The North Atlantic is one of the major sinks for anthropogenic carbon in the global ocean. Improved understanding of the underlying mechanisms is vital for constraining future projections, which presently have high uncertainties. To identify some of the causes behind this uncertainty, this study investigates the North Atlantic's anthropogenically altered carbon uptake and inventory, that is, changes in carbon uptake and inventory due to rising atmospheric CO2 and climate change (abbreviated as $C_{\text{ant}}$-uptake and $C_{\text{ant}}$-inventory). Focus is set on an ensemble of 11 Earth system models and their simulations of a future with high atmospheric CO2. Results show that the model spread in the $C_{\text{ant}}$-uptake originates in middle and high latitudes. Here, the annual cycle of oceanic $pCO_2$ reveals inherent model mechanisms that are responsible for different model behavior: while it is SST-dominated for models with a low future $C_{\text{ant}}$-uptake, it is dominated by deep winter mixing and biological production for models with a high future $C_{\text{ant}}$-uptake. Models with a high future $C_{\text{ant}}$-uptake show an efficient carbon sequestration and hence store a large fraction of their contemporary North Atlantic $C_{\text{ant}}$-inventory below 1000-m depth, while the opposite is true for models with a low future $C_{\text{ant}}$-uptake. Constraining the model ensemble with observation-based estimates of carbon sequestration and summer oceanic $pCO_2$ anomalies yields later flattening of the $C_{\text{ant}}$-uptake than previously estimated. This result highlights the need to depart from the concept of unconstrained model ensembles in order to reduce uncertainties associated with future projections.

1. Introduction

The ocean plays an important role in the mediation of climate change as it takes up CO2 and heat from the atmosphere (Karl and Trenberth 2003; Le Quéré et al. 2016). With ongoing global warming, the ocean’s physical and biogeochemical properties are very likely to undergo fundamental changes (Heinze et al. 2015; Bopp et al. 2013), which are assumed to reduce the ocean’s ability to take up CO2 and would hence result in a larger airborne fraction (Ciais et al. 2013).

To date, the state-of-the-art tools to project future climate change and its consequences are Earth system models (ESMs). ESMs were featured in the latest report on climate change of the Intergovernmental Panel on Climate Change (IPCC; IPCC 2013), where their evaluation found them to be suitable for quantitative future projections (Flato et al. 2013). Yet, ESMs are simplified descriptions of complicated systems, and their development is constantly ongoing: new processes are regularly added, and the representation of already-included processes, as well as the grid resolution, is constantly refined (Flato 2011; Heavens et al. 2013). The evaluation of ESMs

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with all included processes is increasingly complex and, at the same time, limited by the availability of observational data (Flato 2011). Sparsity of data is especially a challenge when trying to evaluate the ocean component and its biogeochemical aspects.

To represent the uncertainties associated with ESMs, the IPCC (2013) considers results from a multi-ESM ensemble and assigns a confidence level to future estimates according to consistency of model results, determined as the standard deviation of the multimodel mean without any consideration of the performance of each individual ensemble member. Hence, a model that does not perform well has the same importance for the end result as a well-performing model, allowing the multimodel mean to degrade more than necessary. Yet, selection or weighting strategies are difficult to find, as a model’s present-day performance is not necessarily related to its ability to project future change (Knutti et al. 2010).

The current generation of ESMs agrees well on global estimates of future carbon uptake by the oceans (Jones et al. 2013), but at the regional level, the spread can be substantial. The North Atlantic has been identified as the region with the largest climate-induced reduction of CO₂ uptake in terms of changes per unit area (Plattner et al. 2001; Roy et al. 2011). This is cause for concern, as the North Atlantic accounts for about one-third of the present oceanic CO₂ uptake (Takahashi et al. 2009) and stores more than 23% of the total oceanic anthropogenic CO₂ content (Sabine et al. 2004). However, current ESM projections for the twenty-first-century North Atlantic CO₂ uptake are highly divergent (Wang et al. 2016), and a more constrained future estimate is urgently needed.

As a first step toward more accurate North Atlantic carbon uptake projections, we analyze the modeled future carbon uptake for this region for an ensemble of ESMs. To be able to quantify the mechanisms behind any projected change, we introduce and analyze the anthropogenically altered carbon budget. A description of this quantity, as well as of the considered ESMs, is given in section 2, while the anthropogenically altered carbon budget and its associated mechanisms are analyzed and discussed in sections 3–6. A summary of our results and our recommendations for future model assessments are found in section 7.

2. Study design

a. Study area, model ensemble, and simulations

For our model-based estimate of the anthropogenically altered carbon budget of the North Atlantic, we define the North Atlantic by combining four of the regions utilized by Mikaloff Fletcher et al. (2006), namely, the North Atlantic high latitudes, the North Atlantic midlatitudes, the North Atlantic low latitudes, and the North Atlantic tropics (regional boundaries are outlined in Fig. 1). As a model ensemble, we employ 11 ESMs that participated in phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012). A very brief description of the ESMs considered here, and their abbreviations can be found in Table 1. The reader is referred to the listed references therein for further details. Some ESMs performed multiple realizations of the CMIP5 experiments considered here (a description of these experiments can be found in the next paragraph); that is, the same experiment was calculated with different initial states, initialization methods, or physical details. Whenever this was the case, we used the output of the first realization (labeled “r1i1p1”) only. A description of the nomenclature for multiple realizations is listed in Taylor et al. (2012).

To calculate carbon uptake, storage rate, inventory, and transport by the ocean for the period 1850–2005, we utilize results from the historical simulation. When considering future scenarios, we focus solely on the experiment with the strongest atmospheric CO₂ increase [representative concentration pathway (RCP8.5), covering the period 2006–99], as emissions are currently just above this scenario (Sanford et al. 2014). For a first-order model drift correction, we consider furthermore the preindustrial control simulation (experiment piControl, covering the period 1850–2009). All experiments are described in detail in Taylor et al. (2012). The piControl experiment is forced with radiative conditions from the year 1850 throughout the entire simulation time, while the historical experiment uses prescribed historical changes in greenhouse gases, aerosols, and land use, as well as solar and volcanic forcing. The future scenario RCP8.5 assumes that emissions continue to rise throughout the twenty-first century ( unabated emissions). Hence, the piControl simulation represents the natural component of the carbon cycle and its variability without the influence of anthropogenic climate change, while the historical simulation followed by RCP8.5 represents the contemporary and future components of the carbon cycle, respectively. Subtracting the piControl simulation from the RCP8.5 experiment or the historical simulation can hence be equated to the anthropogenic component of the carbon cycle combined with climate change–induced differences. All variables calculated in this manner are henceforth marked by the subscript “ant*”: 

\[ X_{\text{ant}*} = X_{\text{historical/RCP8.5}} - X_{\text{piControl}} \]  

where \( X \) represents the different carbon budget related quantities considered here. The variable \( X_{\text{ant}*} \)
represents the total human alteration of the carbon cycle. We therefore use the notation ‘anthropogenically altered’ when describing or analyzing any of these properties. All quantities are already corrected for a first-order model drift.

b. Considered quantities of the anthropogenically altered carbon budget

Our analysis of the carbon budget focuses on (i) the anthropogenically altered carbon uptake, that is, the air–sea flux of \( C_{\text{ant}*} \) into the ocean [abbreviated as \( C_{\text{ant}*}\text{-uptake} \); (ii) the anthropogenically altered carbon storage rate \( C_{\text{ant}*}\text{-storage-rate} \); \( C_{\text{ant}*}\text{-st.-r.} \); and inventory \( C_{\text{ant}*}\text{-inventory} \)] and their distributions with depth; and (iii) the anthropogenically altered carbon difference \( C_{\text{ant}*}\text{-difference} \) [explained below]. Considering \( C_{\text{ant}*}\text{-storage-rate} \) and \( C_{\text{ant}*}\text{-inventory} \), we chose the upper 100 m as one depth range, representing surface concentrations. Further, we assume 1000-m depth as the horizon that separates between \( C_{\text{ant}*}\) stored in the upper and lower limbs of the Atlantic meridional overturning circulation (AMOC). This choice is justified, as most models show the maximum of their AMOC streamfunction at approximately 1000-m depth [see, e.g., Fig. 6 of Zhang and Wang (2013)]. Since the AMOC streamfunction is calculated as the northward mass transport integrated from surface to the depth considered, the decrease of the streamfunction with depth indicates a shift from northward to southward mass transports. Moreover, a modeled mixed layer depth of 1000 m or more is an indicator for deep convection in the North Atlantic (Heuzé et al. 2015), which allows \( C_{\text{ant}*} \) to be sequestered into the deep ocean. Long-term oceanic sequestration of carbon only takes place when carbon is transported below 1000 m on average (Primeau 2005).

Because of different model grids and different specifications of biological processes, as well as the unavailability of some model output data, the calculation of both the exact anthropogenically altered carbon transport via ocean circulation and the anthropogenically altered biologically mediated carbon sequestration is challenging, or in case of some models, impossible with the available output. We deal with this challenge by...

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**Table 1. Earth system models employed in this analysis.**

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>CESM1(BGC)</td>
<td>Community Earth System Model, version 1 (biogeochemistry)</td>
<td>Long et al. (2013)</td>
</tr>
<tr>
<td>GFDL-ESM2M</td>
<td>Geophysical Fluid Dynamics Laboratory Earth System Model with Modular Ocean Model (MOM), version 4 component</td>
<td>Dunne et al. (2012, 2013)</td>
</tr>
<tr>
<td>GFDL-ESM2G</td>
<td>Geophysical Fluid Dynamics Laboratory Earth System Model with Generalized Ocean Layer Dynamics (GOLD) component</td>
<td>Dunne et al. (2012, 2013)</td>
</tr>
<tr>
<td>HadGEM2-CC</td>
<td>Hadley Centre Global Environment Model, version 2 (Carbon Cycle)</td>
<td>Martin et al. (2011)</td>
</tr>
<tr>
<td>HadGEM2-ES</td>
<td>Hadley Centre Global Environment Model, version 2 (Earth System)</td>
<td>Martin et al. (2011)</td>
</tr>
<tr>
<td>IPSL-CM5A-LR</td>
<td>L’Institut Pierre-Simon Laplace Coupled Model, version 5A, low resolution</td>
<td>Dufresne et al. (2013)</td>
</tr>
<tr>
<td>MIROC-ESM-CHEM</td>
<td>Model for Interdisciplinary Research on Climate, Chemistry Coupled Earth System Model</td>
<td>Watanabe et al. (2011)</td>
</tr>
<tr>
<td>MIROC-ESM</td>
<td>Model for Interdisciplinary Research on Climate, Earth System Model</td>
<td>Watanabe et al. (2011)</td>
</tr>
<tr>
<td>MPI-ESM-LR</td>
<td>Max Planck Institute Earth System Model, low resolution</td>
<td>Giorgetta et al. (2013)</td>
</tr>
<tr>
<td>NorESM1-ME</td>
<td>Norwegian Earth System Model version 1 with carbon cycling (intermediate resolution)</td>
<td>Tjiputra et al. (2013)</td>
</tr>
</tbody>
</table>
by calculating the anthropogenically altered carbon difference:

$$C_{\text{ant\*-diff.}}(t, m, r)(\text{PgC yr}^{-1}) = C_{\text{ant\*-st.-r.}}(t, m, r)(\text{PgC yr}^{-1}) - C_{\text{ant\*-upt.}}(t, m, r)(\text{PgC yr}^{-1}).$$ (2)

Here, $t$ denotes the considered point in time, $m$ indicates a specific model, and $r$ indicates a specific region (please note that the $C_{\text{ant\*-diff.}}$-difference is calculated on a regional basis and not per grid point). Positive values indicate that there is regionally more $C_{\text{ant\*}}$ stored in the ocean than was taken up from the atmosphere, and hence, that $C_{\text{ant\*}}$ was added to this region via transport, whereas negative values indicate the opposite. For the time span of the historical simulation, $C_{\text{ant\*-uptake}}$ and $C_{\text{ant\*-storage-rate}}$ represent an approximation of anthropogenic carbon uptake and anthropogenic carbon storage rate, respectively (Frölicher et al. 2015); thus, the $C_{\text{ant\*-diff.}}$-difference represents, first and foremost, the lateral transport of anthropogenic carbon. However, under RCP8.5, the climate-induced changes in both natural and anthropogenic carbon components are nonnegligible (Bernardello et al. 2014), such that the $C_{\text{ant\*-budget}}$ is not only influenced by changes in ocean circulation, but also by changes in biological production.

3. The anthropogenically altered carbon budget of the North Atlantic

Results for the North Atlantic reveal that all models simulate less $C_{\text{ant\*-uptake}}$ than $C_{\text{ant\*-storage-rate}}$.
(Fig. 2); hence, $C_{\text{ant}^*}$ is added to the North Atlantic via physical transport, as illustrated by the $C_{\text{ant}^*}$-difference [consistent with a previous model study by Tjiputra et al. (2010)]. Table 2 exhibits that the multimodel estimates of contemporary $C_{\text{ant}^*}$-uptake, $C_{\text{ant}^*}$-storage-rate, and $C_{\text{ant}^*}$-difference are slightly lower than the corresponding inverse estimates [Mikaloff Fletcher et al. (2006), hereinafter MF06; their estimate is briefly described in the appendix]. The multimodel mean of the $C_{\text{ant}^*}$-inventory is also lower than the Global Ocean Data Analysis Project version 2 (GLODAPv2) estimate [Table 3; the GLODAPv2 estimate is described in the appendix and in Lauvset et al. (2016)]. This discrepancy is present for all depth intervals considered here (Table 3).

For the period 1850–1990, the models agree relatively well for all considered $C_{\text{ant}^*}$-quantities (Fig. 2). Beyond 1990, the model spread for all $C_{\text{ant}^*}$-quantities increases gradually (Fig. 2, Tables 2, 3). There are particularly large differences in the simulated $C_{\text{ant}^*}$-uptake for the 2090s, ranging from a $C_{\text{ant}^*}$-uptake that is 2–3 times higher than in the 1990s to a $C_{\text{ant}^*}$-uptake that has approximately the same strength as in the 1990s. We henceforth denote models that show a high, moderate, or low $C_{\text{ant}^*}$-uptake for the 2090s as FUhigh-models, FUmoderate-models, and FUlow-models, respectively (where FU is an abbreviation for “future uptake”). FUhigh-models are marked in different shades of blue, FUmoderate-models in shades of green, and FUlow-models in shades of red (as seen in Fig. 2). FUhigh-models are not only characterized by a high future $C_{\text{ant}^*}$-uptake, but also by a high future $C_{\text{ant}^*}$-storage-rate and a low future $C_{\text{ant}^*}$-difference (Figs. 2a–c). The opposite is true for FUlow-models, while FUmoderate-models show moderate future $C_{\text{ant}^*}$-storage-rate and $C_{\text{ant}^*}$-difference.

Using the North Atlantic $C_{\text{ant}^*}$-budget, we wish to analyze the mechanisms that determine the future $C_{\text{ant}^*}$-uptake of a model and determine whether it is possible to constrain them with observation-based estimates. Since the focus of our study is to determine the mechanisms leading to the large spread of the future projections, we will focus our analysis mainly on FUhigh- and FUlow-models. For reasons of comprehensiveness, we show results from FUmoderate-models alongside FUhigh- and FUlow-models—in most of the cases, however, without any explicit analysis.

### 4. Key regions for the future anthropogenically altered carbon budget

To determine the origin of the multimodel spread, we consider two subregions of the North Atlantic, namely, the area combining the middle and high latitudes and the area combining the tropics and low latitudes (see Fig. 1). For both subregions, the models agree well within the twentieth century (Fig. 3), and the modeled estimates of $C_{\text{ant}^*}$-uptake, $C_{\text{ant}^*}$-storage-rate, and $C_{\text{ant}^*}$-difference for the 1990s are close to the estimates of MF06 (Fig. 4). For the period 1997–2007, the models simulate the $C_{\text{ant}^*}$-inventory to be $11.50 \pm 1.96$ PgC for the middle and high latitudes and $13.03 \pm 2.91$ PgC for the tropics and low latitudes; that is, most of the modeled estimates are lower than the GLODAPv2 estimates of $14.58$ and $17.86$ PgC, respectively (not shown).

Beyond the 1990s, the model spread is increasing for all $C_{\text{ant}^*}$-quantities in all considered regions. It is evident that the model spread for the $C_{\text{ant}^*}$-uptake originates in the middle and high latitudes (Figs. 3 and 4). For the $C_{\text{ant}^*}$-difference, FUlow-models show higher future values than FUhigh-models in the middle and high

### Table 2. North Atlantic’s anthropogenic carbon budget ($C_{\text{ant}^*}$-budget) for the year 1995 as extracted from Mikaloff Fletcher et al. (2006), as well as North Atlantic’s anthropogenically altered carbon budget ($C_{\text{ant}^*}$-budget) as modeled by our model ensemble (described as multimodel mean and standard deviation). Carbon inventory is expressed in petagrams of carbon (PgC); all other quantities in PgC yr$^{-1}$.

<table>
<thead>
<tr>
<th>Source</th>
<th>Period</th>
<th>$C_{\text{ant}^*}$-uptake</th>
<th>$C_{\text{ant}^*}$-inventory</th>
<th>$C_{\text{ant}^*}$-storage-rate</th>
<th>$C_{\text{ant}^*}$-difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>MF06</td>
<td>1995</td>
<td>0.30</td>
<td>25.5</td>
<td>0.46</td>
<td>0.16</td>
</tr>
<tr>
<td>Model ensemble</td>
<td>1990s</td>
<td>0.29 ± 0.05</td>
<td>21.37 ± 4.26</td>
<td>0.43 ± 0.06</td>
<td>0.14 ± 0.03</td>
</tr>
<tr>
<td>Model ensemble</td>
<td>2090s</td>
<td>0.50 ± 0.25</td>
<td>106.50 ± 19.62</td>
<td>1.15 ± 0.21</td>
<td>0.65 ± 0.10</td>
</tr>
</tbody>
</table>

### Table 3. North Atlantic’s anthropogenic carbon inventory ($C_{\text{ant}^*}$-inventory; PgC) for the year 2002 as extracted from GLODAPv2, as well as North Atlantic’s anthropogenically altered carbon inventory ($C_{\text{ant}^*}$-inventory; PgC) as modeled by our model ensemble (described as multimodel mean and standard deviation).

<table>
<thead>
<tr>
<th>Source</th>
<th>Period</th>
<th>0–100-m depth</th>
<th>100–1000-m depth</th>
<th>1000 m–sea floor</th>
<th>surface–sea floor</th>
</tr>
</thead>
<tbody>
<tr>
<td>GLODAPv2</td>
<td>2002</td>
<td>2.96</td>
<td>13.66</td>
<td>15.83</td>
<td>32.44</td>
</tr>
<tr>
<td>Model ensemble</td>
<td>1997–2007</td>
<td>2.68 ± 0.15</td>
<td>11.16 ± 1.30</td>
<td>10.68 ± 4.43</td>
<td>24.52 ± 4.63</td>
</tr>
<tr>
<td>Model ensemble</td>
<td>2090s</td>
<td>10.58 ± 0.75</td>
<td>49.18 ± 4.55</td>
<td>44.84 ± 15.30</td>
<td>104.60 ± 17.00</td>
</tr>
</tbody>
</table>
latitudes, but lower future values for the tropics and low latitudes. Yet, the middle and high latitudes have a larger multimodel spread than the tropics and low latitudes, making this region the determining factor for the overall model spread of the $C_{\text{ant}}^*$-difference (Figs. 3 and 4). Given the circulation pattern of the North Atlantic, the $C_{\text{ant}}^*$-difference of the middle and high latitudes is dominated by net $C_{\text{ant}}^*$-transport across its southern

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**Fig. 3.** The 10-yr moving averages of anthropogenically altered (a),(f) air–sea carbon uptake (positive values represent uptake by the ocean), (b),(g) oceanic carbon storage rate between surface and 100-m depth, (c),(h) oceanic carbon storage rate between 100- and 1000-m depth, (d),(i) oceanic carbon storage rate between 1000-m depth and ocean floor, and (e),(j) carbon difference as simulated by 11 different CMIP5 models for different regions. Regions are defined as North Atlantic middle and high latitudes (NAM + NAH) and North Atlantic tropics and low latitudes (NAT + NAL). Regional boundaries are illustrated in Fig. 1; color coding as in Fig. 2.
boundary. This boundary is characterized by $C_{\text{ant}}^+$-transport into the region with the upper limb of the AMOC and by $C_{\text{ant}}^+$-transport out of the region with the lower limb of the AMOC. Hence, the difference between $C_{\text{ant}}^+$-transport in the upper and lower limbs of the AMOC is larger in $FU_{\text{low}}$-models than in $FU_{\text{high}}$-models. Positive future values for the $C_{\text{ant}}^+$-difference indicate that the upper limb of the AMOC is the dominant future $C_{\text{ant}}^+$-transport pathway for $FU_{\text{low}}$ models, while values close to zero (as seen for $FU_{\text{high}}$-models) indicate that both limbs of the AMOC transport an equal amount of $C_{\text{ant}}^+$. For the future $C_{\text{ant}}^+$-storage-rate, all models agree well for the first 100 m (Fig. 3a). The model spread only starts to be visible for deeper layers of the North Atlantic Ocean (Figs. 3h,i). In contrast to $C_{\text{ant}}^+$-uptake and difference, the projected model

Fig. 4. Anthropogenically altered carbon properties as simulated by 11 different CMIP5 models for the 1990s and 2090s. Depicted are (top) air–sea carbon uptake (positive values represent uptake by the ocean), (middle) carbon storage rate, and (bottom) carbon difference. Results are presented for North Atlantic middle and high latitudes (NAM + NAH) and tropics and low latitudes (NAT + NAL). Black dots mark the inverse estimates of MF06. Regional boundaries are illustrated in Fig. 1; color coding as in Fig. 2.
spread of the $C_{\text{ant}*}$-storage-rate below 1000-m depth is similarly large for both the middle and high latitudes and the tropics and low latitudes (Figs. 3d,i). As the main carbon entrance portal from surface to the deep ocean is located in the middle and high latitudes, the simulated uncertainties in the $C_{\text{ant}*}$-storage-rate of the deep ocean of the tropics and low latitudes originate in the middle and high latitudes and are transferred to the tropics and low latitudes via the lower limb of the AMOC. Hence, both $C_{\text{ant}*}$-storage-rate and difference point toward transport of $C_{\text{ant}*}$ via the lower limb of the AMOC as a main mechanism responsible for the future model spread. This transport quantity is dependent on how much $C_{\text{ant}*}$ is stored in the lower limb of the AMOC and how vivid the southward mass transport is in this depth range. In the following section, we connect these deep ocean quantities to the $C_{\text{ant}*}$-uptake at the surface and its associated model spread.

5. Mechanisms behind a low or high future oceanic carbon uptake

Based on the results of section 4, we are solely regarding the middle and high latitudes when analyzing $C_{\text{ant}*}$-uptake and related surface processes, but regard the whole North Atlantic when connecting these processes to the deep ocean.

a. Surface ocean

1) OCEANIC CARBON UPTAKE

Oceanic carbon uptake (i.e., net air–sea CO$_2$ flux into the ocean) is dependent on the air–sea gas transfer coefficient $\text{Tr}$ [see, e.g., Takahashi et al. (2009)] and the partial pressure difference of CO$_2$ between air ($p$CO$_2$) and seawater ($p$CO$_2$) (due to the nonseasonality of the atmospheric CO$_2$ forcing). For the sea–air difference $p$CO$_2$ instead of the air–sea difference $p$CO$_2$ and $p$CO$_2$, as this allows for a direct inference to seasonal phasing of $p$CO$_2$ (due to the nonseasonality of the atmospheric CO$_2$ forcing).

For the 1990s, Figs. 6a–e illustrate that $FU_{\text{high}}$ and $FU_{\text{low}}$-models agree relatively well on the annual cycle of SST. However, $FU_{\text{high}}$-models show deep winter mixed layers, yielding high winter surface nutrient concentrations and hence, sufficient supply for biological production, while the opposite is true for $FU_{\text{low}}$-models. As a result, the annual cycle of $p$CO$_2$ of all $FU_{\text{high}}$-models, except GFDL-ESM2G (henceforth, $FU_{\text{high}}(\text{G})$), is inversely phased to that of $FU_{\text{low}}$-models. All $FU_{\text{high}}(\text{G})$-models have a clear minimum in $p$CO$_2$ in June, which coincides approximately with these models’ seasonal maximum in primary production. $FU_{\text{high}}(\text{G})$-models exhibit, furthermore, high $p$CO$_2$ values in winter and spring, when the mixed layer depth deepens and dissolved inorganic carbon (DIC)-rich water upwells. The annual cycle of all $FU_{\text{high}}(\text{G})$-models is hence dominated by biology and mixing. $FU_{\text{low}}$-models, on the other hand, reveal an annual cycle of $p$CO$_2$ that is SST-dominated. Only GFDL-ESM2G has an annual
cycle of $p\text{CO}_2^{\text{sea}}$ that is equally dominated by biology, mixing, and SST.

Observation-based estimates (briefly described in appendix section a) agree best with the seasonal phasing of $p\text{CO}_2^{\text{sea}}$ as simulated by FU$_{\text{high}}$G-models. FU$_{\text{high}}$-models also agree well with the observation-based estimate of primary production but tend to overestimate the nutrient supply. FU$_{\text{low}}$-models underestimate observational estimates of both nutrient concentration and primary production throughout the year and fail to reproduce the observed phasing of the annual cycle of $p\text{CO}_2^{\text{sea}}$. This also holds true for FU$_{\text{moderate}}$-models. For reasons of completeness, we want to add that all considered models show an SST-dominated annual cycle of $p\text{CO}_2^{\text{sea}}$ for the low latitudes and tropics and agree well with observations (not shown).

For the 2090s (Figs. 6f–j), all models simulate an increase in SST and of the amplitude of its annual cycle, compared to the 1990s. The increase in SST is largest for FU$_{\text{low}}$-models. Simultaneously, there is a shallowing of the simulated mixed layer depth with ongoing climate change, but only slight changes in nutrient supply and biological production. Because of this evolution, the annual $p\text{CO}_2^{\text{sea}}$ cycle of FU$_{\text{low}}$-models is still SST-dominated in the 2090s. The summer low of the annual $p\text{CO}_2^{\text{sea}}$ cycle of FU$_{\text{high}}$G-models continues to be biology-dominated, but the SST-driven enhancement leads to a more pronounced autumn high of the annual $p\text{CO}_2^{\text{sea}}$ cycle of FU$_{\text{high}}$-models. In summary, the future annual $p\text{CO}_2^{\text{sea}}$ cycle of FU$_{\text{high}}$G-models is approximately equally determined by biology, mixing, and SST, while the annual cycle of GFDL-ESM2G is slightly SST-dominated.

**Fig. 5.** The 10-yr moving averages of (a) the air–sea difference of the partial pressure of CO$_2$ and (b) the CO$_2$ gas transfer velocity for experiments (left) historical and RCP8.5 and (right) piControl. Results are based on simulations by 11 different CMIP5 models for the North Atlantic’s middle and high latitudes. Regional boundaries are illustrated in Fig. 1; color coding as in Fig. 2.
The annual cycles of $pCO_{2}^{sea}$ reveal inherent mechanisms controlling the strength of the modeled future carbon sink. $FU_{low}$-models show shallow mixed layer depths and little primary production, while the opposite is true for $FU_{high}$-models. In the 2090s, $FU_{high}$-models take up $C_{ant}$ throughout the year with maximum uptake during maximum biological production, whereas $FU_{low}$-models reveal a strong outgassing of $C_{ant}$ during summer due to...
their low biological production and their simulated large future increase in SST.

b. Anthropogenically altered carbon inventory below 1000-m depth

Mixed layer depths and biological production affect not only the annual cycle of $p\text{CO}_2$, but also both the horizontal and the vertical distribution of the $C_{\text{ant}*}$ inventory. The vertical distribution of the $C_{\text{ant}*}$ inventory is also intrinsically connected to the SST evolution, as the AMOC is a key transport process for the transport of both heat and carbon from the surface to the deep ocean (Buckley and Marshall 2016). As the models differ in total $C_{\text{ant}*}$-uptake and in the total amount of $C_{\text{ant}*}$ that was transported into or out of the North Atlantic, they have different $C_{\text{ant}*}$-inventories for each region. To be able to compare their $C_{\text{ant}*}$-inventories directly, we calculate the fraction of the $C_{\text{ant}*}$-inventory that is stored in different depth layers of the North Atlantic. Particularly, the fraction of the $C_{\text{ant}*}$-inventory below 1000-m depth informs us about the amount of $C_{\text{ant}*}$ that has been sequestered into the ocean’s interior.

For the year 2002, $FU_{\text{high}}$-models store a small fraction of $C_{\text{ant}*}$ at the surface (8%–10%) and a large fraction below 1000 m (49%–58%) and, hence, have a very effective carbon sequestration (see Fig. 7). $FU_{\text{low}}$-models show a less effective carbon sequestration, with a $C_{\text{ant}*}$-storage of 11%–15% at the surface and 30%–34% below 1000-m depth. Observational estimates indicate a moderately effective carbon sequestration, with a $C_{\text{ant}*}$-storage of 9% at the surface and 49% below 1000-m depth. We note that the simulated carbon sequestration efficiency is relatively constant with time.

Since we do not only want to compare total numbers, but also to track the pathways of the carbon sequestration, we furthermore calculate the standardized (std) $C_{\text{ant}*}$-inventory (std$C_{\text{ant}*}$-inv.; i.e., $C_{\text{ant}*}$-inventory at one location divided by the total $C_{\text{ant}*}$-inventory—see appendix section b for details). For the period 1997–2007, the standardized $C_{\text{ant}*}$-inventory integrated over all depths is depicted in Fig. 8, while results integrated from 1000-m depth to the sea floor are illustrated in Fig. 9. Figure 8 reveals that the bulk of the standardized $C_{\text{ant}*}$-inventory of $FU_{\text{high}}$-models is stored within the Labrador Sea and west of the Mid-Atlantic Ridge. This distribution is in agreement with the observation-based estimate of GLODAPv2, although the $FU_{\text{high}}$-models overestimate the inventory fraction in the western basins. $FU_{\text{low}}$-models have the bulk of the standardized $C_{\text{ant}*}$-inventory located within the Labrador and Irminger Seas, and their fraction of standardized $C_{\text{ant}*}$-inventory west of the Mid-Atlantic Ridge is small in lower latitudes. A comparison of Figs. 8 and 9 shows that along the northern parts of the deep western boundary current, the bulk of the $C_{\text{ant}*}$-inventory distribution of $FU_{\text{high}}$-models is located below 1000-m depth. This is also in agreement with the observation-based estimate, although the $FU_{\text{high}}$-models overestimate the inventory fraction that is stored below 1000-m depth. For $FU_{\text{low}}$-models, on the other hand, it is only in the Labrador and Irminger Seas that the bulk of their $C_{\text{ant}*}$-inventory is stored below 1000-m depth; elsewhere, the largest fraction is stored in the upper 1000 m. Furthermore, $FU_{\text{low}}$-models show a steep southward gradient for their inventory below 1000-m depth in the western basins. This implies that the southward $C_{\text{ant}*}$-transport through the lower limb of the AMOC is slow in these models, while the opposite is true for $FU_{\text{high}}$-models.

For the years 2090–99, the standardized $C_{\text{ant}*}$-inventory shows approximately the same distribution (not shown), indicating relatively steady carbon sequestration rates for all models. This is also suggested by Fig. 7.

c. Constraining the model ensemble with surface and deep ocean characteristics

We found in sections 5a and 5b that future differences in the modeled $C_{\text{ant}*}$-uptake are caused by different ocean physics and biology that are seasonally influencing surface DIC concentrations and SST, which are linked through deep convection. A simulated deep mixed layer relates to cooler future SST, high biological production, and an efficient $C_{\text{ant}*}$-transport into the interior ocean, as seen for $FU_{\text{high}}$-models. The seasonal characteristics of $FU_{\text{high}}$-G-models translate into $p\text{CO}_2$ values in the middle and high latitudes that are lower in summer months (here defined as May–October) than annually (see Fig. 6e), while the opposite is true for $FU_{\text{low}}$-models. Hence, both the mean summer $p\text{CO}_2$ anomaly in the middle and high latitudes and the fraction of the $C_{\text{ant}*}$-inventory that is stored below 1000-m depth are good indicators for the future $C_{\text{ant}*}$-uptake strength of a model. This is confirmed by the good correlations between the simulated future $C_{\text{ant}*}$-uptake and both quantities (Fig. 10; please note that the negative mean summer $p\text{CO}_2$ anomaly is used here in order to be able to depict positive correlations). We note that the fraction of the $C_{\text{ant}*}$-inventory that is stored below 1000-m depth shows stronger correlations than the mean summer $p\text{CO}_2$ anomaly. We relate this to the fact that the $C_{\text{ant}*}$-inventory reflects not only deep convection, but also the strength of southward mass transport below 1000-m depth. This is consistent with our finding of section 4 that the lower limb of the AMOC is one of the main mechanisms responsible for the future $C_{\text{ant}*}$-uptake model spread.
Next, we use observations to constrain the model ensemble. For the mean summer $pCO_2$ anomaly, the observation-based estimate of Landschützer et al. (2015) yields a value of $-16.64$ μatm (1 atm = 101.325 Pa; the negative value is illustrated with red lines in Figs. 10a,b), and we assume an associated uncertainty of ±28 μatm (gray shades in Figs. 10a,b; see appendix section c). The GLODAPv2-based estimate of the fraction of the $C_{ant*}$-inventory that is stored below 1000-m depth is 49% (illustrated with red lines in Figs. 10c,d), and we estimate an associated uncertainty of ±10% (gray shades in Figs. 10c,d; see appendix section c). Five models perform within the observational uncertainty of both quantities, namely, NorESM1-ME, GFDL-ESM2M, GFDL-ESM2G, CESM1(BGC), and MPI-ESM-LR (see Figs. 10a–d). These models build our constrained model ensemble. We note that our error estimates are very rudimentary and most likely upper estimates. More accurate error estimates would result in an even more constrained ensemble. Figure 11 illustrates the associated $C_{ant*}$-budget for our constrained ensemble. In comparison with the unconstrained ensemble, it shows higher mean values for both future $C_{ant*}$-uptake (constrained 0.70 ± 0.22 PgC yr$^{-1}$, unconstrained 0.50 ± 0.25 PgC yr$^{-1}$) and storage rate (1.30 ± 0.21 vs 1.15 ± 0.21 PgC yr$^{-1}$), while the mean value for the $C_{ant*}$-difference remains approximately the same (0.60 ± 0.09 vs 0.65 ± 0.10 PgC yr$^{-1}$). Most importantly, the constrained ensemble projection shows a significant slowdown of the $C_{ant*}$-uptake only in the second half of the twenty-first century.

**FIG. 7.** Fraction of the anthropogenically altered carbon inventory that is stored between (a) 0 and 100 m, (b) 100 and 1000 m, and (c) 1000 m and sea floor. Results are based on 10-yr moving averages as simulated by 11 different CMIP5 models for the North Atlantic. Black triangles mark the GLODAPv2 observational estimates. Color coding as in Fig. 2.
6. Discussion

Our findings suggest that the projected $C_{ant}$ uptake in the North Atlantic is dependent on the modeled efficiency of the biologically and physically mediated carbon sequestration and the future increase in SST and that an indicator of these quantities can be found (i) in the mean summer $pCO_2^{sea}$ anomaly in the middle and high latitudes and (ii) in the simulated fraction of the $C_{ant}$-inventory that is stored below 1000-m depth in the North Atlantic. Other studies have pointed toward the annual cycle as an indicator of a models' future performance (e.g., Knutti et al. 2006; Kessler and Tjiputra 2016; Mongwe et al. 2016), as seasonal evolution is
partly driven by the same processes as long-term evolution (Knutti et al. 2006). While most models show good agreement for the mean contemporary annual carbon uptake, they show large disagreement for the associated seasonal cycle (e.g., Lenton et al. 2013; Kessler and Tjiputra 2016). We find that for the middle and high latitudes of the North Atlantic, FU_{high}-models have a different contemporary seasonal $pCO_2$ phasing than FU_{low}-models. Kessler and Tjiputra (2016) found a corresponding result for the Southern Ocean for an ensemble of nine ESMs: here, models with a strong future carbon uptake have a different contemporary seasonal phasing than models with a weak future carbon uptake. Both studies find that the phasing (in terms of either $pCO_2$ or carbon uptake) of the models with a strong future carbon uptake are in agreement with
observational estimates, although the amplitude of the seasonal cycle is overestimated. This indicates that models with a high future carbon uptake emphasize the right mechanisms and that the future carbon uptake might be higher than currently estimated by the IPCC (2013).

The mean summer $pCO_{2}^{sea}$ anomaly in the middle and high latitudes is strongly connected to the mixed layer...
Anthropogenically altered carbon budget [Pg C/yr]

**a)** Uptake

**b)** Storage-rate

**c)** Difference

![Figure 11](image)

**Fig. 11.** The 10-yr moving averages of anthropogenically altered (a) air–sea carbon uptake (positive values represent uptake by the ocean), (b) oceanic carbon storage rate, and (c) carbon difference for the North Atlantic as simulated by our full ensemble of 11 different CMIP5 models (dashed red lines: mean values; orange shades: standard deviation), as well as by the observationally constrained ensemble of five CMIP5 models (blue lines: mean values; light blue shades: standard deviation).

Sgubin et al. (2017) found that the mixed layer depth is crucial for determining the SST evolution in the subpolar North Atlantic. They state that a model’s ability to accurately project SST is related to its ability to accurately simulate the contemporary winter stratification. Each CMIP5 model that they studied was assigned a skill score between zero and one, depending on the accuracy of the modeled contemporary winter stratification. We extract the skill scores for our model ensemble and find that FU_{high}-models have skill scores between 0.63 and 0.99, while FU_{low}-models have skill scores between 0.08 and 0.13. Sgubin et al. (2017) find that the most skilled models project a much more moderate warming or even cooling trend over the North Atlantic than the least skilled models. Yet, they also find that even the most skilled models simulate too-warm and salty surface conditions for the contemporary subpolar gyre. The study of Sgubin et al. (2017) confirms that FU_{high}-models are in better agreement with observations than FU_{low}-models, although there is no perfect agreement to be found.

The usage of the fraction of the $C_{\text{ant}}$-inventory of the North Atlantic that is stored below 1000-m depth is a new approach to estimate the efficiency of the carbon sequestration. It has the disadvantage of offering no means to separate the contributions of biology and physics. It offers, however, an integrated measure of the combined strength of sinking organic particles and deep convection, as well as of the transport along the lower limb of the AMOC. The latter is important, as observations about the transport strength of the lower limb of the AMOC are rare. The observation-based standardized $C_{\text{ant}}$ inventory of the North Atlantic below 1000-m is in good agreement with schematics of the deep limb of the AMOC (Gary et al. 2011).

### 7. Summary and conclusions

We investigated the North Atlantic carbon budget in terms of uptake, storage rate, inventory, and transport of carbon by the oceans for an ensemble of 11 Earth system models. All considered models participated in the carbon uptake projections of the latest report of the Intergovernmental Panel on Climate Change (IPCC 2013). We focused on the period 1850–2099 and investigated the historical experiment (period 1850–2005) as well as the high CO$_2$ future scenario RCP8.5 (period 2006–99).

We found that there is large uncertainty in the future anthropogenically altered carbon budget: that is, the changes caused by rising atmospheric CO$_2$ and climate change (marked by the subscript “ant*”). The model spread in $C_{\text{ant}}$-uptake is due to the contrast between (i) models showing a late flattening of their $C_{\text{ant}}$-uptake and a high future uptake (FU_{high}-models) and (ii) models with an early flattening of their $C_{\text{ant}}$-uptake and a low future uptake (FU_{low}-models). FU_{high} models simultaneously show a high future $C_{\text{ant}}$-storage-rate and a low future $C_{\text{ant}}$-difference (i.e., difference between $C_{\text{ant}}$-uptake and $C_{\text{ant}}$-storage-rate). The opposite is true for FU_{low}-models.

Further analysis revealed that the model spread in the $C_{\text{ant}}$-uptake originates in middle and high latitudes. Here, the annual cycle of pCO$_2$\textsubscript{eq} pointed toward model mechanisms that are responsible for different model behavior: while it is mainly driven by winter mixing and biology for FU_{high} models, it is mainly SST-driven for...
FU\textsubscript{low} models. The deep winter mixing and high primary production of FU\textsubscript{high} models enable an efficient carbon sequestration into the deep ocean. We found that the simulated mean summer pCO\textsubscript{sea} anomaly in the North Atlantic middle and high latitudes is well correlated to the future C\textsubscript{ant*-uptake}. Another indicator that is tightly correlated to a model’s future C\textsubscript{ant*-uptake} is the fraction of the simulated C\textsubscript{ant*-inventory} that is stored below 1000-m depth. Here, FU\textsubscript{high}-models exhibit not only an efficient C\textsubscript{ant*-drawdown} in the Labrador and Irminger Seas (as indicated by their simulated deep mixed layers), but furthermore an efficient southward transport of C\textsubscript{ant*-out} of the middle and high latitudes via the lower limb of the Atlantic meridional overturning circulation. The C\textsubscript{ant*-difference} confirms that FU\textsubscript{high}-models transport the most C\textsubscript{ant*-out} of the middle and high latitudes. The opposite is true for FU\textsubscript{low}-models: their C\textsubscript{ant*-inventory points toward a less efficient carbon sequestration into the deep ocean.

Our analysis shows that the response of the considered models to future climate change depends on mechanisms that are relatively well constrained by available observations: that is, the annual cycle of pCO\textsubscript{sea} in the middle and high latitudes of the North Atlantic and the fraction of the carbon inventory that is stored below 1000-m depth. We showed that observation-based estimates falsify the SST-driven annual cycle of pCO\textsubscript{sea}, as well as the low inventory rate below 1000-m depth, as simulated by FU\textsubscript{low}-models. A constrained model-ensemble was created, which includes only models that are close enough to current observational estimates of both performance indicators. The constrained model ensemble showed that a significant slowdown of the C\textsubscript{ant*-uptake} occurs only in the second half of the twenty-first century. This opens up the question of whether regarding the unconstrained multimodel mean and standard deviation as the best possible estimates for the ocean’s future carbon uptake is an overly simplified approach. While different uncertainty estimates would have allowed for several constrained ensembles, the exclusion of FU\textsubscript{low}-models would always result in a delayed flattening of the future C\textsubscript{ant*-uptake}. Hence, we propose to put into place simple processes that evaluate the models’ performance with respect to crucial mechanisms such as, for example, the annual cycle of pCO\textsubscript{sea} in the middle and high latitudes of the North Atlantic. Several other model studies found similar performance indicators for other climate-relevant variables. While it is debatable if it is good practice to reject a model from the Model Intercomparison Project, it seems beneficial to group models into different performance categories whenever performance indicators are available. An indication of multimodel mean and standard deviation for each of the subensembles can help climate scientists to be of better service for policymakers.

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**APPENDIX**

**Observational Data and Analysis Methods**

**a. Observation-based estimates**

To compare our model-based estimate and the associated mechanisms, we use several observation-based estimates. For anthropogenic carbon, we focus on the anthropogenic carbon budget estimates of Mikaloff Fletcher et al. (2006) and a climatology of the anthropogenic CO\textsubscript{2} inventory from GLODAPv2 (Lauvset et al. 2016). Mikaloff Fletcher et al. (2006) estimate anthropogenic CO\textsubscript{2} uptake, transport, and storage for 24 oceanic regions by using the Green’s function inversion method to combine database estimates of anthropogenic CO\textsubscript{2} storage with information about ocean transport and mixing from a suite of ocean general circulation models. The gridded version of GLODAPv2 is a mapped climatology of ocean biogeochemical variables with a horizontal resolution of 1° × 1° for 33 standard depth surfaces. It is based on quality-controlled data for the period 1972–2013 (Olsen et al. 2016). The climatology of the anthropogenic CO\textsubscript{2} inventory has the reference year 2002 and is based on an application of the transit time distribution method (e.g., Waugh et al. 2006) on all available dichlorodifluoromethane (CFC-12) data in GLODAPv2 (Olsen et al. 2016).
For the verification of mechanisms associated with the anthropogenically altered carbon budget, we utilize the data product of Landschützer et al. (2015), which consists of monthly fields of oceanic partial pressure of CO$_2$ (pCO$_2$) at 1° × 1° resolution created with a two-step neural network technique and different observational estimates. Further, we employ objectively analyzed climatological fields from version 2 of World Ocean Atlas 2013 with a horizontal resolution of 1° × 1° for in situ temperature (Locarnini et al. 2013) as well as nutrients (Garcia et al. 2014). Last, a satellite-based estimate of oceanic net primary productivity (NPP) was employed to generate a monthly NPP climatology for the years 1997–2007. The NPP data are calculated with the vertically generalized production model (Behrenfeld and Falkowski 1997), using input from the Sea-Viewing Wide Field-of-View Sensor (SeaWIFS), and was downloaded from the Ocean Productivity website (http://www.science.oregonstate.edu/ocean.productivity/index.php).

b. Standardized oceanic inventory of anthropogenically altered carbon

To be able to compare the efficiency directly, we standardize the C$_{ant}$-inventory by dividing it gridpoint and modelwise through the total amount of C$_{ant}$ stored in the North Atlantic:

\[
\text{stdC}_{ant}\text{-inv.}(t, m, g)(\%) = \frac{\text{C}_{ant}\text{-inv.}(t, m, g)(\text{mol})}{\sum_g \text{C}_{ant}\text{-inv.}(t, m, g)(\text{mol})} \times 100. \quad (A1)
\]

Here, t denotes the considered point in time, m indicates a specific model, and g indicates a grid point within the North Atlantic region. As fractional measures are strongly influenced by different model grid resolutions, we interpolate all model outputs first to a regular 1° × 1° grid before computing the standardized C$_{ant}$-inventory.

c. Uncertainty estimates

We utilize observation-based estimates of Landschützer et al. (2015) for the mean summer pCO$_2$ anomaly in the middle and high latitudes and of GLODAPv2 for the fraction of the C$_{ant}$-inventory that is stored below 1000-m depth to constrain our model ensemble. Both observation-based estimates offer no detailed error estimates, which are needed for a precise constraint. However, Landschützer et al. (2015) offer a comparison of their database with Surface Ocean CO$_2$ Atlas version 2 [SOCATv2; Bakker et al. (2014)] and LDEO version 2013 (Takahashi et al. 2014), yielding root-mean-square errors (RMSEs) between 13.8 and 26.9 μatm for the period 1990–99 [see Table S1 of Landschützer et al. (2015)]. Moreover, Landschützer et al. (2014) compare an earlier version of their estimate with observations from different time series stations, resulting in RMSEs between 11 and 28 μatm. Because of a lack of a better uncertainty estimate, we take 28 μatm as an upper error estimate (gray shaded areas in Figs. 10a,b).

GLODAPv2 only introduces a mapping error and not an overall error of the estimate. Here, we use the overall error of 29%, as introduced for the C$_{ant}$-storage of the North Atlantic by Steinfeldt et al. (2009). If we assume a systematic error and add an uncertainty of 29% to the C$_{ant}$-storage of the upper 1000-m depth (16.62 + 4.82 = 21.44 PgC) while we subtract an uncertainty of 29% from the C$_{ant}$-storage below 1000-m depth (15.83 – 4.59 = 11.24 PgC), then the fraction of the C$_{ant}$-inventory that is stored below 1000-m depth is 34%, and hence, the associated uncertainty is 15%. A systematic error of 29% that is subtracted from the C$_{ant}$-storage of the upper 1000-m depth and added to the C$_{ant}$-storage below 1000-m depth gives 63% for the fraction of the C$_{ant}$-inventory that is stored below 1000 m and an associated uncertainty of 14%. Evidently, the fraction of the C$_{ant}$-inventory that is stored below 1000 m is a relatively stable measure, where a systematic error of 29% is translated into an error of 14%–15%. As such a strong anticorrelation with depth seems unlikely for the error (Vázquez-Rodríguez et al. 2009), we reduce the uncertainty associated with the fraction of the C$_{ant}$-inventory that is stored below 1000 m to 10% (gray shaded areas in Figs. 10c,d).

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