Simulated sea-to-air CO_2 flux from 1948 to 2003 using NCEP reanalysis surface fluxes

Arne Winguth[1]
Patrick Wetzel[2] and Ernst Maier-Reimer[2]

[1] Dept. of Atmospheric and Oceanic Science, UW-Madison, 1225 W. Dayton St., Madison, WI 53706
Email: amwinguth@wisc.edu

[2] Max-Planck-Institut fuer Meteorologie, Bundesstr. 55, 20146 Hamburg, Germany.
Email: wetzel@dkrz.de, maier-reimer@dkrz.de

Abstract

A carbon cycle model embedded in a state-of-the-art ocean general circulation and sea ice model has been applied to quantify the important mechanisms of the interannual and decadal sea-to-air CO_2 flux variability and the variations in the uptake of anthropogenic CO_2. The model is forced by daily NCEP/NCAR reanalysis data over a 56 year period of time, showing trends and variability on interannual and decadal scales. The total interannual variability of the model is ±0.50 Pg C yr^{-1}, confirming estimates from previous studies. This is largely dominated by ocean dynamics in the equatorial Pacific, where changes in air-to-sea CO_2 fluxes have a variability of ±0.33 Pg C yr^{-1} and are characterized by wind stress-induced changes in the surface dissolved inorganic carbon concentration. We estimate an average CO_2 flux into the ocean of -1.74 Pg C yr^{-1} for the period between 1990 and 2000, with extremes of -1.20 Pg C yr^{-1} at the La Niña in 1996 and -2.10 Pg C yr^{-1} flux during the El Niños in 1993 and 1998. Overall about 124.3 Pg of anthropogenic carbon have accumulated in the model ocean for the period between 1800 to 2000. This estimate of anthropogenic CO_2 uptake is consistent with a recent study from Sabine et al. (2004).

1. Background

The ocean absorbs carbon dioxide and other trace gases and exchanges them with the atmosphere. These exchanges might interact with the ocean circulation, climate variability (e.g. El Nino/Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), North Atlantic Oscillation (NAO), or Antarctic Circum Polar Wave), and climate change. Estimates of chlorophyll, primary production and temperature on the sea surface from satellite and in situ observations have documented a high variability of the marine carbon cycle in the tropics, the northern high latitudes, and the Southern Ocean. Natural variability and trend of the carbon fluxes in the marine ecosystem are determined by the variability of the circulation, diffusive processes, freshwater and carbon fluxes at the air-sea interface and biogeochemical processes. Estimates of the interannual variability of the sea-to-air CO_2 flux differ considerably:

1. Studies based on variations of the sea surface temperature and CO_2 partial pressure (Feely et al., 1999) indicate a low variability (below ±0.5 Pg C yr-1). This estimate is generally supported

by ocean model simulations (Winguth et al., 1994; Le Quere et al., 2000; Obata and Kitamura, 2003; McKinley et al., 2004). Recent atmospheric inversion estimates (Rödenbeck et al., 2003) show an oceanic interannual variability that agree generally with the ocean model studies.

2. The $\delta^{13}C$ signal and terrestrial studies (Knorr, 2000) clearly indicate a dominant role of the terrestrial biosphere for the variability. However, Keeling et al. (1995) suggest that a considerable sea-air gas exchange, mostly anticorrelated to the terrestrial flux, is needed to explain the atmospheric CO_2 signal.

Here, we reevaluate the prediction of the CO_2 fluxes at the air-sea interface and estimate the oceanic uptake of anthropogenic CO_2 from fossil fuel and land-use change with a state-of-the-art carbon cycle model coupled to an ocean general circulation model and compare the results with previous studies. A detailed description of the work can be found in Wetzel et al. (2004).

2. Model Description

2.1. Ocean general circulation model

To investigate the variability in the marine carbon cycle, we are using the OM (Marseland et al., 2003) developed at the MPI in Hamburg. OM is a z-coordinate free surface global general circulation model based on primitive equations with Boussinesq and hydrostatic approximation. The horizontal model grid is a curvilinear C grid (ranging from 20 to 350 km) with one pole centered over Greenland and the other over Antarctica. The vertical resolution is 20 model levels (layer thicknesses of 20 m to 1400 m). The model runs with a time step of 2.4 hours. Isopycnal diffusion of the thermohaline fields is included as well as eddy-induced tracer transport by the Gent-McWilliams parameterization (Gent et al., 1995). Bottom boundary layer slope convection scheme is considered. The model includes a sea ice model with viscous-plastic rheology (Hibler, 1979).

2.2. Carbon cycle model

The HAMOCC5 is coupled online to the MPI OM, running the same spatial and temporal resolution. The carbon chemistry is based on Maier-Reimer (1993) with air-sea gas exchange parameterized after Wanninkhof (1992). HAMOCC5 includes an ecosystem model that is based on nutrient, phytoplankton, zooplankton and detritus (NPZD; Six and Maier-Reimer, 1996). Nutrient co-limitation of phytoplankton growth is determined by macronutrient concentration of nitrate and phosphate. Siliceous plankton types are limited by available silicate concentration. The growth of phytoplankton is also controlled by micronutrient concentration of iron. The iron cycle model used considers a prescribed atmospheric dust input, an iron uptake by marine organisms using a fixed Fe:P ratio, and a scavenging model based on the concept of Johnson et al. (1997). Export of particles into the deep sea is dependant on particle aggregation (Kriest, 2002). A 10-layer sediment model with pore-water chemistry is similar to Heinze et al. (1999).

3. NCEP experiment set-up

The daily forcing used in the simulation has been derived from the NCEP reanalysis data (Kalnay et al. 1996) of near surface air and dew point temperature, 10 m wind speed, downward shortwave radiation, precipitation, cloudiness and wind stress. The heat fluxes are calculated from the atmospheric forcing data and the actual model distribution of SST and sea ice using a bulk formula. For freshwater, a mass flux boundary condition is implemented, where the actual flux is calculated from prescribed interannual varying precipitation and evaporation, and climatological river runoff. Evaporation is calculated from latent heat fluxes. Additionally, a weak restoring of the surface salinity towards the Levitus climatology (Conkright et al., 1998) is used with a time constant of 386 days. No surface salinity restoring is applied under sea ice.

The model is spun up for 500 years using climatological forcing from the Ocean Model Intercomparison Project (OMIP; http://www.omip.zmaw.de/omip.php) after starting from Levitus climatology as initial condition. Thereafter, the forcing is switched to detrended NCEP data for the period 1948 to 2000 and applied several times subsequently for a total integration of 1700 years.

The biogeochemical model starts with an uniform tracer distribution and tracer adjusted to the circulation after about 1,000 model years with a preindustrial CO_2 concentration of 280 ppm. Two experiments where carried out: A control run from 1765 to 2003 without anthropogenic forcing (CR) and an anthropogenic run (AR) with prescribed atmospheric concentrations from ice cores and from Mauna Loa. In the AR run non-detrended NCEP forcing fields are used for the period from 1948 to 2003.

4. Air-sea CO_2 flux and anthoprogenic CO_2 uptake

The air-sea CO_2 flux partial and the pressure difference of CO_2 between the ocean and the atmosphere (ΔpCO_2) are tightly connected. The mean spatial pattern of ΔpCO_2 is shown in Figure 1 for the reference year 1995 of the "climatological" data compilation from Takahashi et al. (2002). The modeled pattern agrees reasonably well with the observations, but there are local differences. For example, small areas such as the Greenland Sea, Iceland Sea, Norway Sea and Labrador Sea are not well resolved in the course grid of Takahashi et al. (2002).

The effective flux of CO_2 represents the sum of the natural CO_2 fluxes and the uptake of anthropogenic CO_2. Because of the trend of the fluxes of the CR experiment, the total flux of CO_2 into the ocean between 1980 and 2000 is about 0.18 Pg C yr^{-1} smaller than the computed uptake of anthropogenic CO_2. The average CO_2 flux of -1.49 Pg C yr^{-1} for 1980-89 and -1.74 Pg C yr^{-1} for 1990-99 agrees reasonably well with the estimates from atmospheric inversions (Rödenbeck et al., 2003; Gurney et al., 2002) (Table 1).

The inventory of anthropogenic carbon is computed as the difference between the dissolved inorganic carbon (DIC) inventory of the AR and the CR experiment. Storage of anthropogenic CO_2 is high in areas with a low buffer factor (or Revelle factor). Most of the anthropogenic carbon that is taken up in the high latitudes and the upwelling regions is transported to the subtropics, and accumulated mostly in the upper few hundred meters of the water-column. The modeled global spatial distribution and concentrations of anthropogenic DIC are in good agreement with the estimates from other studies with a total uptake for the period from 1800 to

1994 of 104.3 Pg C as compared to a recent estimate of 118 ± 19 Pg C from Sabine et al. (2004) by using inorganic carbon measurements. In the Pacific and Indian Oceans, the simulated inventory almost matches the calculations of Sabine et al. (2002, 1999), however our estimate for the Atlantic is about 25% lower than the calculation by Lee et al. (2003). The difference may be related to the different approaches of Sabine et al. (2002, 1999) and Lee et al. (2003) to estimate the air-sea CO_2 disequilibrium term for shallow surfaces.

5. Variability in the Pacific

The global interannual CO_2 flux variability is controlled by the equatorial Pacific, in agreement with other ocean carbon cycle model studies (Winguth et al., 1994; Le Quere et al., 2000; Obata and Kitamura, 2003; McKinley et al., 2004). In the control run (CR) experiment, the equatorial Pacific from 10°N to 10°S accounts for ±0.33 Pg C yr^{-1}, which is about 65% of the global interannual CO_2 flux variability (Figure 2). The total change of pCO_2 can be calculated (Le Quere et al., 2000; Wetzel et al., 2004) by a change of total dissolved inorganic carbon (DIC), total alkalinity (TALK), temperature (T), and salinity (S):

$$\frac{dpCO2}{dt} = \frac{\partial pCO2}{\partial DIC}\frac{dDIC}{dt} + \frac{\partial pCO2}{\partial TALK}\frac{dTALK}{dt} + \frac{\partial pCO2}{\partial T}\frac{dT}{dt} + \frac{\partial S}{\partial S}\frac{dS}{dt}$$

In the equatorial Pacific, the total change of pCO_2 is mainly dominanted by the change of DIC, and to a lesser degree by temperature. Changes in DIC are mainly affected by the solubility of CO_2 in seawater, the biological production, and changes in the circulation (advection and diffusion). Variability of Ekman induced upwelling by changes in the strength and direction of the trade winds is strongly influenced by the El Nino/Southern Oscillation (ENSO). These changes influence the carbon concentration at surface significantly, because changes in upwelling have an effect on the strong vertical gradient in DIC (Figure 2; Winguth et al., 1994).

A regime shift in 1975-77 is inherent in the NCEP/NCAR reanalysis. Most pronounced is the change in the equatorial Pacific, but it is evident in the Atlantic and the Southern Ocean, too. The interannual flux variability in the equatorial Pacific changes from ±0.33 Pg C yr^{-1} (1948 to 1976) to ±0.27 Pg C yr^{-1} (1977 to 2003), the overturning of both equatorial cells changes from 100±11 Sv to 77±5 Sv. The mean outgasing is reduced from 0.70 Pg C yr^{-1} to 0.58 Pg C yr^{-1}. The transition can also be seen in the Pacific Decadal Oscillation.

6. Variability in the Atlantic

It is generally accepted that the North Atlantic is a strong sink for CO_2. The modeled CO_2 flux and variability of the North Atlantic north of 50°N is -0.31 ±0.04 Pg C yr^{-1} in the CR and -0.41±0.05 Pg C yr^{-1} for 1990-99 in the AR experiment. The model produces a deep penetration of anthropogenic CO_2 in the high northern latitudes of the Atlantic, which is in agreement with estimates by Lee et al. (2003), Gruber (1996) and McNeil et al. (2003). In areas of North Atlantic Deep Water formation, e.g. GIN Sea and Labrador Sea, intense surface cooling and deep winter convection carry the anthropogenic CO_2 signal down to the bottom of the ocean. Accumulation of anthropogenic CO_2 in the GIN Sea is lower than the uptake of anthropogenic CO_2 by air-sea

gas exchange. Most of the water masses transported into the GIN sea originate in the North Atlantic Current and are not fully equilibrated with the atmospheric CO_2 concentrations, and a significant fraction of the anthropogenic CO_2 taken up by air-sea gas exchange is transported out of the GIN Sea with the East Greenland Current and the Denmark Strait overflow. In the Labrador Sea and the North Atlantic south of the GIN Sea and Labrador Sea the in- and out-flows of anthropogenic CO_2 are roughly in balance. Highest rates of anthropogenic CO_2 uptake are simulated in deep convection areas near Greenland. This uptake is reduced by a rising buffer factor, causing only a small increase from 1980-89 to 1990-99.

7. Conclusions

We have used a carbon cycle model coupled to an ocean circulation model, forced with surface fluxes computed from NCEP/NCAR reanalysis data for the period 1948-2003. The natural variability and the trends are analyzed in a control run with preindustrial atmospheric concentrations. In addition, the response of the system to anthropogenic atmospheric CO_2 concentrations have been tested. The total interannual variability of the model is ±0.50 Pg C yr^{-1}. This is within the range of previous ocean model studies (Le Quere et al., 2000; Obata and Kitamura, 2003; McKinley et al., 2004) and recent atmospheric inversions (Rödenbeck et al., 2003). The CO_2 flux is -1.49 Pg C yr^{-1} for 1980-89 and -1.74 Pg C yr^{-1} for 1990-99. Our simulated fluxes agree well with flux estimates from atmospheric inversions (Rödenbeck et al., 2003; Gurney et al., 2002) (Table 1).

We define the uptake of anthropogenic CO_2 as the difference between the CO_2 fluxes of the anthropogenic run and the control run. The uptake of anthropogenic CO_2 is 1.65 Pg C yr^{-1} for 1980-89 and 1.91 Pg C yr^{-1} for 1990-1999, with an accumulation of about 105 Pg of anthropogenic carbon by 1995. The carbon inventory estimate is compatible with estimates from chlorofluorocarbons (McNeil et al., 2003) and from the ΔC^* technique (Gruber et al., 1996; Lee et al., 2003; Sabine et al., 2002 and 1999). A shift in 1975-76 reduces the outgasing of CO_2 in the equatorial Pacific in the control run from 0.70 to 0.58 Pg C yr^{-1}. Nevertheless the total sea-to-air CO_2 flux increases by 0.006 Pg C per year on average. This trend mainly originates in the Southern Ocean and is caused by increasing wind stress. This trend is the reason why the total flux of CO_2 is about 0.18 Pg C yr^{-1} smaller than the computed uptake of anthropogenic CO_2 in the periods 1980-89 and 1990-99.

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8. References.

Conkright, M., S. Levitus, T. O'Brien, T. Boyer, J. Antonov, and C. Stephens, World Ocean Atlas 1998 CD-ROM Data Set Documentation., Tech. Rep. 15, NODC Internal Report, Silver Spring, MD, 16pp., 1998.

Feely, R.A., R. Wanninkhof, T. Takahashi, P. Tans, Influence of El Nino on the equatorial Pacific contribution to atmospheric CO_2 accumulation, Nature, 398, 597, 1999.

Gent, P.R., J. Willebrand, T. McDougall, and J.C. McWilliams, Parameterizing eddy-induced tracer transport in ocean circulation models, J. Phys. Oceanogr., 25, 463-474, 1995.

Gruber, N., J. L. Sarmiento, and T. F. Stocker, An improved method for detecting anthropogenic CO_2 in the oceans, Global Biogeochem. Cycles, 10, 809-837, 1996.

Gurney, K. R., et al., Transcom 3 inversion intercomparison: Model mean results for the estimation of seasonal carbon sources and sinks, Global Biogeochem. Cycles, 18, GB1010, doi:10.1029/2003GB002111, 2002.

Heinze, C., D. Archer, E. Maier-Reimer, and A. Winguth, A global oceanic sediment model for long-term climate studies, Global Biogeochem. Cycles 13, 221-250, 1999.

Hibler, W.D., A dynamic thermodynamic sea ice model, J. Phys. Oceanogr., 9, 815-846, 1979.

Houghton, J. T., Y. Ding, D.J. Griggs, M. Noguer, P. J. van der Linden and D. Xiaosu (Eds.) Climate Change 2001: The Scientific Basis, Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change (IPCC), Cambridge University Press, UK., 2001.

Johnson, K. S., R. M. Gordon, and K. H. Coale, What controls dissolved iron concentrations in the world ocean?, Mar. Chem., 57, 137-161, 1997.

Kalnay E., et al., The NCEP/NCAR 40-Year reanalysis proj, Bull. Amer. Meteor. Soc., Volume 77, No.3, 437-470, 1996.

Keeling, C.D., T.P. Whorf, M. Wahlen, and J. van der Plicht, Interannual extremes in the rate of rise of atmospheric carbon dioxide since 1980, Nature, 375, 666-670, 1995.

Knorr, W., Annual and interannual CO_2 exchanges of the terrestrial biosphere: process-based simulations and uncertainties, Ecology & Biogeography, 9, 225-252, 2000.

Kriest, I., different parmetrizations of marine snow in a 1D-model and their influence on representation of marine snow, nitrogen budget, and sedimentation, Deep-Sea Res., I , 49, 2133-2162, 2002.

Lee, K., et al., An updated anthropogenic CO_2 inventory in the Atlantic Ocean, Global Biogeochem. Cycles, 17(4), 1116, doi:10.1029/2003GB002067, 2003.

Le Quere, C., J.C. Orr, P. Monfray, and O. Aumont, Interannual variability of the oceanic sink of CO2 from 1979 through 1997, Global Biogeochem. Cycles, 14, 1247-1265, 2000.

Maier-Reimer, E., Geochemical cycles in an ocean general circulation model: Preindustrial tracer distribution, Global Biogeochem. Cycles, 7, 645-677, 1993.

Marsland S.J., H. Haak, J.H. Jungclaus, M. Latif and F. Roeske, The Max-Planck-Institute global ocean/sea ice model with orthogonal curvilinear coordinates, Ocean Modeling Volume 5, Issue 2, 91-127, 2003.

McKinley, G.A, M. J. Follows, and J. Marshall, Mechanisms of air-sea CO_2 flux variability in the equatorial Pacific and the North Atlantic Global Biogeochem. Cycles, 18, doi:10.1029/2003GB002179, 2004.

McNeil, B.I, R.J. Matear, R.M.Key, J.L. Bullister, and J.L. Sarmiento, Anthropogenic CO_2 Uptake by the ocean based on the global chlorofluorocarbon data set, Science, 299, 2003.

Obata, A., and Y Kitamura, Interannual variability of sea-air exchange of CO_2 from 1961 to 1998 with a global ocean circulation biogeochemistry model, J. Geophys. Res., 108(C11), 3337, doi:10.1029/2001JC001088, 2003.

Rödenbeck, C., S. Houweling,M. Gloor, and M. Heimann, CO_2 flux history 1982-2001 inferred from atmospheric data using a global inversion of atmospheric transport, Atmos. Chem. Phys., 3, 1919-1964. 2003.

Sabine, C. L., R. M. Key, K. M. Johnson, F. J. Millero, A. Poisson, J. L. Sarmiento, D. W. R. Wallace, and C. D. Winn, Anthropogenic CO_2 inventory of the Indian Ocean, Global Biogeochem. Cycles, 13, 179-198, 1999.

Sabine, C. L., R. A. Feely, R. M. Key, J. L. Bullister, F. J. Millero, K. Lee, T.-H. Peng, B. Tilbrook, T. Ono, and C. S. Wong, Distribution of anthropogenic CO2 in the Pacific Ocean, Global Biogeochem. Cycles, 16(4), 1083, doi:10.1029/2001GB001639, 2002.

Sabine, C.L., et al., The oceanic sink for anthropogenic CO_2, Science, 305, 367-371, 2004.

Six, K.D. and E. Maier-Reimer, Effects of plankton dynamics on seasonal carbon fluxes in an ocean general circulation model, Global Biogeochem. Cycles, 10, 559-583, 1996.

Takahashi, T.,S.C. Sutherland, C. Sweeney, A. Poisson, N. Metzel, B. Tilbrook, N. Bates, R. Wanninkhof, R.A. Feely,C. Sabine, J. Olafsson, Y. Nojiri, Global sea-air CO_2 flux based on climatological surface ocean pCO2, and seasonal biological and temperature effects, Deep-Sea Research II, 49, 1601-1622, 2002.

Wanninkhof, R., Relationship between wind speed and gas exchange over the ocean, J. Geophys. Res., 97, 7373-7382, 1992.

Wetzel, P., A., Winguth, and E., Maier-Reimer, Sea-to-air CO2 flux from 1948 to 2003 - a model study. Global Biogeochem. Cycles, (in revision 2004).

Winguth, A.M.E., M. Heimann, K.D. Kurz, E. Maier-Reimer, U. Mikolajewicz, and J. Segschneider, El Niño-Southern Oscillation related fluctuations of the marine carbon cycle, Global Biogeochem. Cycles 8, 39-63, 1994.

	Time span	South (≤20°S) Pg C yr^{-1}	Tropic Pg C yr^{-1}	North (≥20°N) Pg C yr^{-1}	Total Pg C yr^{-1}
This Study (CR control)	1980-1999	-0.11	1.18	-0.88	0.18
This Study (AR anthropogenic)	1980-1989	-0.9	0.65	-1.26	-1.5
	1992-1996	-1	0.54	-1.32	-1.78
	1996-1999	-1.03	0.6	-1.29	-1.72
	1995	-1.05	0.64		-1.75
Rödenbeck et al. (2003)	1980-1989				-1.2±0.3
	1990-1996	-1.0±0.1	0.9±0.2	-1.6±0.1	-1.7±0.3
	1996-1999	-1.2±0.2	1.1±0.2	-1.7±0.1	-1.7±0.4
Gurney et al. (2002)	1992-1996	-0.9±0.7	0.5±0.6	-1.1±0.4	-1.5±0.4
Takahashi et al. (2002)	1995	-1.51	0.9	-1.03	-1.64
Houghton et al. (2001)	1980-1989				-1.9±0.6
	1990-1999				-1.7±0.5

Table 1. Sea-to-air flux of CO_2 into the northern hemisphere, tropics and southern hemisphere in Pg C yr^{-1}.

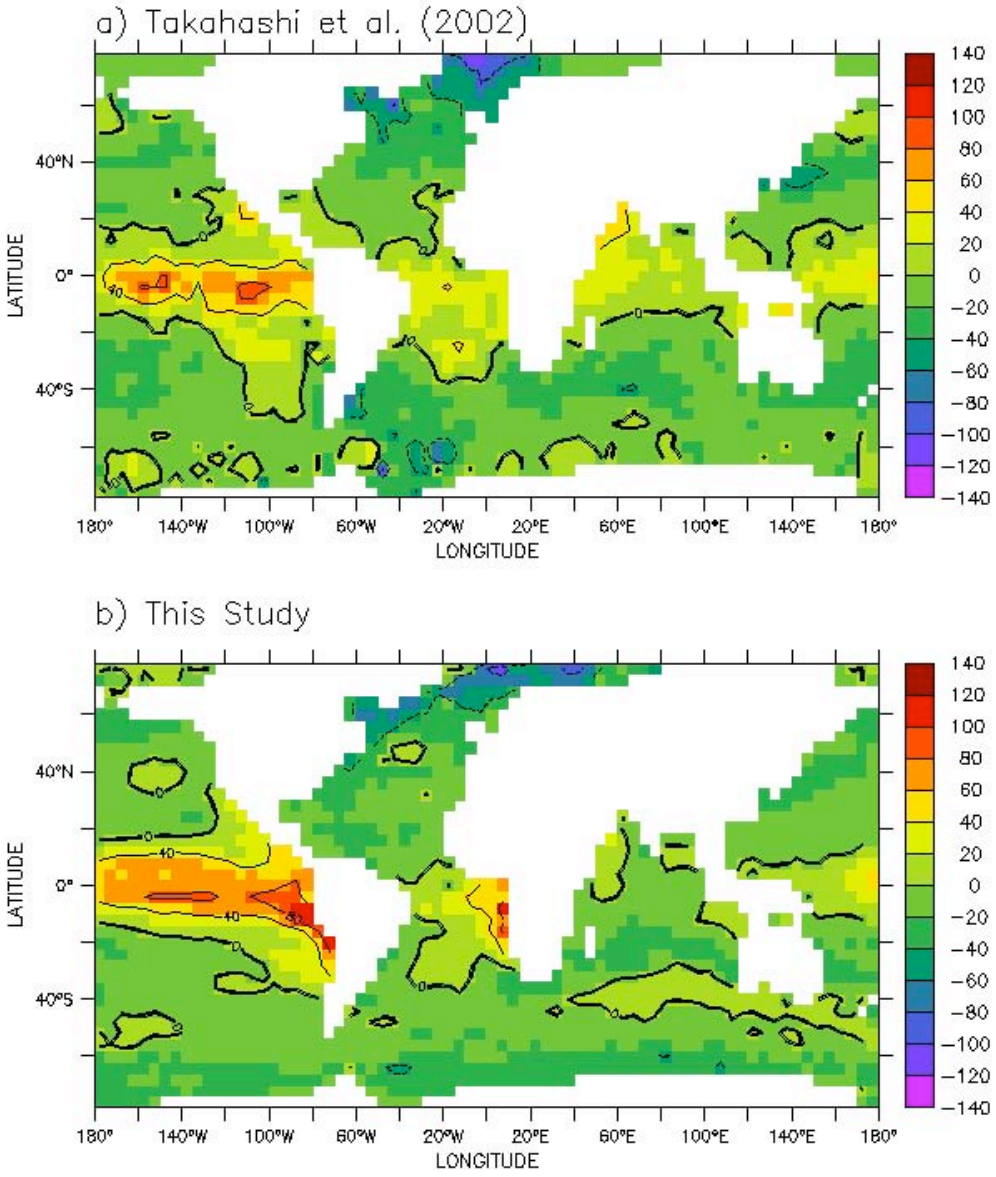

Figure 1. A) Mean ΔpCO2 values from Takahashi et al. (2002), computed for the reference year 1995 in ppm. B) Mean ΔpCO2 of the year 1995 of the anthropogenic run in ppm; interpolated to the 4° x 5° grid of Takahashi et al. (2002).

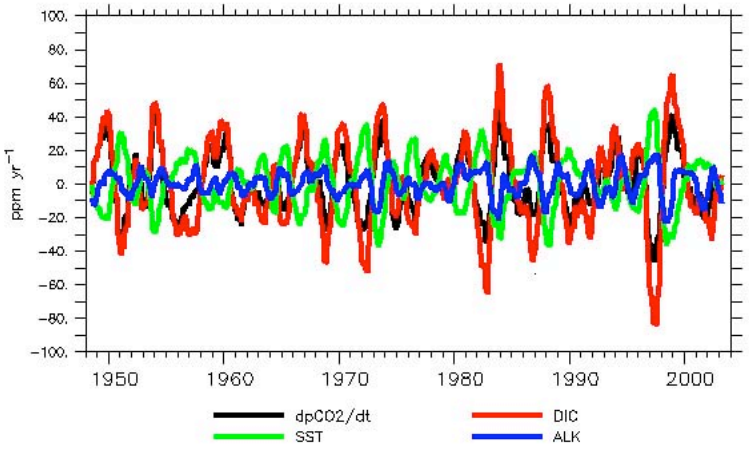

Figure 2. Component analysis for the terms of equation (1) in the surface box of the equatorial Pacific from 10°S to 10°N and 80°W to 135°E. All fluxes are from the control run and smoothed with a 12 month running mean.

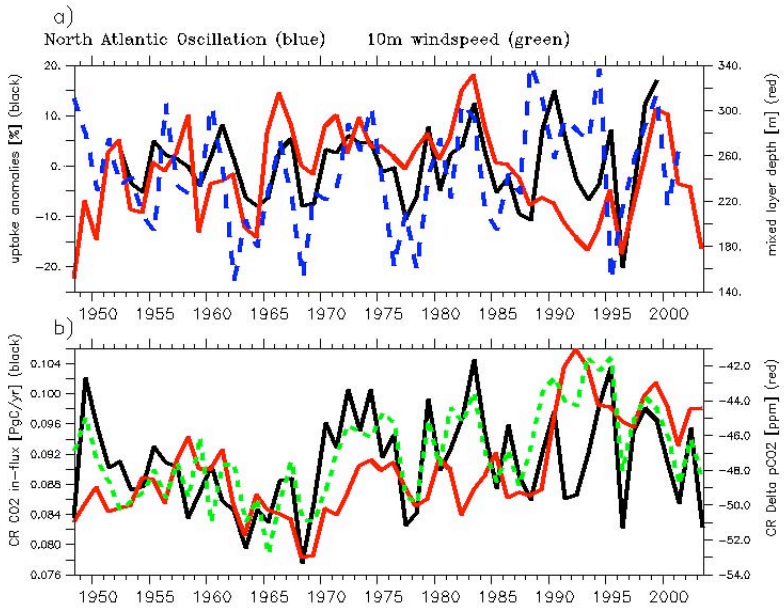

Figure 3. Yearly means for Greenland, Iceland and Norwegian Sea. In blue is the Arctic Oscillation Index and in green is the 10 m wind speed from the NCEP reanalysis. A) Anomaly in the uptake of anthropogenic CO_2 (AR experiment) and the average mixed layer depth. B) CO_2 in-flux and ΔpCO2 from the control run (CR).