The impact of the North Atlantic Oscillation on the uptake and accumulation of anthropogenic CO_2 by North Atlantic Ocean mode waters

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Received 9 June 2010; revised 3 May 2011; accepted 10 June 2011; published 21 September 2011.

[1] The North Atlantic Ocean accounts for about 25% of the global oceanic anthropogenic carbon sink. This basin experiences significant interannual variability primarily driven by the North Atlantic Oscillation (NAO). A suite of biogeochemical model simulations is used to analyze the impact of interannual variability on the uptake and storage of contemporary and anthropogenic carbon (C_{anthro}) in the North Atlantic Ocean. Greater winter mixing during positive NAO years results in increased mode water formation and subsequent increases in subtropical and subpolar C_{anthro} inventories. Our analysis suggests that changes in mode water C_{anthro} inventories are primarily due to changes in water mass volumes driven by variations in water mass transformation rates rather than local air-sea CO_2 exchange. This suggests that a significant portion of anthropogenic carbon found in the ocean interior may be derived from surface waters advected into water formation regions rather than from local gas exchange. Therefore, changes in climate modes, such as the NAO, may alter the residence time of anthropogenic carbon in the ocean by altering the rate of water mass transformation. In addition, interannual variability in C_{anthro} storage increases the difficulty of C_{anthro} detection and attribution through hydrographic observations, which are limited by sparse sampling of subsurface waters in time and space.

Citation: Levine, N. M., S. C. Doney, I. Lima, R. Wanninkhof, N. R. Bates, and R. A. Feely (2011), The impact of the North Atlantic Oscillation on the uptake and accumulation of anthropogenic CO_2 by North Atlantic Ocean mode waters, *Global Biogeochem. Cycles*, *25*, GB3022, doi:10.1029/2010GB003892.

1. Introduction

[2] Since the industrial revolution, human activity has released large quantities of carbon dioxide (CO_2), resulting in increased atmospheric concentrations [e.g., *Keeling et al.*, 1976; *Keeling and Whorf*, 1994; *Le Quéré et al.*, 2009]. However, the observed atmospheric increase accounts for only approximately half of anthropogenic carbon emissions [*Canadell et al.*, 2007; *Le Quéré et al.*, 2009; *Sabine et al.*, 2004] (http://www.globalcarbonproject.org). The remaining anthropogenic CO_2 has been taken up by the oceans and terrestrial biospheres, with approximately 25% of anthropogenic CO_2 emissions currently being sequestered in the oceans [*Le Quéré et al.*, 2009, 2010]. The future trajectory of atmospheric CO_2, and the resulting impact on the global climate, is therefore dependent on the magnitude and stability of the ocean and terrestrial carbon sinks [*Canadell et al.*, 2007; *Friedlingstein et al.*, 2006; *Fung et al.*, 2005; *Le Quéré et al.*, 2009].

[3] Over the past three decades, the oceanographic community has devoted significant time and resources to accurately detect both the accumulation of anthropogenic carbon (C_{anthro}) in the ocean and variability in the ocean carbon sink [*Sabine and Tanhua*, 2010]. This is done through full depth surveys of water column dissolved inorganic carbon (DIC) concentrations and anthropogenic tracers and through measurements of surface-ocean partial pressure of carbon dioxide (pCO_2) and air-sea CO_2 fluxes. Several global surveys of the ocean inorganic carbon system have been conducted including the Geochemical Ocean Section Study (GEOSECS) in the 1970s and the World Ocean Circulation Experiment (WOCE)/Joint Global Ocean Flux Study (JGOFS), and Ocean-Atmosphere Carbon Exchange Study (OACES) surveys in the 1990s [*Key et al.*, 2004; *Lee et al.*, 2003; *Sabine et al.*, 2004]. Currently, the U.S. and International Climate Variability and Predictability (CLIVAR)/CO_2 Repeat Hydrography Program is continuing to monitor changes in ocean DIC by resurveying key hydrographic cruises from

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0886-6236/11/2010GB003892

the WOCE/JGOFS era. In addition, approximately 4.8 million measurements of surface water pCO_2 have been made between 1970 and 2006 and assembled into a uniform data set. These prodigious field efforts have provided data to assess the magnitude and variability of the ocean carbon sink [e.g., *Takahashi et al.*, 2009; *Watson et al.*, 2009]. However, while significant headway has been made, the detection of C_{anthro} in the ocean still faces several major challenges.

[4] The C_{anthro} signal is superimposed upon a large dissolved inorganic carbon (DIC) background, with C_{anthro} accounting for only about 5% of the total reservoir of DIC in surface waters in the 1990s [*Sabine et al.*, 2004]. In addition, significant short-term natural variability in the ocean carbon system makes the detection of relatively small, long-term temporal trends in C_{anthro} difficult [*Levine et al.*, 2008]. Several empirical methods have been proposed to deconvolve the C_{anthro} signal from natural variability in the ocean carbon system [e.g., *Brewer et al.*, 1995; *Friis et al.*, 2005; *Gruber et al.*, 1996; *Khatiwala et al.*, 2009; *Matsumoto and Gruber*, 2005; *Touratier and Goyet*, 2004; *Wallace*, 1995; *Waugh et al.*, 2006]. However, each of these methods has their own biases resulting in significant uncertainties in C_{anthro} estimates, particularly in regions of water mass formation [*Levine et al.*, 2008; *Vázquez-Rodríguez et al.*, 2009]. Finally, hydrographic cruises are expensive and provide limited temporal and spatial resolution.

[5] The synergistic use of observations and numerical models has become an important tool for understanding the ocean carbon sink; observations provide an important check of model output, and model results provide a context for interpreting observations and insight into the mechanisms controlling the uptake and accumulation of C_{anthro} [e.g., *Sarmiento et al.*, 1995]. This study uses the output of a global ocean model to investigate the impact of interannual variability on the uptake and storage of C_{anthro} in the North Atlantic Ocean and to provide a context for observations of C_{anthro} in this basin. Nearly 25% of anthropogenic carbon sequestered in the ocean is found in the North Atlantic Ocean, despite the fact that this basin only accounts for 15% of the global area [*Sabine et al.*, 2004]. Changes in uptake and accumulation of carbon in this region therefore have the potential to significantly impact the global inventory of C_{anthro}. In addition, a mechanistic understanding of North Atlantic Ocean dynamics may provide insight into interannual variability in C_{anthro} storage in other basins. Finally, understanding the underlying mechanisms driving C_{anthro} trends will also help improve models used to project future changes in atmospheric CO_2 and global climate.

[6] Mode waters play an important role in the circulation, heat transport, and biogeochemistry of ocean gyres [e.g., *Marshall et al.*, 2009; *McCartney and Talley*, 1982]. These waters form during the winter and carry water with high C_{anthro} into the ocean interior where residence times range from years to many decades [*Wallace*, 2001]. Here we investigate carbon uptake by subtropical and subpolar mode waters, two key water masses for determining variability in North Atlantic Ocean C_{anthro} sequestration in our model. The primary subtropical mode water in the western North Atlantic Ocean is the Eighteen Degree Water (EDW), which forms southeast of the Gulf Stream during the winter months between 33°N–40°N and 30°W–75°W [e.g., *Talley and Raymer*, 1982]. Variability in EDW formation and CO_2 uptake has been shown to correlate with climate modes such as the North Atlantic Oscillation [*Bates et al.*, 2002; *Bates*, 2007; *Gruber et al.*, 2002; *Joyce et al.*, 2000]. However, the driving mechanisms behind these correlations remain unclear [*Bates*, 2007]. A summary of previous findings in the context of this study is presented in section 6.1.

[7] Mode water formation occurs primarily through water mass transformation (formation through diapycnal mixing and surface buoyancy flux) [e.g., *Brambilla et al.*, 2008] and is therefore strongly influenced by wintertime mixed layer depths (MLD) [e.g., *Brambilla and Talley*, 2008; *Marshall et al.*, 1993; *McCartney and Talley*, 1982]. In the subpolar gyre, dense mode water (SPMW) is formed in the northeastern subpolar gyre and travels counterclockwise around the gyre and into the Labrador Sea where some of it ultimately becomes part of the Labrador Sea Water (LSW), a component of North Atlantic Deep Water [*Brambilla et al.*, 2008; *McCartney and Talley*, 1982; *Talley and McCartney*, 1982]. Understanding the properties and history of the densest SPMW, therefore, provides insight into the properties of LSW [*McCartney and Talley*, 1982], an important sink for C_{anthro}. A number of studies have quantified changes in the subpolar ocean carbon sink based on observations of surface pCO_2 [*Corbière et al.*, 2007; *Lefèvre et al.*, 2004; *Omar and Olsen*, 2006; *Schuster and Watson*, 2007; *Takahashi et al.*, 2009; *Watson et al.*, 2009] and profiles of DIC and anthropogenic tracers [*Friis et al.*, 2005; *Olsen et al.*, 2006; *Pérez et al.*, 2008, 2010; *Steinfeldt et al.*, 2009]. The observed temporal rates of change of air-sea CO_2 flux and C_{anthro} inventories, which range in different analyses and for different regions from an increase in the North Atlantic Ocean carbon sink to a decrease in the basin sink, are highly dependent on the time period of the observations. *Thomas et al.* [2008], *Schuster et al.* [2009] and *Watson et al.* [2009] suggest that these differences may be due to limited observational records which alias interannual variability, such as the North Atlantic Oscillation, into estimates of long-term trends.

[8] The North Atlantic Oscillation (NAO) is the major climate mode driving interannual and decadal variability in the North Atlantic Ocean [*Hurrell*, 1995]. During positive NAO years, the combination of a strong atmospheric low pressure over Iceland and a strong atmospheric high pressure in the tropics results in increased storm frequency and current strength in the North Atlantic, Labrador and Nordic Seas [e.g., *Hurrell et al.*, 2003, 2001; *Visbeck et al.*, 2003]. During a negative NAO, the Icelandic atmospheric low pressure system shifts southward resulting in more southerly storm tracks, fewer storms, and weaker currents in the North Atlantic Ocean [e.g., *Hurrell et al.*, 2001]. The shift from a negative to positive NAO has been shown to correlate with a northward migration of the Gulf Stream [*Hurrell and Deser*, 2009; *Marshall et al.*, 2001]. While there is insufficient data to determine the impact of the NAO on MLD in the subpolar and northern subtropical gyres, positive NAO years correlate with decreased sea surface temperatures (SST) in this region indicating that MLD may increase during positive NAO years as there is a negative relationship between SST and MLD [*Hurrell and Deser*, 2009]. The strength of the NAO (the NAO index) is typically measured during the winter months, December–March [*Rogers*, 1984]. The past four decades have seen significant interannual variability in

the NAO index, with the most prominent feature being an extended period of positive NAO from 1988 to 1995. For this study we use the station-based wintertime NAO record for 1970–2004 compiled by James Hurrell (National Center for Atmospheric Research, Boulder CO USA, http://jisao.washington.edu/data_sets/nao/).

[9] Several recent modeling studies have looked at the impact of the NAO on the North Atlantic Ocean carbon sink [*McKinley et al.*, 2004; *Thomas et al.*, 2008; *Ullman et al.*, 2009; *Watson et al.*, 2009]. These studies have focused on interannual changes in air-sea CO_2 flux, surface pCO_2, and upper ocean (100 m) DIC. *Thomas et al.* [2008] propose that changes in the NAO have resulted in significant interannual variability in the subpolar and basin-scale air-sea CO_2 flux. In contrast, *McKinley et al.* [2004] find no significant relationship between air-sea CO_2 flux and the NAO index, which they attribute to long air-sea equilibrium timescales for the carbonate system resulting in a slow response in surface pCO_2. Using the same model as *McKinley et al.* [2004], *Ullman et al.* [2009] find a strong relationship between the first principal component of the subpolar air-sea CO_2 flux and the NAO index and a significantly weaker correlation between pCO_2 and the NAO index. These authors suggest that surface variability in pCO_2 and DIC is driven by changes in vertical transport in the subpolar gyre and by fluctuations in temperature in the subtropical gyre. However, the implication of these surface changes on the thermocline inventory of anthropogenic carbon in the North Atlantic Ocean and the underlying mechanisms driving these changes remains unclear.

[10] Here we use a suite of model simulations to deconvolve the mechanisms driving interannual variability in the North Atlantic Ocean C_{anthro} inventory in the subpolar and subtropical gyres. The NAO plays a significant role in modulating the formation of mode waters and the accumulation of anthropogenic carbon in model simulations. Specifically, shifts from a negative to positive NAO phase result in deeper mixing in mode water formation regions in both gyres. This in turn increases the modeled anthropogenic CO_2 air-sea flux and rates of water mass transformation, thereby increasing C_{anthro} inventories along mode water isopycnals. The impact of NAO shifts on mode water C_{anthro} inventories has important implications for the future of the North Atlantic Ocean carbon sink as climate models predict an increased frequency of positive NAO years [*Meehl et al.*, 2007].

2. Methods

2.1. Model Description

[11] The ocean Biogeochemical Element Cycle (BEC) component of the National Center for Atmospheric Research (NCAR) global Community Climate System Model (CCSM-3) is used for this study [*Doney et al.*, 2009a, 2009b]. The model is noneddy resolving with a grid spacing of 3.6° longitude by 0.8°–1.8° latitude and 25 vertical levels. Mesoscale eddies are parameterized according to *Gent and McWilliams* [1990]. The model contains a full ecological module [*Moore et al.*, 2004] with several phytoplankton function groups and multinutrient limitation including iron limitation. The fourteen main compartments of the ecological module are: small/pico phytoplankton, large phytoplankton/diatoms, nitrogen fixing diazotrophs, zooplankton, suspended and sinking detritus, nitrate, ammonia, phosphate, iron, silicate, oxygen, DIC and alkalinity. The biogeochemistry module [*Doney et al.*, 2006, 2009a] includes full carbonate system thermodynamics, CO_2 air-sea gas exchange, a dynamic iron cycle, and dust deposition from an atmospheric transport model. Neither photosynthesis nor calcification is dependent on CO_2 variables. A full description of the BEC model, including the spin-up and initialization procedure, can be found in the work of *Doney et al.* [2009a, 2009b]. Briefly, the model was spun-up for several hundred years to a quasi-equilibrium state using preindustrial CO_2 (280 ppm) and a repeating annual cycle of atmospheric state variables [*Large and Yeager*, 2004]. The 'repeat annual cycle' is constructed based on a typical year from the hindcast period and maintains realistic high frequency forcings (e.g., storms) consistent with the climatological record.

[12] For this study, we use four companion model simulations that were branched in a consistent fashion from the end point of the spin-up simulation:

[13] 1. Preindustrial CO_2 Repeat Annual Cycle (RAC): The model is forced with the 'repeat annual cycle' used for the long-term spin-up. Atmospheric CO_2 is fixed at preindustrial levels (280 ppm).

[14] 2. Transient CO_2 Repeat Annual Cycle: The model is branched from the quasi-equilibrium steady state in 1870 and run through 2004 forced with the 'repeat annual cycle' and atmospheric CO_2 prescribed following ice core measurements and observations (Pieter Tans, NOAA/ESRL (http://www.esrl.noaa.gov/gmd/ccgg/trends/, download date 2010).

[15] 3. Preindustrial CO_2 Variable Physics (VP): The model is branched from the Preindustrial CO_2 Repeat Annual Cycle in 1958 and run for a historical hindcast simulation from 1958 to 2004 forced with atmospheric reanalysis and satellite data products [*Doney et al.*, 2007, 2009a]. Atmospheric CO_2 is fixed at preindustrial levels (280 ppm).

[16] 4. Transient CO_2 Variable Physics: The model is branched from the Transient CO_2 Repeat Annual Cycle in 1958 and run with the same forcing as Preindustrial CO_2 Variable Physics but with atmospheric CO_2 prescribed following observations (Pieter Tans, NOAA/ESRL (http://www.esrl.noaa.gov/gmd/ccgg/trends/, download date 2010).

[17] For this study we focus on the period of 1970–2004 as these years show the greatest interannual variability in C_{anthro} inventory and have available historical field observations.

2.2. Calculations

[18] Contemporary carbon ($C_{contemp}$) is defined as the DIC concentration for model simulations with increasing atmospheric CO_2. Anthropogenic carbon in the model is determined by subtracting the DIC concentration for paired model simulations with identical physics but with varying and constant atmospheric CO_2:

$$C_{anthro} = C_{contemp} - C_{preindustrial} \quad (1)$$

where $C_{preindustrial}$ is the DIC concentration for the simulation with constant preindustrial atmospheric CO_2 concentrations. The impact of variable ocean physics is investigated by differencing C_{anthro} for the 'Repeat Annual Cycle' (RAC)