Centennial-scale interactions between the carbon cycle and anthropogenic climate change using a dynamic Earth system model

A. Winguth

Center for Climatic Research, Department of Atmospheric and Oceanic Sciences, Madison, Wisconsin, USA

U. Mikolajewicz, M. Gröger, E. Maier-Reimer, G. Schurgers, and M. Vizcaíno

Max-Planck-Institut für Meteorologie, Hamburg, Germany

[1]   A complex Earth system model including atmosphere, ocean, ice sheets, marine carbon cycle and terrestrial vegetation was used to study the long-term response (100–1000 yrs) of the climate to different increased atmospheric CO_2 concentrations. A 3.2 K global mean surface temperature increase is simulated for a $3 \times CO_2$ experiment. The freshwater input by melting of the Greenland Ice Sheet due to global warming is of minor importance compared to hydrological changes in the atmosphere. Increased equatorial upwelling enhances the tropical outgassing of CO_2 from the oceans, lowering the total marine carbon uptake by 16–22%. On land, carbon release due to increase in soil temperature reduces the anthropogenic carbon uptake from CO_2 fertilization up to 43%. Thus, we show that both marine and terrestrial carbon cycle have a positive feedback on climate, which has to be considered for future carbon emission scenarios. **Citation:** Winguth, A., U. Mikolajewicz, M. Gröger, E. Maier-Reimer, G. Schurgers, and M. Vizcaíno (2005), Centennial-scale interactions between the carbon cycle and anthropogenic climate change using a dynamic Earth system model, *Geophys. Res. Lett.*, *32*, L23714, doi:10.1029/2005GL023681.

1. Introduction

[2]   The potential for two-way interactions and feedbacks between the carbon cycle and the climate system comes from the radiative properties of CO_2 and other greenhouse gases in the atmosphere. Increases of atmospheric CO_2 through anthropogenic emissions from fossil fuel burning, for example, may lead to significant changes in climate including changes in the ocean circulation, and magnitude and extent of continental ice sheets [*Intergovernmental Panel on Climate Change (IPCC)*, 2001]. Whereas the studies using complex 3-dimensional models are typically restricted to the next 100 years and do not include interactive ice sheets, only strongly reduced models [*IPCC*, 2001] have been used to study longer time scales. Recent studies of Earth system models (ESMs) including both terrestrial and marine carbon cycles [*Cox et al.*, 2000; *Dufresne et al.*, 2002; *Govindasamy et al.*, 2005] are controversial regarding feedbacks between climate and biogeochemical pools due to global warming. In this study, feedbacks arising from continental ice sheets and the global carbon cycle are considered by using an ESM including all major components of the Earth system. With simulations of 1000-year length each, we investigate the long-term adjustment of the system to various enhanced atmospheric CO_2 levels. Specifically, we focus on two questions: i) how will a polar fresh water input by integrated changes in surface circulation, continental ice sheets, and enhanced precipitation affect the thermohaline circulation on decadal-to-centennial time scales, and ii) how may these changes in the ocean circulation affect carbon uptake in the ocean and on land.

2. Description of the Earth System Model

[3]   In this study, we apply an Earth system model developed at the Max-Planck-Institute for Meteorology and the University of Wisconsin-Madison (MPI/UW ESM) that includes the dominant physical and biogeochemical components with climatic relevance of the Earth system. These climate components are atmosphere, ocean, land biosphere, oceanic carbon cycle, and cryosphere.

[4]   The physical ocean and atmosphere components of the ESM are the coupled atmosphere-ocean ECHAM3/LSG [*Mikolajewicz and Voss*, 2000]. In the 22-layer ocean component, improved parameterizations of the mixed-layer and eddy-induced tracer transport [*Gent et al.*, 1995; *Visbeck et al.*, 1997] have been implemented. To reduce computational costs, we applied the periodically synchronous coupling technique [*Sausen and Voss*, 1996], in which the surface fluxes from the atmosphere are coupled periodically to the upper ocean boundary. In contrast to the original technique, here a nonlinear energy balance model is used for calculating heat fluxes for the ocean/sea ice model.

[5]   The dynamic vegetation part of the ESM is the Lund-Potsdam-Jena dynamic global vegetation model (LPJ) [*Sitch et al.*, 2003]. Climate data from the atmosphere are used as input for LPJ, which returns monthly land surface fields (surface albedo, roughness length, vegetation and forest fraction) to ECHAM3 every year. The marine biogeochemical model part of the ESM is HAMOCC3 [*Maier-Reimer*, 1993]. Model parameters have been optimized in order to reduce the long-term drift.

[6]   The dynamic continental ice sheet component of the ESM is the three-dimensional thermomechanical ice sheet model SICOPOLIS [*Greve*, 1997]. Atmospheric climatologies (seasonal temperature and precipitation) are used as input for SICOPOLIS which returns surface fields such as glacier mask and orography to ECHAM3 and runoff and glacier calving (sea-level) to LSG every year.
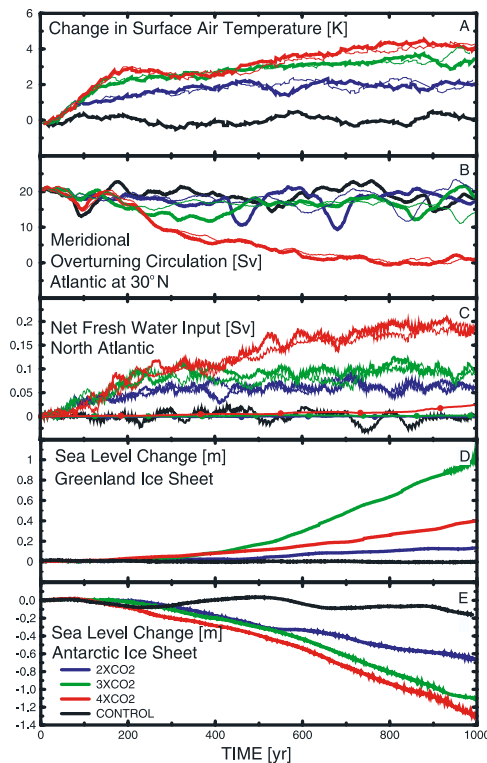
Figure 1. (a) Change in global mean surface temperature, (b) strength of the North Atlantic thermohaline circulation, (c) freshwater transport into the North Atlantic (filled circle: contributions from the Greenland ice sheet), (d) global sea level change by melt water input from Greenland, and (e) Antarctica (black: control run; blue: 2xCO₂; green: 3xCO₂; red: 4xCO₂; thick: with ice sheet feedback; thin: without ice sheet feedback).

[7] A multimillenia control run has been carried out for all model components of the ESM. Analogously to previous studies [*Friedlingstein et al.*, 2003], the following simplified assumptions were made for future emission scenarios: change in land use was not accounted for in LPJ, anthropogenic nitrogen sources were neglected, and sulfate emissions were ignored. Moreover, the ESM does not consider changes in dust, and future changes in the aerosol loading might be of importance for the coupling between climate and marine biosphere [*Jickells et al.*, 2005].

3. CO₂ Stabilization Scenarios

[8] In order to study the sensitivity of the feedbacks between the carbon cycle and the climate, a control experiment and seven atmospheric sensitivity experiments with prescribed atmospheric CO_2 concentrations and with different stabilization levels (as used by *IPCC* [2001]) were conducted. The prescribed CO_2 concentration increases by 1% per year until it stabilizes after 70 years for 2xCO_2, 105 years for 3xCO_2, and 140 years for 4xCO_2, respectively.

[9] Relatively moderate climate sensitivity to changes in the atmospheric CO_2 concentration [*IPCC*, 2001] is indicated by an increase of global mean surface temperature of 2.1 K, 3.2 K, and 4.2 K for the 2xCO_2, 3xCO_2, and 4xCO_2 experiment, respectively (Figure 1a). A drastic change is

predicted for the Arctic Ocean, with a substantial decline in sea ice thickness due to a warming of the atmosphere (Figure 2). Reduced formation of NADW and a decline in the strength of the thermohaline circulation (Figure 1b) are particularly obvious in the 4xCO_2 simulation. In the experiments with 2xCO_2 and 3xCO_2, this reduction in the thermohaline circulation is modest and the strength of the overturning reaches almost the values of the control run after a century, but the North Atlantic overturning cell is shallower than in the control run (not shown). In this scenario ocean convection is intensified in the northeastern Pacific (intermediate depths) by reduced freshwater input from ECHAM3 and Southern Ocean convection is limited to mid-depth with increased ventilation over the last 100 years of integration. Overall, the deep Pacific circulation in the 3xCO_2 run below 2.5 km is substantially reduced and thus prone to low oxygen or anoxic conditions.

[10] The reduction in the North Atlantic thermohaline circulation in all the CO_2 perturbation scenarios is related to an increased stratification, resulting from a positive freshwater anomaly and surface warming (Figure 2). This freshwater perturbation is caused by a strengthening of the atmospheric moisture transports into the North Atlantic and Arctic of up to 0.2 Sv in the case of the 4xCO_2 experiment (Figure 1c). The changes in the surface buoyancy fluxes result in a weakening of the meridional overturning circulation involving all the well-known feedbacks shown in other melt-water experiments, such as reduced tropical salt transport by a more sluggish Gulf Stream [e.g., *Mikolajewicz and Voss*, 2000, and references therein], and additional melt-water input from the Greenland Ice Sheet (Figures 1c and 1d). Comparison of three further experiments, 2x, 3x, and 4xCO_2_NOICE (without ice sheet feedback), clearly reveals that the changes in atmospheric surface buoyancy fluxes are sufficient to bring the system across the threshold for a collapsed thermohaline circulation and that the meltwater input from the Greenland Ice Sheet, which is substantially (roughly a factor of 10) smaller than the changes in atmospheric freshwater input, is of minor importance. The melt water input from the Greenland Ice Sheet is sensitive to changes in air temperatures, with the largest changes predicted for the 3XCO_2 experiment which simulates the warmest surface temperature over Greenland, with an overall contribution to global mean sea level rise of about 1.0 m after 1000 years (Figure 1d). The 2xCO_2 and 3xCO_2 experiment exhibit a rather similar two-dimensional pattern of ice thickness changes in the Greenland Ice Sheet: reduced thickness in the margins due to intensified melting along the shore by the warming and increased thickness in
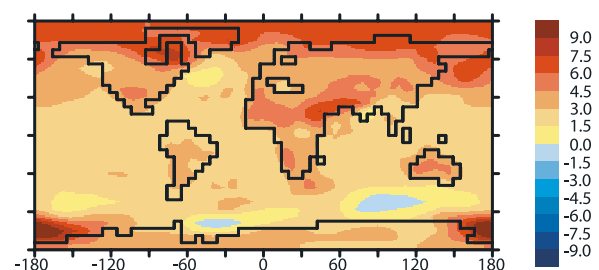

Figure 2. Change in surface temperature [K] for 3xCO_2 minus control (100-year mean for model years 900–1000).
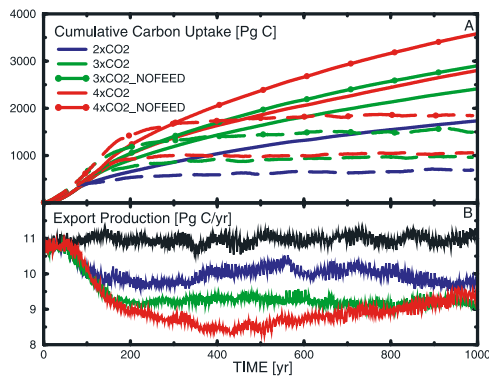
Figure 3. (a) Cumulative carbon uptake by the ocean (solid) and by land (dashed). NOFEED denotes an experiment without climate change due to radiative forcing. (b) Export of carbon from the surface layer into the abyssal layers (black: control run; legend see also Figure 3a).

the interior of North Greenland due to higher snowfall rates. The decrease in Greenland ice volume in the $3xCO_2$ experiment is almost balanced by the increase in volume of the Antarctic Ice Sheet (Figure 1e), which sees enhanced snowfall. In the experiment with $4xCO_2$, the pattern clearly differs from the 2x and $3xCO_2$ experiments. Cooling over the North Atlantic in the $4xCO_2$ experiment leads to a reduction in snowfall and moderate melting rates.

[11] The feedbacks between climate and carbon cycle are most obviously documented by Figures 3 and 4a. CO_2 uptake of the oceans by climate change [*Sabine et al.*, 2004] is significantly reduced because of an increased equatorial gas exchange from the ocean to the atmosphere by increased upwelling.[1] Moreover, a decrease in solubility due to increased stratification and warming in high latitudes reduces the storage of carbon by the oceans. The marine CO_2 storage is also affected by the change in the biological pump, which is remarkably reduced by a decline of the carbon export (16% for experiment $3xCO_2$) from the euphotic layer into the abyssal layers (Figure 3b). The reduced export in the Pacific is somewhat contrary to an increased upwelling in the global warming scenarios. Increased upwelling (e.g., as observed during La Nina events) [e.g., *Feely et al.*, 1999] is commonly associated with an increased nutrient supply from subsurface water. However, in experiment $3xCO_2$ the surface ocean is generally more depleted in nutrients than in experiment CONTROL (where a change in the atmospheric CO_2 concentration does not have an effect on climate change) (Figure 4b). This change of the vertical nutrient gradient is highly related to a change in the high-latitude ventilation of subsurface water masses [*Sarmiento et al.*, 2004], especially in the Pacific Ocean. A stagnant deep sea is prone to nutrient accumulation, since the loss by mixing is much slower than the gain due to the particulate organic carbon flux from the surface. Thus, surface and intermediate water masses are low in nutrients, showing similarities to glacial circulation reconstructions [*Boyle*, 1988]. Accordingly, negative nutrient anomalies are most pronounced in

areas with intense upwelling, i.e., the east Pacific divergence zone and the North Pacific. With relatively low nutrient concentrations in equatorial surface waters, export of carbon into the twilight zone is substantially reduced (Figure 3b), exerting a positive feedback on these negative nutrient anomalies.

[12] The decrease of the export of carbon into the deep-sea in the $4xCO_2$ experiment relative to the control run also affects the rain ratio of $CaCO_3$ to organic carbon with a change up to +0.02 or 20% in the eastern equatorial Pacific. Anthropogenic-induced changes in pH are predicted to be the largest in the Cenozoic (the last 66 million years) [*Zeebe*, 1999]. Our simulations show maximum changes of the pH value at the surface ocean related to the high atmospheric CO_2 concentration and a more stagnant thermohaline circulation [*Caldeira and Wickett*, 2003; *Heinze*, 2004]. The pH value in the Southern Ocean and Antarctic Intermediate Water (Figure 4c) decreases up to -0.7 after 1000 years integration for the $4xCO_2$ experiment, which could enhance carbonate dissolution and changes in calcification of calcareous organisms such as corals, foraminifera, pteropods, or coccolithophores [e.g., *Riebesell et al.*, 2000].

[13] The sensitivity of carbon uptake on land by global change is mainly determined by the CO_2 fertilization of plants, which has a positive effect on carbon storage (Figure 4a), and an increase in soil temperature with overall global warming of the landmasses, which has a negative
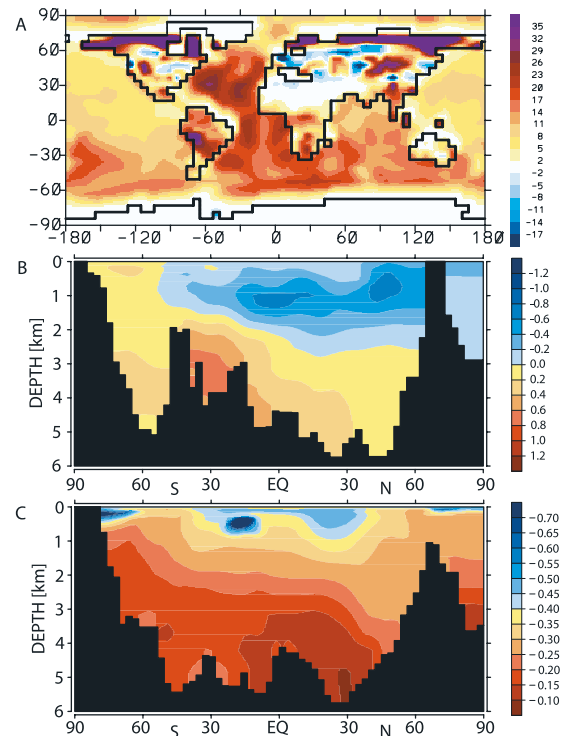

Figure 4. Differences between the $3xCO_2$ and the control run (a) of cumulative uptake of carbon [kg C m^{-2}] in the water column and on the land surface, (b) in a vertical section of phosphate concentration [μmol L^{-1}] in the central Pacific, and (c) in a vertical section of pH value in the western Atlantic (100-year mean for model years 900–1000).

effect. Increase in temperatures causes a poleward shift of the boreal forest cover, with high soil carbon storage in the Northern Hemisphere high latitudes (Figure 4a). Due to increased soil temperatures, respiration of soil carbon is enhanced in the temperate latitudes, which causes a decrease in carbon storage. Changes in soil moisture have a minor influence on carbon storage. The overall changes associated with climate change reduce the carbon uptake of the land biosphere up to 43% (Figure 3a).

[14] The total sensitivity of carbon uptake by the ocean and by the land with respect to global warming is about the same in the initial phase of the CO_2 perturbation. After that, the terrestrial biosphere gets saturated quickly and the total uptake of carbon by ocean and land is primarily determined by the oceans (Figure 4a; 69% over the last 100 years of integration for 4xCO_2). Overall, total carbon uptake of the ocean and the land is reduced by 24–28%, if changes in the atmospheric CO_2 concentration have an effect on climate change (Figure 3a).

4. Summary and Conclusions

[15] In summary, our simulations with the Earth system model confirm previous results that there is a strong feedback between the carbon cycle and the climate. In contrast to previous model simulations, we consider interactions between climate and continental ice sheets. The following main conclusions can be drawn from this study:

[16] 1. Changes in freshwater input from ice sheets by global warming have a secondary impact on variations in the thermohaline circulation and carbon cycle for the first 1000 years. However, uncertainties remain in ice sheet modeling, for example with regard to the role of ice shelves (not modeled in this approach but relevant for the evolution of the Antarctic Ice Sheet) and basal temperate ice layers potentially causing ice sheet instabilities.

[17] 2. Increased equatorial upwelling related to a decrease in the pole-to-equator temperature gradient and to an increase in moisture transport and a reduced biological pump enhance the gas exchange of carbon from the oceans to the atmosphere, lowering the CO_2 storage by the oceans. On land, CO_2 fertilization of plants and carbon emissions by increase in soil temperature reduce the anthropogenic carbon uptake, if changes in the atmospheric CO_2 concentration have an effect on climate change. The importance of feedback of the land biosphere on climate is particular significant in the northward migration of forest over Asia and North America. The reduction of carbon storage due to climate change is in agreement with other studies [*Cox et al.*, 2000; *Dufresne et al.*, 2002; *Govindasamy et al.*, 2005]. However, the reduction due to climate change does not exceed the uptake due to fertilization, as was reported by *Cox et al.* [2000] due to changes in the Amazon region. Uncertainties in the feedbacks between the climate and the marine carbon cycle remain and a more detailed representation of the important processes [*IPCC*, 2001] needs to be subject of future investigations.

References

Boyle, E. A. (1988), Cadmium: Chemical tracer of deepwater paleoceanography, *Paleoceanography*, 3, 471–489.

Caldeira, K., and M. E. Wickett (2003), Anthropogenic carbon and ocean pH, *Nature*, 425, 365–368.

Cox, P. M., R. A. Betts, C. D. Jones, S. A. Spall, and I. J. Totterdell (2000), Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model, *Nature*, 408, 184–187.

Dufresne, J.-L., L. Fairhead, H. Le Treut, M. Berthelot, L. Bopp, P. Ciais, P. Friedlingstein, and P. Monfray (2002), On the magnitude of positive feedback between future climate change and the carbon cycle, *Geophys. Res. Lett.*, 29(10), 1405, doi:10.1029/2001GL013777.

Feely, R., R. Wanninkhof, T. Takahashi, and P. Tans (1999), Influence of El Nino on the equatorial Pacific contribution to atmospheric CO_2, *Nature*, 98, 597–601.

Friedlingstein, P., J.-L. Dufresne, P. M. Cox, and P. Rayner (2003), How positive is the feedback between climate change and the carbon cycle, *Tellus, Ser. B*, 55, 692–700.

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

Govindasamy, B., S. Thompson, A. Mirin, P. Friedlingstein, M. Wickett, K. Caldeira, and C. Delire (2005), Increase of carbon cycle feedback with climate sensitivity: Results from a coupled carbon climate and carbon cycle model, *Tellus, Ser. B*, 57, 153–163.

Greve, R. (1997), Application of a polythermal three dimensional ice sheet model to the Greenland Ice Sheet: Response to steady-state and transient climate scenarios, *J. Clim.*, 10, 901–918.

Heinze, C. (2004), Simulating oceanic $CaCO_3$ export production in the greenhouse, *Geophys. Res. Lett.*, 31, L16308, doi:10.1029/2004GL020613.

Intergovernmental Panel on Climate Change (IPCC) (2001), *Climate Change 2001: The Scientific Basis: Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, edited by J. T. Houghton et al., 881 pp., Cambridge Univ. Press, New York.

Jickells, T. D., et al. (2005), Global iron connections between desert dust, ocean biogeochemistry, and climate, *Science*, 308, 67–71.

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

Mikolajewicz, U., and R. Voss (2000), The role of the individual air-sea flux components in CO_2-induced changes of the ocean's circulation and climate, *Clim. Dyn.*, 16, 627–642.

Riebesell, U., I. Zondervan, P. D. Tortell, R. Zeebe, and F. M. M. Morel (2000), Reduced calcification of marine plankton in response to increased atmospheric CO_2, *Nature*, 407, 364–367.

Sabine, C. L., et al. (2004), The oceanic sink for anthropogenic CO_2, *Science*, 305, 367–371.

Sarmiento, J. L., N. Gruber, M. A. Brzezinski, and J. P. Dunne (2004), High-latitude controls of thermocline nutrients and low latitude biological productivity, *Nature*, 427, 56–60.

Sausen, R., and R. Voss (1996), Techniques for asynchronous and periodically synchronous coupling of atmosphere and ocean models. part I: Impact of variability, *Clim. Dyn.*, 12, 605–614.

Sitch, S., et al. (2003), Evaluation of ecosystem dynamics, plant geography and terrestrial carbon cycling in the LPJ Dynamic Global Vegetation Model, *Global Change Biol.*, 9, 161–185.

Visbeck, M., J. Marshall, T. Haine, and M. Spall (1997), Specification of eddy transfer coefficients in coarse resolution ocean circulation models, *J. Phys. Oceanogr.*, 27, 381–402.

Zeebe, R. E. (1999), An explanation of the effect of seawater carbonate concentration on foraminiferal oxygen isotopes, *Geochim. Cosmochim. Acta*, 63, 2001–2007.

————————————

M. Gröger, E. Maier-Reimer, U. Mikolajewicz, G. Schurgers, and M. Vizcaíno, Max-Planck-Institut für Meteorologie, Bundesstraße 53, D-20146 Hamburg, Germany.

A. Winguth, Center for Climatic Research, Department of Atmospheric and Oceanic Sciences, 1225 W. Dayton St., Madison, WI 53706, USA. (amwinguth@wisc.edu)