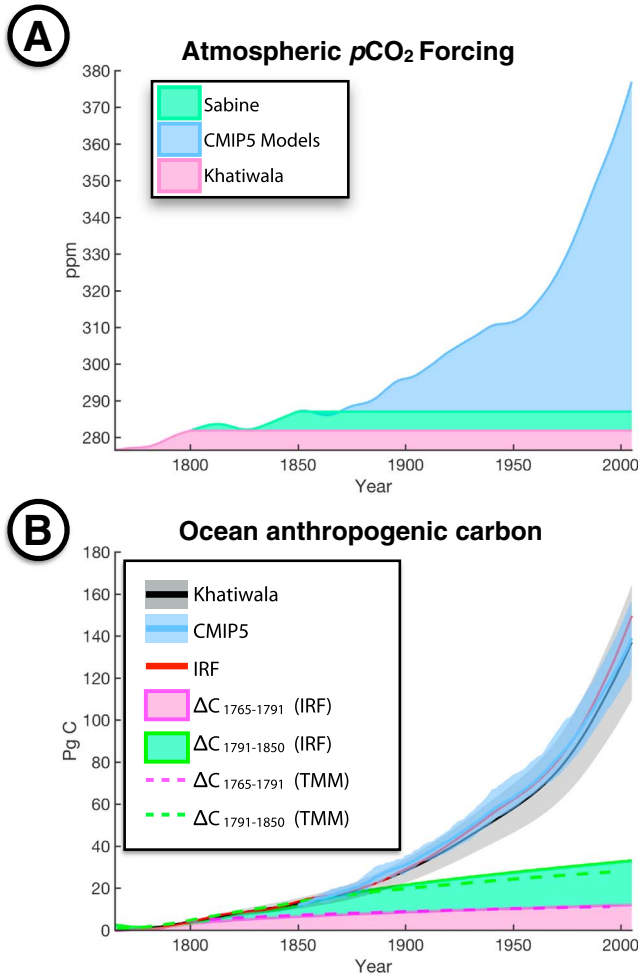

Figure 1. (a) Atmospheric pCO_2 as a function of time. The blue-shaded region is the forcing applied to CMIP5 historical simulations, while the green and magenta regions are the additional forcings of the Sabine et al. (2004) and Khatiwala et al. (2009) estimates. (B) Historical adjusted ocean anthropogenic carbon from Khatiwala et al. (2009) (black), using impulse response functions (red), adjusted CMIP5 models (blue), $\Delta C_{1765-1791}$ and $\Delta C_{1791-1850}$ (magenta and green, respectively). For the adjustments, solid lines show IRF results and dashed lines show the results using the TMM. Note the different vertical scales on all the panels.

atmospheric pCO_2. In section 2 we discuss the observational products we will use. Section 3 presents the methods used to adjust C_{ant} estimates to the same start date, and section 4 discusses the results of the adjustment. Section 5 summarizes the results.

2. Estimates of Anthropogenic Carbon

In CMIP5 historical simulations, generally from 1850 to 2005, C_{ant} is taken as the difference in dissolved inorganic carbon (DIC) over the historical period with the preindustrial control simulation (atmospheric pCO_2 levels are prescribed). In the real ocean, anthropogenic carbon is difficult to estimate directly because we lack sufficiently accurate measurements over the historical period. Consequently, alternative methods have been developed to estimate anthropogenic carbon. We discuss two methods which make estimates of C_{ant} based on observational constraints: the ΔC^* method (Gruber et al., 1996), which was applied by Sabine et al. (2004), and the Green's function method by Khatiwala et al. (2009). The ΔC^* method uses contemporary measurements of DIC in the ocean and estimates the preindustrial and biological carbon to calculate the remainder or the anthropogenic component. The Green's function method is based on the transit tracer method as employed by Waugh et al. (2006). It relies on constructing an estimate of ocean transport based on direct observations of ocean tracers and using the transport matrix to propagate surface carbon anomalies into the ocean interior. There are some well-understood systematic differences, in particular, the ΔC^* method overestimates C_{ant} in relatively young water and underestimates it in older water. For both observational products, the largest common source of random error comes from sparse spatial and temporal sampling. While sampling error does provide the largest source of error, it does not explain a systematic global bias between both observational estimates and the CMIP5 models.

Previous studies have discussed and tested various assumptions of each method, such as the effects of changes in circulation, ocean temperature, and ocean chemistry (see supporting information) (e.g., Haine & Hall, 2002; Keeling, 2005; Matsumoto & Gruber, 2005; McNeil & Matear, 2013; Plattner et al., 2001; Wang et al., 2012; Weber & Deutsch, 2010). However, none of the assumptions tested have been able to explain the global difference between observational products and between the observations and model estimates. We show below that the largest difference between the three estimates is due to differences in the assumed start date of the measurement: method 1 measures ocean C_{ant} from 1791, method 2 from 1765, and CMIP5 simulates from 1850. We note that while the Sabine et al. (2004) estimate is nominally from 1800, the preindustrial ocean carbon is calculated with respect to an atmospheric pCO_2 of 280 ppm, which was reached in 1791 (MacFarling Meure et al., 2006). Therefore, we reference to the year 1791.

3. Adjusting Uptake to Start Date: Methods

Between 1765 and 1850, atmospheric pCO_2 rose roughly 10 ppm. We reference all three methods to the same time period—the period of the Sabine et al. (2004) estimate—by accounting for the ocean carbon that is due to rises in atmospheric CO_2 between 1765 and 1791, a negative adjustment to the Khatiwala et al. (2009) estimate, and between 1791 and 1850, a positive adjustment to the CMIP5 estimate. This concept is illustrated in Figure 1a, where the blue-shaded region shows the atmospheric changes modeled by the CMIP5 simulations, the green region the additional atmospheric CO_2 perturbation forcing for the Sabine et al. (2004) estimate, and the magenta region the additional atmospheric pCO_2 history of the Khatiwala et al. (2009) estimate. The Khatiwala et al. (2009) estimate is from 1765 to 1995 and Khatiwala et al. (2013) extended this to 2011.

Table 1
Ocean Anthropogenic Carbon

Number	Model	Carbon storage 1791–1995 (PgC)	Carbon storage 1791–2011 (PgC)
1	CMCC-CESM	100.5	137.4
2	CNRM	87.5	120.9
3	IPSL-CM5A LR	96.2	135.1
4	IPSL-CM5A MR	101.7	145.1
5	IPSL-CM5B LR	95.2	125.6
6	MIROC ESM	93.9	128.4
7	MIROC ESM CHEM	93.2	127.3
8	HadGEM2-ES*	102.4	129.3
9	HadGEM2-CC*	103.9	130.5
10	MPI ESM MR	96.0	142.1
11	GISS E2 R CC	107.4	148.6
12	NorESM	105.9	146.1
13	GFDL ESM2M*	102.0	142.3
14	GFDL ESM2G*	93.9	135.5
-	Model Mean	98.6 ± 10.0	136.6 ± 14.0
-	Khatiwala	100.8 ± 20.2	134.9 ± 24
-	Sabine	97.5 ± 14.4	-

Note. Models denoted by an asterisk begin their historical simulations in 1860.

Between 1765 and 1850, the Khatiwala et al. (2009) data set estimates that the ocean had stored 14.3 PgC. However, ocean uptake of atmospheric carbon occurs over multiple time scales, such that the ocean will continue to absorb carbon due to a rise in atmospheric pCO_2 long after that rise has ended. In order to make the smallest adjustment possible, we reference the CMIP5 models and the Khatiwala et al. (2009) data to the Sabine et al. (2004) period: 1791–1995. We also note that some CMIP5 models begin their historical simulations in 1860, denoted by an asterisk in Table 1, so a different adjustment is made to these models to reference them to a 1791 start date.

We define the ocean anthropogenic carbon as function of time t due to the 1765–1791 rise in atmospheric pCO_2 as $\Delta C_{1765-1791}(t)$, and the ocean carbon due to the 1791–1850 rise in atmospheric pCO_2 as $\Delta C_{1791-1850}(t)$. $\Delta C_{1765-1791}(t)$ and $\Delta C_{1791-1850}(t)$ are calculated using two separate methods: a transport matrix and impulse response functions.

3.1. Transport Matrix Method

The transport matrix method (TMM) simulates ocean biogeochemical tracers in an "off-line" manner that is designed to be more efficient than a full general circulation model (GCM) (Khatiwala et al., 2005). The transport matrix for tracer propagation itself can be extracted from any GCM. To simulate ocean anthropogenic carbon, we prescribe atmospheric pCO_2. The surface ocean carbon anomaly is then transported into the ocean interior as an abiotic tracer, dependent on local chemistry, according to the Ocean Carbon-Cycle Model Intercomparison Project 2 (OCMIP-2) formulation (Orr, 1999). In this study, we use the matrix extracted from the 1° physical ocean state estimate from the Estimating the Circulation and Climate of the Ocean (ECCO) project v1 state estimate (Stammer et al., 2004). The ECCO state estimate was constructed using the MITgcm (Marshall et al., 1997) and data assimilation. The ECCO matrix has been applied to simulate anthropogenic carbon uptake (Graven et al., 2012) and was found to match the Khatiwala et al. (2013) estimate well, within less than 2% error over the 1765–1995 period. While the ECCO configuration is a GCM, it has been optimized with observations and produces carbon uptake that is very close to the observational product of Khatiwala et al. (2013). Similar to the Khatiwala et al. (2009) method, ocean dynamics is assumed to be constant in the TMM.

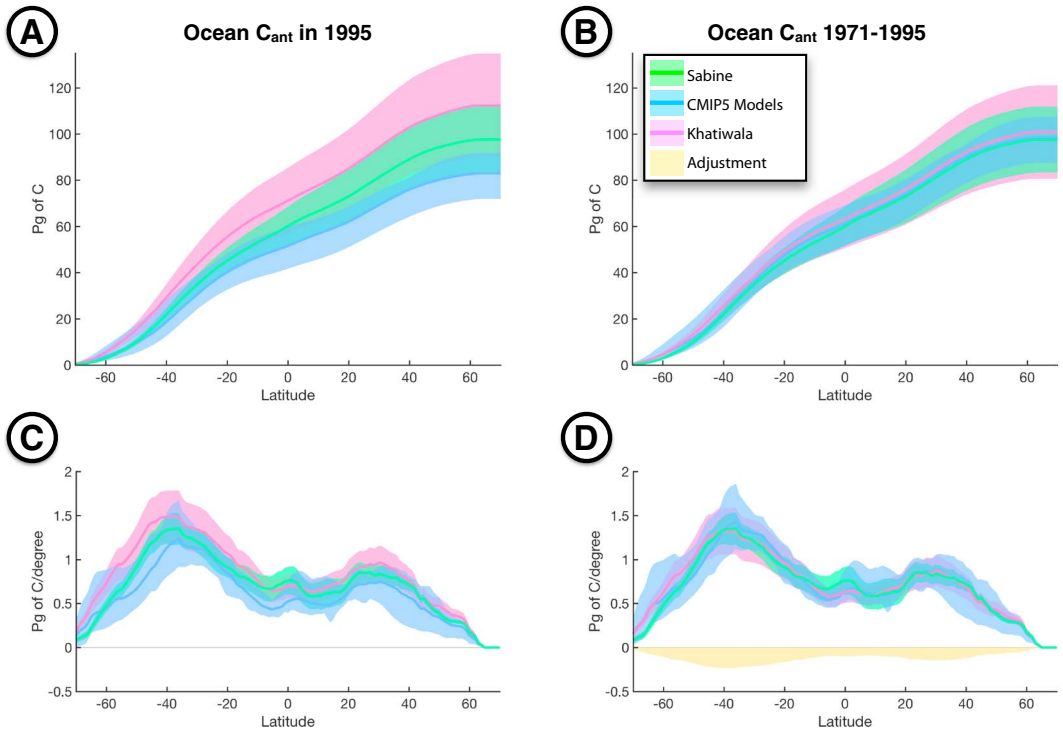

Figure 2. (a, b) The 1995 oceanic anthropogenic carbon content integrated with latitude. The carbon content is integrated zonally and then integrated cumulatively from 70°S to 70°N (which covers all the data from the three estimates), so that the value at 70°N is the total carbon uptake in the year 1995. Figure 2a is unadjusted and Figure 2b is adjusted relative to 1791 start date. (c, d) The 1995 oceanic anthropogenic carbon content with latitude. Figure 2c is unadjusted and Figure 2d is adjusted. Magenta and green show the Khatiwala et al. (2009) and Sabine et al. (2004) estimates, respectively. The blue line shows the CMIP5 mean for the models listed in Table 1, with the blue-shaded region showing the model spread. The yellow-shaded region in Figure 2d shows $\Delta C_{1765-1791}$, which is removed from the Khatiwala et al. (2009) observations.

We estimate $\Delta C_{1791-1850}$ from 1791 to 1995 by prescribing measured atmospheric pCO_2 levels from 1791 to 1850 and then fixing them at the 1850 value from 1850 to 1995 to adjust the CMIP5 models. We also estimate $\Delta C_{1765-1791}$ by the same method to reference the Khatiwala et al. (2009) data to the Sabine et al. (2004) period.

3.2. Impulse Response Functions

To provide a second estimate of $\Delta C_{1791-1850}$ and $\Delta C_{1765-1791}$ for verification, we use the impulse response functions (IRFs) from Joos et al. (2013). The IRF as function of time t, $I(t)$, is the response of a system due to a pulse emission. Joos et al. (2013) presents the result of multiple model experiments whereby a pulse of 100 Gt C is emitted in both preindustrial as well as 2010 conditions. The model mean IRFs for both atmospheric, oceanic, and land carbon are given and they capture the responses of a set of fully coupled climate models and models of intermediate complexity. The IRFs can then be used to find the response of the system given any emission history $E(t)$. Assuming that every year's carbon emission is an impulse, we can convolve the emission history that corresponds with atmospheric pCO_2 adjustments from Figure 1a, with the ocean IRFs from Joos et al. (2013) to calculate $\Delta C_{1765-1791}$ and $\Delta C_{1791-1850}$. The full details of this procedure is outlined in the supporting information (Boden et al., 2009; Clark, 1982; Houghton, 2003).

4. Adjusting Uptake to Start Date: Results

To compare the results of both the TMM and the IRFs, the time evolution of both $\Delta C_{1791-1850}$ and $\Delta C_{1765-1791}$ evaluated using each method is shown in Figure 1b. Despite different approaches, both the TMM and IRFs show a similar time evolution and the magnitude of $\Delta C_{1791-1850}$ and $\Delta C_{1765-1791}$ in 1995 agree to within 14% (4 PgC). While the magnitudes agree, only the TMM gives a spatial estimate. If we add the mean TMM and IRF estimates of $\Delta C_{1791-1850}$ and $\Delta C_{1765-1791}$ to the time evolution of total CMIP5 C_{ant}, we see that this gives a remarkable agreement with the Khatiwala et al. (2013) data in 1995 (Figure 1b). In the midtwentieth century,

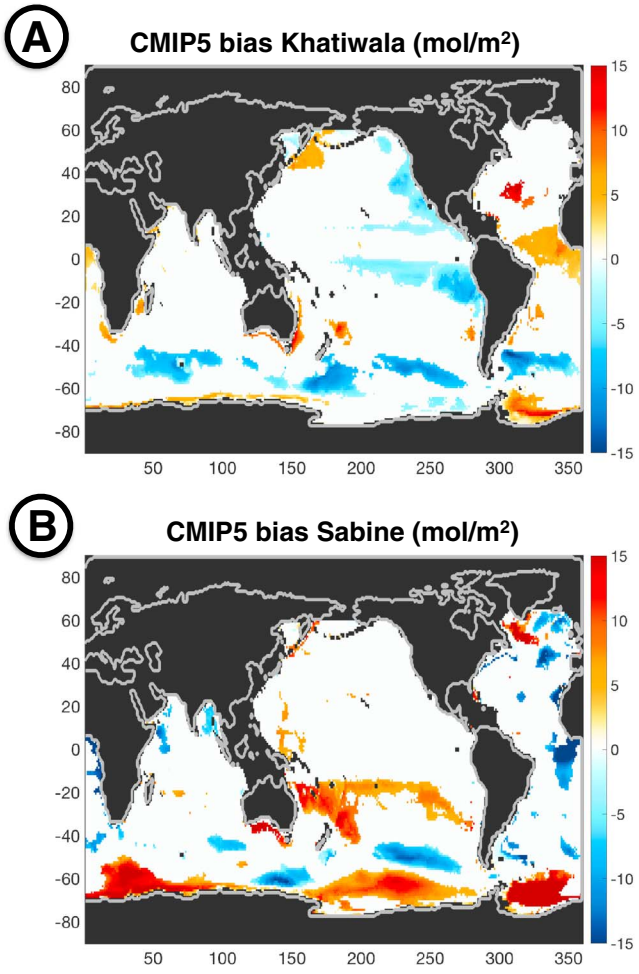

Figure 3. CMIP5 model bias of column-integrated 1995 anthropogenic carbon relative to the adjusted (a) Khatiwala et al. (2009) and (b) Sabine et al. (2004) estimates. The white areas are where the bias is less than the error in the observations, and the dark grey areas are those not covered by the observations.

however, we note that CMIP5 models consistently overestimate C_{ant} compared to Khatiwala et al. (2013) before coming together at the end of the twentieth century. This shift in CMIP5 carbon uptake may be due to changes in circulation or uncertainties in atmospheric pCO_2 histories used.

In 1791, the total of $\Delta C_{1765-1791}$ in the entire ocean is estimated to have been 2.1 ± 1.2 PgC. However, in 1995 the total had grown to 12.1 ± 3.8 PgC, an increase of over 400%. Similarly, we estimate $\Delta C_{1791-1850}$ to be 5.2 ± 1.0 PgC in 1850 and 16.6 ± 2.6 PgC in 1995. These results show that the ocean carbon due to a relatively small rise in atmospheric pCO_2 can increase drastically over time. Ice core atmospheric pCO_2 values prior to 1850 are quoted with a ±3 ppm error. While we use the same atmospheric pCO_2 time series as Khatiwala et al. (2009) and our adjustments are therefore independent of these errors, the error might affect the date at which an atmospheric pCO_2 of 280 ppm was reached. Adjusting for a ±3 ppm error in atmospheric pCO_2 gives an error of less than ±1 PgC in both $\Delta C_{1765-1791}$ and $\Delta C_{1791-1850}$ in 1995.

The majority of $\Delta C_{1765-1791}$ and $\Delta C_{1791-1850}$ is absorbed after 1791 and 1850, respectively, after atmospheric pCO_2 levels have stabilized. CMIP5 simulations are spun up to different equilibration degrees, but they aim to start the historical simulations from a state equilibrated to 1850 atmospheric pCO_2 levels. The surface carbon concentrations in this equilibrated state will not be the same as a transient system with the same atmospheric pCO_2 which has experienced a complete pCO_2 history. This can be illustrated by model simulations whereby atmospheric pCO_2 is increased steadily with time and then held constant, such as the TMM simulations (see supporting information) or RCP scenarios followed by stabilization experiments performed by Zickfeld et al. (2013).

In these experiments, the surface DIC at the time of atmospheric pCO_2 stabilization is less than it would be at equilibrium for the same atmospheric pCO_2 value. During the stabilization following an atmospheric pCO_2 ramp, the surface carbon concentration increases over time. The surface DIC in the CMIP5 experiments at 1850 will therefore be higher than it was in the real ocean, resulting in a larger air-sea pCO_2 disequilibrium. This difference in air-sea pCO_2 disequilibrium drives an additional flux of carbon into the real ocean that is lacking in the models, causing a bias in the ocean carbon uptake rate between 1850 and 1995. The carbon content of the models and the observations therefore evolve at different rates over the historical period. As a result, the difference in carbon content of the ocean in 1995 between the CMIP5 models and the observations is not the same as the difference in carbon in 1850 and 1765 in the CMIP5 models and the observations, respectively.

While the long time scale of ocean carbon uptake has been studied before (Eby et al., 2013; Siegenthaler & Sarmiento, 1993), our results show how a modest difference in atmospheric pCO_2 at the beginning of the historical period can result in a significant fraction of ocean uptake in 1995. In the case of the Khatiwala et al. (2009) estimate, a difference of 3.4 ppm between 1765 and 1791, 3% of the change in atmospheric pCO_2 between 1765 and 1995, results in 11% of ocean C_{ant} in 1995.

Using our results to adjust the start date of the CMIP5 and Khatiwala et al. (2009) estimates of C_{ant} to 1791, we find good agreement between the methods. Figure 2a shows the unadjusted 1791–1995 C_{ant} integrated with latitude. Figure 2b shows the same but adjusted to a start time of 1791. To ensure that all data sets cover the same area, we apply the most limiting mask, that of Khatiwala et al. (2009). We see that without the start time adjustment, both the Khatiwala et al. (2009) estimate and the CMIP5 lie outside the quoted error of the Sabine et al. (2004) estimate. However, when adjusted, all three methods agree very well, to within 4% (see Table 1). The CMIP5 models, as a set, are therefore in agreement with the observations, and both observational products agree with each other within the observational error.

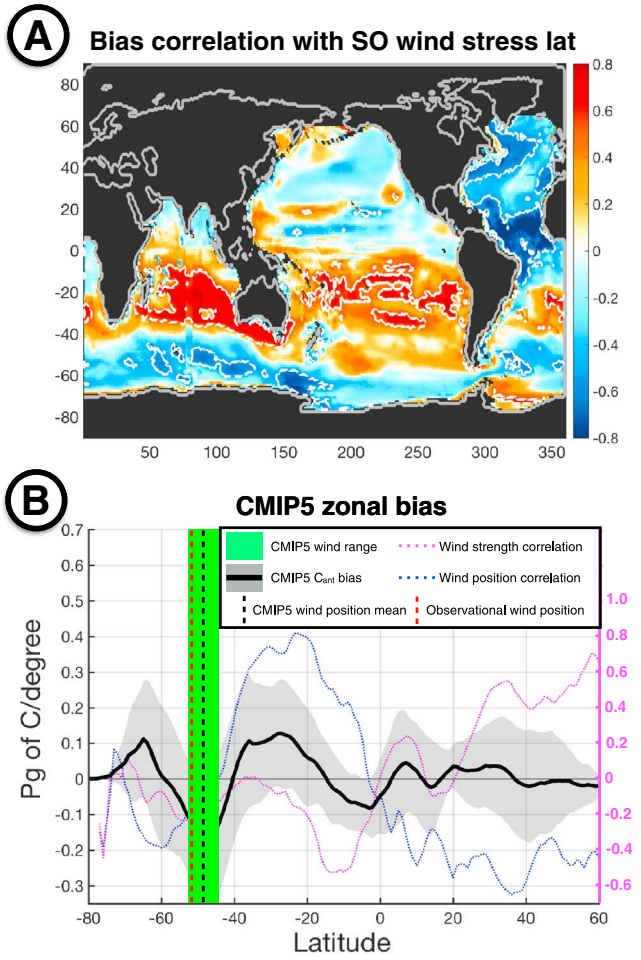

Figure 4. (a) Correlation of the latitude of the maximum in Southern Hemisphere zonal wind stress with the CMIP5 C_{ant} bias relative to the mean of Khatiwala et al. (2009) and Sabine et al. (2004). The areas inside the white dashed contours have a significance level of $p < 0.05$. (b) Zonally and depth-integrated bias of CMIP5 with the mean of Khatiwala et al. (2009) and Sabine et al. (2004). The black solid line shows the mean CMIP5 bias, and the grey-shaded area shows the range of individual models. The green bar shows the range of the maximum in CMIP5 preindustrial Southern Hemisphere zonal wind stress, with the black dashed line showing the CMIP5 mean and the red dashed line the observational estimate for the period 1979–2010 (Swart & Fyfe, 2012). The blue dashed line is the correlation of the zonally integrated CMIP C_{ant} bias with the latitude of the maximum in Southern Hemisphere zonal wind stress, and the magenta dashed line is the correlation with the maximum strength of the wind stress, with corresponding *y* axis shown on the right.

When comparing 1791–1995 carbon uptake as a function of latitude, Figure 2c shows the unadjusted values while Figure 2d shows the adjusted uptake. In Figure 2d, the only significant (outside of the observational error) CMIP5 model bias lies in the Southern Hemisphere. There, the models underestimate carbon uptake in the Southern Ocean between 60° and 40°S and overestimate it in the Indian and Pacific Oceans 40°–10°S. A more comprehensive picture of spatial model biases is shown by the column-integrated C_{ant} CMIP5 mean bias relative to Khatiwala et al. (2009) and Sabine et al. (2004) in Figures 3a and 3b, respectively (for individual model biases, see supporting information). Relative to both observational products, the models store too much carbon in the southern subtropical Indian and Pacific Oceans and not enough in the open Southern Ocean, by as much as 10 mol m^{-2}. There are significant differences between the observational products around Antarctica and in the Atlantic Ocean.

We propose that the model bias in the Southern Ocean and Subtropical Indian/Pacific Oceans is due to a bias in the position of the preindustrial Southern Ocean westerlies. The majority of CMIP5 models have equatorward biased winds (Swart & Fyfe, 2012) leading to increased carbon uptake in the subtropics. Figure 4a shows the spatial correlation of column-integrated CMIP5 C_{ant} bias with the position of the maximum in Southern Hemisphere westerlies. Figure 4b shows the zonally integrated CMIP5 C_{ant} bias, the position of the winds in CMIP5 and observations as well as the correlation of the CMIP5 C_{ant} bias with the latitude of the westerlies. We note a strong positive correlation in the Indian and Pacific Oceans, and a negative correlation in the Southern Ocean. We find that equatorward winds result in stronger subtropical gyres which are more latitudinally confined compared to models with poleward winds. As a result, the smaller and stronger gyres accumulate more carbon. Meanwhile, equatorward winds also result in a more stratified Southern Ocean and less vertical mixing of carbon throughout the water column (e.g., Lauderdale et al., 2013).

The wind bias produces a dipole in the carbon uptake but does not affect the overall magnitude since we do not find correlation between Southern Hemisphere wind position and total C_{ant}. Studies such as Russell et al. (2006) propose that equatorward biased winds lead to decreased global carbon uptake. However, Russell et al. (2006) compared models with differences in both westerly position and strength. Comparing the correlation of CMIP5 C_{ant} bias with westerly position and strength in Figure 4b, we note that equator-biased winds suggest decreased carbon uptake in the Northern Hemisphere, particularly in the Northern Atlantic. However, North Atlantic C_{ant} bias also shows a significant correlation with Southern Hemisphere westerly wind strength, suggesting that poleward intensified winds can lead to higher Atlantic C_{ant} uptake. In CMIP5, however, we find no significant correlation between preindustrial wind position and strength such that there is no significant bias in the Atlantic. If Southern Ocean winds intensify poleward into the future, however, it is possible that wind position and strength together will have an impact on Atlantic Ocean and global carbon uptake.

5. Conclusions

A comparison between observational and model estimates of global historical ocean anthropogenic carbon uptake has revealed a large bias in the models. In this paper, we bring the estimates in closer agreement by adjusting for the effects of different start dates between the CMIP5, Khatiwala et al. (2009), and Sabine et al. (2004) estimates of ocean anthropogenic carbon.

CMIP5 historical simulations start in 1850, while the Khatiwala et al. (2009) estimate begins in 1765 and Sabine et al. (2004) uses 1791 as the start date. We adjust the CMIP5 and Khatiwala et al. (2009) to a 1791 start date by accounting for the present-day ocean carbon that is due to perturbations in atmospheric pCO_2 between the periods 1765–1791 and 1791–1850. To do so, we use the ocean impulse response functions published by Joos et al. (2013) and a transport matrix method derived from ECCOv1. We find that there is considerable uptake of ocean carbon after 1791 due to the pre-1791 rise in atmospheric pCO_2. After referencing all estimates to the period 1791–1995, we show that the Khatiwala et al. (2009), Sabine et al. (2004), and CMIP5 ocean carbon uptake agree to within 3 Pg of C. However, individual model bias compared to the observations of up to 12% remain. We do note that both the IRFs used to calculate the adjustments assume that the adjustment in ocean uptake occurs relative to the 1850 ocean. However, ocean chemistry will have changed with time due to actual changes in ocean carbon. Due to this effect, the uptake measured by the IRF adjustment is an underestimate prior to 1850 and an overestimate after 1850. However, we expect the error induced by this effect to be no larger than 4.5% in 1995 (see supporting information) (Goodwin et al., 2007).

The CMIP5 model bias compared to the observations in most regions of the ocean is smaller than the observational error. In some areas the observations disagree in the magnitude of the carbon uptake. In the Southern Indian and Pacific Oceans 40°–10°S, the models store too much carbon and not enough in the Southern Ocean 60°–40°S. We believe this dipole bias is likely due to the bias in the position of Southern Hemisphere westerlies. However, we do not think that this dipole bias has affected the global carbon uptake to date, though the wind position may affect total carbon uptake in the future.

Uncertainties in the observational products and discrepancies between the Khatiwala et al. (2009) and Sabine et al. (2004) estimates prevent us from identifying additional model biases outside the Southern Ocean and the Southern Indian and Pacific Oceans. Figure 2b shows that the CMIP5 model spread lies well within the quoted observational errors. This feature also pertains to 2011. The Sabine estimate does not extend to 2011, but the adjusted Khatiwala et al. (2013) data set gives a value of 135.1 ± 24 PgC. The CMIP5 models estimate the 1791–2011 carbon uptake as 135.3 ± 14.0 PgC. We also find that the spread in CMIP5 model C_{ant} scales linearly with cumulative ocean carbon uptake. By using the impulse response functions and assuming a best case scenario where the error in the observations remains constant at ± 24 PgC (i.e., assuming "perfect" observations in the future with no error), it is not until the year 2040 that a single model would emerge outside the observational error, for an RCP8.5 scenario. For more moderate emission scenarios such as RCP4.5, this year would be 2045. Given perfect observations, it would still take decades until we could begin to constrain the global spread in CMIP5 ocean anthropogenic carbon using observations. This result emphasizes the importance of improving global estimates of cumulative ocean carbon uptake to date as well as into the future.

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References

Boden, T. A., Marland, G., & Andreas, R. J. (2009). *Global, regional, and national fossil-fuel CO_2 emissions*. Oak Ridge, TN: Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory.

Clark, W. C. E. (1982). *Carbon dioxide review*. New York: Oxford University Press.

Eby, M., Weaver, A. J., Alexander, K., Zickfeld, K., Abe-Ouchi, A., Cimatoribus, A. A., … Zhao, F. (2013). Historical and idealized climate model experiments: An intercomparison of Earth system models of intermediate complexity. *Climate of the Past*, *9*(3), 1111–1140. https://doi.org/10.5194/cp-9-1111-2013

Frölicher, T. L., Sarmiento, J. L., Paynter, D. J., Dunne, J. P., Krasting, J. P., & Winton, M. (2015). Dominance of the southern ocean in anthropogenic carbon and heat uptake in CMIP5 models. *Journal of Climate*, *28*(2), 862–886. https://doi.org/10.1175/JCLI-D-14-00117.1

Goodwin, P., Williams, R. G., Follows, M., & Dutkiewicz, S. (2007). Ocean-atmosphere partitioning of anthropogenic carbon dioxide on centennial timescales. *Global Biogeochemical Cycles*, *21*, GB1014. https://doi.org/10.1029/2006GB002810

Graven, H. D., 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. *Journal of Geophysical*, *117*, C10005. https://doi.org/10.1029/2012JC008074

Gruber, N., Sarmiento, J., & Stocker, T. (1996). An improved method for detecting anthropogenic CO_2 in the oceans. *Global Biogeochemical Cycles*, *10*(4), 809–837. https://doi.org/10.1029/96GB01608

Haine, T., & Hall, T. (2002). A generalized transport theory: Water-mass composition and age. *Journal of Physical Oceanography*, *32*(6), 1932–1946. https://doi.org/10.1175/1520-0485(2002)032<1932:AGTTWM>2.0.CO;2

Houghton, R. (2003). Revised estimates of the annual net flux of carbon to the atmosphere from changes in land use and land management 1850–2000. *Tellus Series B-Chemical and Physical Meteorology*, *55*(2), 378–390. https://doi.org/10.1034/j.1600-0889.2003.01450.x

Joos, F., Roth, R., Fuglestvedt, J. S., Peters, G. P., Enting, I. G., von Bloh, W., … Weaver, A. J. (2013). Carbon dioxide and climate impulse response functions for the computation of greenhouse gas metrics: A multi-model analysis. *Atmospheric Chemistry and Physics*, *13*(5), 2793–2825. https://doi.org/10.5194/acp-13-2793-2013

Keeling, R. (2005). Comment on "The ocean sink for anthropogenic CO_2". *Science*, *308*(5729), 1743. https://doi.org/10.1126/science.1109620

Khatiwala, S., Primeau, F., & Hall, T. (2009). Reconstruction of the history of anthropogenic co2 concentrations in the ocean. *Nature*, *462*(7271), 346–349.

Khatiwala, S., Visbeck, M., & Cane, M. (2005). Accelerated simulation of passive tracers in ocean circulation models. *Ocean Modelling*, *9*(1), 51–69. https://doi.org/10.1016/j.ocemod.2004.04.002

Khatiwala, S., Tanhua, T., Fletcher, S. M., Gerber, M., Doney, S. C., Graven, H. D., … Sabine, C. L. (2013). Global ocean storage of anthropogenic carbon. *Biogeosciences*, *10*(4), 2169–2191. https://doi.org/10.5194/bg-10-2169-2013

Lauderdale, J. M., Garabato, A. C. N., Oliver, K. I. C., Follows, M. J., & Williams, R. G. (2013). Wind-driven changes in Southern Ocean residual circulation, ocean carbon reservoirs and atmospheric CO_2. *Climate Dynamics*, *41*(7–8), 2145–2164. https://doi.org/10.1007/s00382-012-1650-3

Le Quere, C., Andrew, R. M., Canadell, J. G., Sitch, S., Korsbakken, J. I., Peters, G. P., … Zaehle, S. (2016). Global carbon budget 2016. *Earth System Science Data*, *8*(2), 605–649. https://doi.org/10.5194/essd-8-605-2016

MacFarling Meure, C., Etheridge, D., Trudinger, C., Steele, P., Langenfelds, R., van Ommen, T., … Elkins, J. (2006). Law dome CO_2, CH_4 and N_2O ice core records extended to 2000 years BP. *Geophysical Research Letters*, *33*, L14810. https://doi.org/10.1029/2006GL026152

Marshall, J., Adcroft, A., Hill, C., Perelman, L., & Heisey, C. (1997). A finite-volume, incompressible Navier-Stokes model for studies of the ocean on parallel computers. *Journal of Geophysical Research*, *102*, 5753–5766.

Matsumoto, K., & Gruber, N. (2005). How accurate is the estimation of anthropogenic carbon in the ocean? An evaluation of the Delta C* method. *Global Biogeochemical Cycles*, *19*, GB3014. https://doi.org/10.1029/2004GB002397

McNeil, B. I., & Matear, R. J. (2013). The non-steady state oceanic CO_2 signal: Its importance, magnitude and a novel way to detect it. *Biogeosciences*, *10*(4), 2219–2228. https://doi.org/10.5194/bg-10-2219-2013

Orr, J. (1999). On ocean carbon-cycle model comparison. *Tellus Series B-Chemical and Physical Meteorology*, *51*(2), 509–510.

Plattner, G., Joos, F., Stocker, T., & Marchal, O. (2001). Feedback mechanisms and sensitivities of ocean carbon uptake under global warming. *Tellus Series B-Chemical and Physical Meteorology*, *53*(5), 564–592. https://doi.org/10.1034/j.1600-0889.2001.530504.x

Revelle, R., & Suess, H. S. (1957). Carbon dioxide exchange between atmosphere and ocean and the questions of an increase of atmospheric CO_2 during the past decades. *Tellus*, *9*, 18–27. https://doi.org/10.1111/j.2153-3490. 1957.tb01849.x

Russell, J. L., Dixon, K. W., Gnanadesikan, A., Stouffer, R. J., & Toggweiler, J. R. (2006). The Southern Hemisphere westerlies in a warming world: Propping open the door to the deep ocean. *Journal of Climate*, *19*(24), 6382–6390. https://doi.org/10.1175/JCLI3984.1

Sabine, C., Feely, R., Gruber, N., Key, R., Lee, K., Bullister, J., … Rios, A. (2004). The oceanic sink for anthropogenic CO_2. *Science*, *305*(5682), 367–371. https://doi.org/10.1126/science.1097403

Siegenthaler, U., & Sarmiento, J. (1993). Atmospheric carbon-dioxide and the ocean. *Nature*, *365*(6442), 119–125. https://doi.org/10.1038/365119a0

Stammer, D., Ueyoshi, K., Kohl, A., Large, W., Josey, S., & Wunsch, C. (2004). Estimating air-sea fluxes of heat, freshwater, and momentum through global ocean data assimilation. *Journal of Geophysical Research*, *109*, C05023. https://doi.org/10.1029/2003JC002082

Swart, N. C., & Fyfe, J. C. (2012). Observed and simulated changes in the Southern Hemisphere surface westerly wind-stress. *Geophysical Research Letters*, *39*, L16711. https://doi.org/10.1029/2012GL052810

Wang, S., Moore, J. K., Primeau, F. W., & Khatiwala, S. (2012). Simulation of anthropogenic CO_2 uptake in the CCSM3.1 ocean circulation-biogeochemical model: Comparison with data-based estimates. *Biogeosciences*, *9*(4), 1321–1336. https://doi.org/10.5194/bg-9-1321-2012

Waugh, D. W., Hall, T. M., McNeil, B. I., Key, R., & Matear, R. J. (2006). Anthropogenic CO_2 in the oceans estimated using transit time distributions. *Tellus Series B-Chemical and Physical Meteorology*, *58*(5), 376–389.

Weber, T., & Deutsch, C. (2010). Ocean nutrient ratios governed by plankton biogeography. *Nature*, *467*, 550–554.

Zickfeld, K., Eby, M., Weaver, A. J., Alexander, K., Crespin, E., Edwards, N. R., … Zhao, F. (2013). Long-term climate change commitment and reversibility: An EMIC intercomparison. *Journal of Climate*, *26*(16), 5782–5809. https://doi.org/10.1175/JCLI-D-12-00584.1