GEOPHYSICAL RESEARCH LETTERS, VOL. 34, L19707, doi:10.1029/2007GL030905, 2007

Click Here for Full Article

Long term fate of anthropogenic carbon

Alvaro Montenegro,[1] Victor Brovkin,[2] Michael Eby,[1] David Archer,[3] and Andrew J. Weaver[1]

Received 5 June 2007; revised 29 August 2007; accepted 12 September 2007; published 10 October 2007.

[1] Two earth-system models of intermediate complexity are used to study the long term response to an input of 5000 Pg of carbon into the atmosphere. About 75% of CO_2 emissions have an average perturbation lifetime of 1800 years and 25% have lifetimes much longer than 5000 years. In the simulations, higher levels of atmospheric CO_2 remain in the atmosphere than predicted by previous experiments and the average perturbation lifetime of atmospheric CO_2 for this level of emissions is much longer than the 300–400 years proposed by other studies. At year 6800, CO_2 concentrations between about 960 to 1440 ppmv result in global surface temperature increases between 6 and 8°C. There is also significant surface ocean acidification, with pH decreasing from 8.16 to 7.46 units between years 2000 and 2300. **Citation:** Montenegro, A., V. Brovkin, M. Eby, D. Archer, and A. J. Weaver (2007), Long term fate of anthropogenic carbon, *Geophys. Res. Lett.*, *34*, L19707, doi:10.1029/2007GL030905.

1. Introduction

[2] Most studies of the future impact of anthropogenic CO_2 on the climate system focus their attention on the next few decades, or at most up to the end of the 21st century. These periods are comparable to policy planning and implementation times and comprehensible in terms of human life span. The less scrutinized climate response on longer time scales (10^3 to 10^4 years) can offer insight into pertinent scientific questions, like the onset of the next glaciation [*Archer and Ganopolski*, 2005] or how the dynamics of sediment $CaCO_3$ influence atmospheric CO_2 concentrations. Also, as in the cases of nuclear wastes or human-induced species extinctions [*Novacek and Cleland*, 2001], the depth of the impact in time could be used in the social and political arenas as a way to quantify the seriousness of such impact.

[3] Different processes, with very distinct time scales, are responsible for determining the perturbation lifetime of anthropogenic CO_2. For time scales of decades to centuries, the response to excess CO_2 includes ocean uptake, changes in land carbon associated with increase in soil respiration, CO_2 fertilization and alterations to vegetation cover. In scales of centuries to about 5000–10000 years, ocean uptake becomes dominant, with $CaCO_3$ compensation [*Broecker and Peng*, 1987] playing a significant role in the longer time scales within this interval. Models that have simulated the long time response to the consumption of all available fossil fuel reserves (~5000 PgC [*Rogner*, 1997]) have indicated that roughly 80% of the anthropogenic input has an average perturbation lifetime in the atmosphere of approximately 300–450 years. The remaining 20% could remain in the atmosphere for more than 5 thousand years after emissions cease and the atmosphere would still hold 5% to 10% of the anthropogenic CO_2 hundreds of thousands of years into the future [*Archer et al.*, 1997, 1998; *Archer*, 2005; *Lenton and Britton*, 2006]. Apart from the rate with which CO_2 is removed from the atmosphere, the fraction of the perturbation remaining in the atmosphere depends on the ocean's buffering capacity, which is inversely proportional to the magnitude of the perturbation. The result is that the portion of the anthropogenic CO_2 left in the atmosphere increases as emissions increase [*Sarmiento and Gruber*, 2006; *Archer et al.*, 1997].

[4] Here we use two earth-system models of intermediate complexity (EMIC) to analyze the long term (4500 years after the end of emissions) response of the climate system to an anthropogenic input of 5000 PgC into the atmosphere. The aim is to describe the fate of anthropogenic CO_2, its pathways through the different components of the carbon system and the mechanisms responsible for this partition. The effects of climate change on the carbon cycle are also analyzed.

2. Model Descriptions

[5] CLIMBER-2 is a coupled climate-carbon cycle model with a 2.5-dimensional dynamical-statistical atmosphere model with a spatial resolution of 10° latitude and 51° longitude, a 3-basin, zonally-averaged ocean model and a sea-ice model with latitudinal resolution of 2.5° [*Sitch et al.*, 2005; *Petoukhov et al.*, 2000]. The oceanic carbon cycle includes standard inorganic biogeochemistry [*Brovkin et al.*, 2002] and a marine biota NPZD model [*Six and Maier-Reimer*, 1996]. CLIMBER-2 also includes a dynamic sediment model [*Archer et al.*, 1998] which allows for $CaCO_3$ compensation. Due to its box-like form this model overestimates the deep ocean volume. To correct for this factor in the oceanic biogeochemistry module, a globally-averaged hypsometric function is taken into account in calculation of deep sea floor and volume [*Brovkin et al.*, 2007].

[6] The University of Victoria Earth System Climate Model (UVic ESCM) has horizontal resolution of 1.8° × 3.6°. It consists of a vertically integrated, energy-moisture balance, atmospheric model, coupled to the MOM2 ocean general circulation model with 19 vertical levels and a dynamic-thermodynamic sea-ice model [*Weaver et al.*, 2001]. The terrestrial carbon model is a modified version of the MOSES2 land surface model and the TRIFFID

[1]Department of Earth and Ocean Science, University of Victoria, Victoria, British Columbia, Canada.
[2]Potsdam Institute for Climate Impact Research, Potsdam, Germany.
[3]Department of the Geophysical Sciences, University of Chicago, Chicago, Illinois, USA.

Copyright 2007 by the American Geophysical Union.
0094-8276/07/2007GL030905$05.00

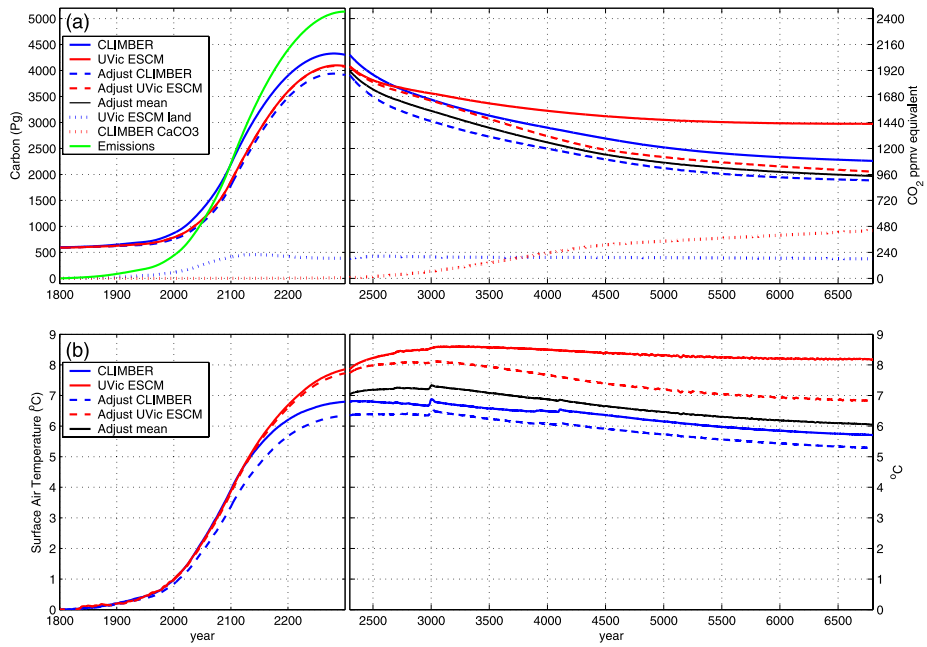

Figure 1. (a) Carbon stocks for the coupled experiments. Emissions, green; atmospheric carbon for CLIMBER-2, solid blue; adjusted CLIMBER-2, dashed blue; UVic, solid red; adjusted UVic, dashed red; average of adjusted values, solid black; UVic land carbon anomaly, dotted blue; CLIMBER-2 carbon from dissolved $CaCO_3$ anomaly, dotted red. Gap indicates change in temporal resolution at year 2300. On the right vertical axis carbon stocks are expressed in terms of equivalent atmospheric CO_2 concentration. (b) Surface air temperature anomaly, same colour coding as in Figure 1a.

dynamic vegetation model [*Meissner et al.*, 2003; *Matthews et al.*, 2005]. Ocean inorganic carbon is based on the OCMIP abiotic protocol. Ocean biology is simulated by an ecosystem model of nitrogen cycling [*Schmittner et al.*, 2005; *Oschlies and Garçon*, 1999]. Water, energy and carbon are conserved with no flux adjustments. It should be noted that in a recent carbon cycling intercomparison project, the response of atmospheric carbon of both CLIMBER and UVic models to anthropogenic CO_2 emissions of about 2200 Pg C were well within the range of the other nine models analyzed [*Friedlingstein et al.*, 2006].

3. Experiments

[7] Both models are integrated for 5000 years starting from equilibrium at year 1800. The simulations are forced with CO_2 emissions prescribed by historical values up to the year 2000 [*Marland et al.*, 2006] and according to the IPCC A2 scenario [*Intergovernmental Panel on Climate Change*, 2000] from 2000 to 2100. After year 2100, emissions decline linearly to zero at 2300, resulting in 5134 Pg of carbon (Pg C) being added to the system (Figure 1a).

[8] To examine the climate-carbon cycle feedback, two distinct integrations are performed with each model. In one set (coupled simulations), the radiative forcing associated with atmospheric CO_2 concentration is taken into account; in the other set (uncoupled simulations), atmospheric CO_2 concentrations have no direct effect on the radiative forcing.

[9] The CLIMBER-2 simulations reported here did not include terrestrial carbon dynamics. A rationale was an absence of land use dynamics and the uncertainty in future land use scenario which can turn land into a sink or a source [*Sitch et al.*, 2005]. While the UVic ESCM model takes into consideration biological $CaCO_3$ dynamics in the water column, it has no ocean sediment component and no input of $CaCO_3$ into the ocean due to weathering.

[10] To compare model results, adjustments are made to both CLIMBER-2 and UVic ESCM atmospheric CO_2. These adjustments are conducted a *posteriori* and do not influence the numerical experiments. The assumption is that the missing components act as sinks to atmospheric carbon.

[11] The CLIMBER-2 results are adjusted by removing from the atmosphere the amount of carbon equivalent to the UVic ESCM terrestrial carbon anomaly. The adjustment to the UVic ESCM consists in reducing its atmospheric CO_2 in direct proportion to the amount of $CaCO_3$ dissolved from CLIMBER-2's sediment and weathering model. This is because dissolution of $CaCO_3$ modifies the alkalinity in such a way as to eventually cause the draw down of an equivalent amount of CO_2. In reality this draw down would take some time to occur, depending on mixing the alkalinity signal from the sediments to the surface, so this is an overestimate (or at least a faster estimate) of the draw down. Both adjustments are time dependent (Figure 1a). The adjusted atmospheric carbon is used to generate adjusted surface air temperature series based on the time dependent atmospheric temperature to carbon ratios (Figure 1b).

[12] Had the processes associated with the adjustments been represented in the simulations the atmospheric CO_2 values during the integrations would have been lower. This in turn would have resulted in smaller oceanic and terrestrial carbon uptake than actually recorded. The simulations then will overestimate carbon draw down and the adjusted values should be considered a lower bound on atmospheric carbon stocks. This is confirmed for the UVic ESCM model by an experiment where the dissolved $CaCO_3$ from CLIMBER-2

Table 1. Comparison to Previous Experiments[a]

Emissions Pg C	CO_2 Year 3000, ppmv	CO_2 Year 6800, ppmv	Land Carbon	Sediment $CaCO_2$	Ocean Dynamics[b]	Reference
5134	1708 [0.58]	1427 [0.47]	yes	no	yes	UVic ESCM
5134	1642 [0.55]	985 [0.29]	yes	†[c]	yes	Adj. UVic ESCM
5134	1617 [0.54]	1063 [0.32]	no	yes	yes	CLIMBER-2
4000	~1200 [0.48]	-	yes	no	yes	[*Lenton et al.*, 2006]
5134	1423 [0.46]	886 [0.25]	†[c]	yes	yes	Adj. CLIMBER-2
4546	~1180 [0.41]	~700 [0.19]	no	yes	no	[*Archer et al.*, 1998]
6579	~1200 [0.38]	-	yes	yes	yes	[*Mikolajewicz et al.*, 2007]
5000	1120 [0.35]	833 [0.23]	no	yes	no	[*Archer*, 2005]
4173	~918 [0.32]	~640[.18]	no	yes	yes	[*Ridgwell and Hargreaves*, 2005]
4546	923 [0.29]	~700 [0.19]	yes	yes	no	[*Lenton and Britton*, 2006]
3817	~343 [0.18]	-	yes	yes	yes	[*Mikolajewicz et al.*, 2007]

[a]Values preceded by ~ were obtained by visual inspection of published figures. Numbers in brackets give the fraction of anthropogenic CO_2 present in the atmosphere.

[b]Ocean dynamics refers to the model's capability of representing changes in ocean state with climate.

[c]These processes are not considered by the model but represented through the adjustments described in the text. The CLIMBER-2 and UVic ESCM results are from the coupled simulations.

was introduced into the UVic ESCM ocean as the simulation progressed. At year 6800, the atmospheric carbon stock of this experiment (not shown) had 400 Pg more carbon than the adjusted UVic ESCM value and the global surface temperature was 0.52°C higher than the adjusted UVic ESCM temperature. The use of these adjustments is a compromise and presents some problems. Most notably, while we show adjusted results for the two models, these are not independent from each other. Also, the adjustments assume that the missing component on each model would have responded to the CO_2 forcing in the same manner as the other model did.

4. Results and Discussion

[13] The uptake by vegetation is characterized by the rapid increase in land carbon anomaly over the 21st century, which reaches a peak of 458 Pg C in year 2142. Between the end of emissions and the year 6800 the UVic ESCM land acts as a small source to the atmosphere, with terrestrial carbon anomaly declining 3% (from 389 Pg to 377 Pg). Both models also have rapid initial oceanic uptake, with ocean carbon increasing by 700 Pg between the years 2000 and 2150 (not shown).

[14] The $CaCO_3$ dissolution modelled by CLIMBER-2 is small at the beginning of the simulation but starts increasing at around year 2500, with peak dissolution rates occurring between years 3500 and 4500. By around year 4500 all sediment $CaCO_3$ has been depleted and any further $CaCO_3$ dissolution is associated with the input from continental weathering.

[15] The introduction of 5000 Pg C as CO_2 into the models generate adjusted global mean surface air temperatures ~6°C warmer and nearly one third of the anthropogenic CO_2 still present in the atmosphere 4500 years after emissions stop (Figure 1b). Based on a exponential decay fit to the average of the adjusted CO_2 curves starting at year 2300, 75% of the anthropogenic CO_2 has an average perturbation lifetime of approximately 1800 years. The removal of the remaining 25% of the anthropogenic input requires silicate weathering, which has an estimated time scale of several hundred thousand years [*Archer*, 2005].

[16] In millennial time scale simulations which account for the effects of both sediment CaCO3 and vegetation dynamics, the anthropogenic CO_2 fraction present in the atmosphere at year 3000 range from 0.18 to 0.55. At year 6800, the range is from 0.19 to 0.29 (Table 1). The values from the UVic ESCM and CLIMBER-2 are at the higher limits of this distribution. The total amount of anthropogenic CO_2, the rate of emissions and the oceanic response to climate vary significantly among the simulations that make up this range, complicating comparisons between their results.

[17] In agreement with previous simulations [*Caldeira and Wickett*, 2003], global mean surface pH decreases, going from the initial state of 8.16 units to 7.42 units(UVic) and 7.41 units (CLIMBER) at the year 2300. It has been estimated [*Orr et al.*, 2005] that surface pH levels of 7.8 to 7.9 would bring the aragonite saturation depth to the surface in the Southern Ocean, with large negative impacts to the biota.

4.1. Effects of Ocean Dynamics

[18] Decreases in CO_2 solubility due to higher oceanic temperatures [*Archer*, 2005] and changes in circulation and water mass distribution are factors influencing the response of atmospheric CO_2 to climate. While present results cannot be used to isolate and quantify the impact of changes in ocean dynamics, they offer insight into some potentially pertinent processes.

[19] One of these relate to changes in the overturning circulation. The Atlantic meridional overturning decreases by about 30% for both models between years 2005 and 2300, going from 19.5 to 13.2 Sv in the UVic ESCM and from 20.4 to 14.7 Sv in CLIMBER. In separate experiments (not shown), complete shut down of the overturning in the UVic ESCM model generates a positive feedback on atmospheric CO_2 on the order of 3%. While smaller overturning rates could have contributed to a slow down in the oceanic invasion of atmospheric CO_2, this was probably not the dominant factor.

[20] As in previous experiments [*Friedlingstein et al.*, 2001; *Bopp et al.*, 2005; *Ridgwell and Hargreaves*, 2005], the warming simulations registered a reduction in the global

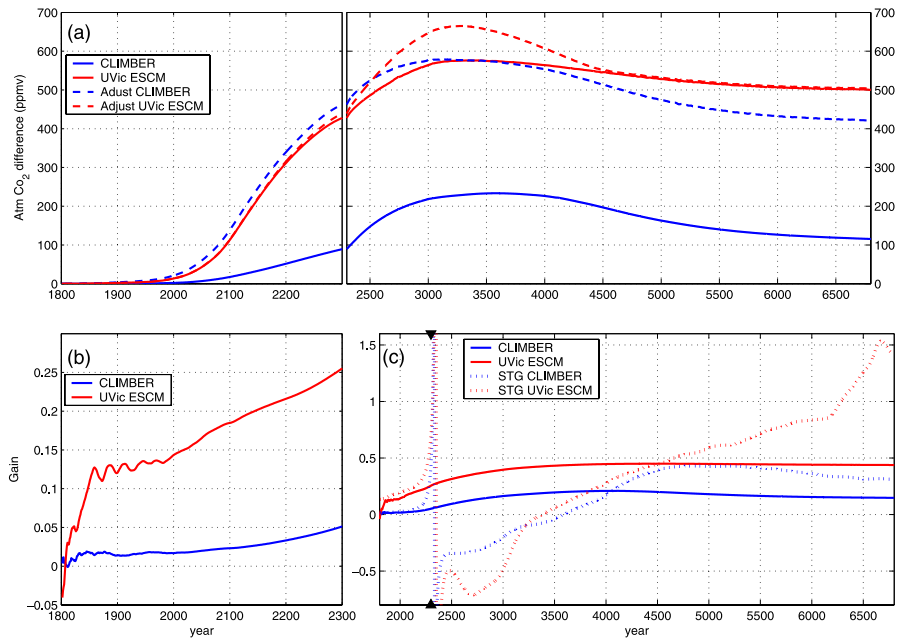

Figure 2. (a) Atmospheric CO_2 concentration difference between coupled and uncoupled simulations, same colour coding as in Figure 1. (b) Atmospheric carbon gain for the first 500 years of the simulation. (c) Atmospheric carbon gain for the whole experiment (solid lines) and the short term gain starting at year 2400 (dotted lines). The black triangles at year 2300 mark the end of the period shown in Figure 2b.

export of near surface particulate organic mater, which decreased by 10% (UVic) and by 4% (CLIMBER) between the years of 2005 and 2300. Part of this change could be related to the stronger stratification in the warmer climate state. In the UVic ESCM experiment, near surface stratification at year 2300 (estimated based on the mean buoyancy frequency of the first 400 m) is 25% to 40% larger than in the year 2000. Stronger stratification leads to less nutrient upwelling and warmer surface temperatures lead to greater nutrient recycling, reducing biological export to deeper waters. An increase in stratification would also reduce the physical downward transport of dissolved inorganic carbon.

4.2. Climate-Carbon Feedback

[21] The climate-carbon feedback of the UVic ESCM model is larger than CLIMBER-2's (Figure 2a), in part due to CLIMBER-2's lack of terrestrial carbon feedbacks which dominate the feedback of the UVic ESCM model during the first few hundred years. Even with its terrestrial carbon cycle, CLIMBER-2 shows a lower overall carbon-climate feedback compared to the UVic ESCM model [*Friedlingstein et al.*, 2006]. The adjusted CLIMBER-2 values (Figure 2a), are similar to the UVic ESCM feedback values.

[22] The larger input of carbon from $CaCO_3$ dissolution of the uncoupled run up to about year 4500 is caused by differences in sediment $CaCO_3$ dissolution and is associated with the higher CO_2 concentrations and the cooler temperatures of the deep ocean in the CLIMBER-2 uncoupled simulations. In both coupled and uncoupled experiments, sediment $CaCO_3$ is depleted at around year 4500. After this time, $CaCO_3$ dissolution is controlled by the terrestrial input from weathering, at the prescribed rate 0.12 Pg C/year. The end of sediment $CaCO_3$ dissolution is reflected in the convergence of the UVic ESCM original and adjusted atmospheric CO_2 differences (Figure 2a).

[23] The gain of the climate-carbon feedback (g) was estimated following [*Friedlingstein et al.*, 2006, equation 1]:

$$g = 1 - \frac{\Delta C_A^u}{\Delta C_A^c} \tag{1}$$

where ΔC_A^u and ΔC_A^c refer, respectively, to the uncoupled and coupled atmospheric CO_2 anomaly. The gain varies in time in all experiments and reaches values larger than 0.4 in the UVic ESCM experiments. (Figures 2b and 2c).

[24] As defined by equation 1, with anomalies referenced to initial CO_2 concentrations, the gain index is not well suited to deal with periods of declining atmospheric CO_2. We estimate the climate gain over the latter portion of the simulation using a slightly different calculation. We define "short term gain" using equation 1 but calculating $\Delta\ C_A^u$ and ΔC_A^c locally, as anomalies over the past 10 years and not referenced to the initial atmospheric concentration. Short term gain values are close to the regular estimate in the beginning of the simulation but tend to positive and then negative infinity as it approaches and goes over the inflection point of maximum atmospheric carbon. During the period of diminishing atmospheric CO_2, the short term gain shows that both UVic ESCM and CLIMBER-2 uncoupled experiments have larger drawdown up to about years 3400–3500 (negative short term gain, Figure 2c). This is caused by larger drawdown due to the lower water temperatures of the uncoupled simulations. After year 3500, the short term gain is positive, a consequence of a reduction of the uncoupled oceanic drawdown and the larger atmospheric concentrations of the coupled experiments.

5. Summary

[25] In experiments that have emissions similar to all known fossil fuel reserves it is estimated that 75% of the

anthropogenic CO_2 has an average perturbation lifetime of ~1800 years with the remaining 25% having average lifetime much longer than 5000 years. This conclusion is not dependent on the adjustments performed on model results. Global temperatures are shown to raise between 6 and 8°C and remain at least 5°C higher than preindustrial for more than 5000 years. These conclusions are supported by two very different EMICs and are considered to be conservative estimates.

[26] The results also suggest that changes in ocean dynamics due to climate change may cause a positive long term feedback for atmospheric CO_2. Over the next 300 years, the modelled carbon gain due to climate-carbon feedback tends to increase, with values at year 2300 that range from ~5% to ~25%. The higher value is mainly due to much lower terrestrial carbon stocks registered under warmer climate.

[27] Our simulations show higher levels of atmospheric CO_2 remaining in the atmosphere longer than predicted by previous modelling experiments [*Archer*, 2005; *Archer and Ganopolski*, 2005; *Lenton and Britton*, 2006]. The average perturbation lifetime of 1800 years is much longer than the 300–450 years proposed by some other studies [*Archer et al.*, 1997; *Archer*, 2005]. Given the large differences in model type and experiment set up between the present and previous experiments, these comparisons should be made with care. While there still is a great deal of uncertainty at these longer timescales, our results indicate that the long term consequences of anthropogenic climate change may be much greater than previously thought.

[28] **Acknowledgments.** We are grateful for funding from CFCAS through its Polar Climate Stability and Operating grant programs.

References

Archer, D. (2005), Fate of fossil fuel CO_2 in geologic time, *J. Geophys. Res.*, *110*, C09S05, doi:10.1029/2004JC002625.

Archer, D., and A. Ganopolski (2005), A movable trigger: Fossil fuel CO_2 and the onset of the next glaciation, *Geochem. Geophys. Geosyst.*, *6*, Q05003, doi:10.1029/2004GC000891.

Archer, D., H. Kheshgi, and E. Maier-Reimer (1997), Multiple timescales for neutralization of fossil fuel CO_2, *Geophys. Res. Lett.*, *24*(4), 405–408.

Archer, D., H. Kheshgi, and E. Maier-Reimer (1998), Dynamics of fossil fuel CO_2 neutralization by marine $CaCO_3$, *Global Biogeochem. Cycles*, *12*, 259–276.

Bopp, L., O. Aumont, P. Cadule, S. Alvain, and M. Gehlen (2005), Response of diatoms distribution to global warming and potential implications: A global model study, *Geophys. Res. Lett.*, *32*, L19606, doi:10.1029/2005GL023653.

Broecker, W., and T. Peng (1987), The role of $CaCO_3$ compensation in the glacial to interglacial atmospheric CO_2 change, *Global Biogeochem. Cycles*, *1*, 15–29.

Brovkin, V., J. Bendtsen, M. Claussen, A. Ganopolski, C. Kubatzki, V. Petoukhov, and A. Andreev (2002), Carbon cycle, vegetation, and climate dynamics in the Holocene: Experiments with the CLIMBER-2 model, *Global Biogeochem. Cycles*, *16*(4), 1139, doi:10.1029/2001GB001662.

Brovkin, V., A. Ganopolski, D. Archer, and S. Rahmstorf (2007), Lowering of glacial atmospheric CO_2 in response to changes in oceanic circulation and marine biogeochemistry, *Paleoceanography*, doi:10.1029/2006PA001380, in press.

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

Friedlingstein, P., L. Bopp, P. Ciais, J. L. Dufresne, L. Fairhead, H. LeTreut, P. Monfray, and J. Orr (2001), Positive feedback between future climate change and the carbon cycle, *Geophys. Res. Lett.*, *28*(8), 1543–1546.

Friedlingstein, P., et al. (2006), Cimate-carbon feedback analysis: Results from the C^4MIP model intercomparison, *J. Clim.*, *19*, 3337–3353.

Intergovernmental Panel on Climate Change (2000), *Special Report on Emissions Scenarios, Working Group III of the Intergovernmental Panel on Climate Change*, Cambridge Univ. Press, New York.

Lenton, T. M., and C. Britton (2006), Enhanced carbonate and silicate weathering accelerates recovery from fossil fuel CO_2 perturbations, *Global Biogeochem. Cycles*, *20*, GB3009, doi:10.1029/2005GB002678.

Lenton, T. M., M. S. Williamson, N. R. Edwards, R. Marsh, A. R. Price, A. J. Ridgwell, J. G. Shepherd, and S. J. Cox (2006), Millennial timescale carbon cycle and climate change in an efficient Earth system model, *Clim. Dyn.*, *26*, 687–711.

Marland, G., T. Boden, and R. J. Andres (2006), Global, regional, and national CO_2 emissions, in *Trends: A Compendium of Data on Global Change*, Carbon Dioxide Inf. Anal. Cent., Oak Ridge Natl. Lab., Oak Ridge, Tenn., available at http://cdiac.ornl.gov/trends/emis/tre_glob.htm.

Matthews, H. D., A. Weaver, and K. J. Meissner (2005), Terrestrial carbon cycle dynamics under recent and future climate change, *J. Clim.*, *18*, 1609–1628.

Meissner, K. J., A. J. Weaver, H. D. Matthews, and P. M. Cox (2003), The role of land-surface dynamics in glacial inception: A study with the UVic Earth System Model, *Clim. Dyn.*, *21*, 515–537.

Mikolajewicz, U., M. Gröger, E. Maier-Reimer, G. Schurgers, M. Vizcaíno, and A. M. E. Winguth (2007), Long-term effects of anthropogenic CO_2 emissions simulated with a complex Earth system model, *Clim. Dyn.*, *28*, 599–633.

Novacek, M. J., and E. E. Cleland (2001), The current biodiversity extinction event: Scenarios for mitigation and recovery, *Proc. Natl. Acad. Sci. U. S. A.*, *98*(10), 5466–5470.

Orr, J. C., et al. (2005), Anthropogenic ocean acidification over the twenty-first century and its impact on calcifying organisms, *Nature*, *437*, 681–686.

Oschlies, A., and V. Garçon (1999), An eddy-permiting coupled physical-biological models of the North Atlantic: part 1. Sensitivity to advection numerics and mixed layer physics, *Global Biogeochem. Cycles*, *13*, 135–160.

Petoukhov, V., A. Ganopolski, V. Brovkin, M. Claussen, A. Eliseev, C. Kubatzki, and S. Rahmstorf (2000), CLIMBER-2: A climate system model of intermediate complexity. Part I: Model description and performance for present climate, *Clim. Dyn.*, *16*, 1–17.

Ridgwell, A., and J. Hargreaves (2005), Regulation of atmospheric CO_2 by deep sea sediments in an Earth system model, *Earth Planet. Sci. Lett.*, *234*, 299–315.

Rogner, H.-H. (1997), An assessment of world hydrocarbon resources, *Annu. Rev. Energy Environ.*, *22*, 217–262.

Sarmiento, J., and N. Gruber (2006), *Ocean Biogeochemical Dynamics*, Princeton Univ. Press, Princeton, N. J.

Schmittner, A., A. Oschlies, X. Giraud, M. Eby, and H. L. Simmons (2005), A global model of the marine ecosystem for long-term simulations: Sensitivity to ocean mixing, buoyancy forcing, particle sinking, and dissolved organic matter cycling, *Global Biogeochem. Cycles*, *19*, GB3004, doi:10.1029/2004GB002283.

Sitch, S., V. Brovkin, W. von Bloh, D. van Vuuren, B. Eickhout, and A. Ganopolski (2005), Impacts of future land cover changes on atmospheric CO_2 and climate, *Global Biogeochem. Cycles*, *19*, GB2013, doi:10.1029/2004GB002311.

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

Weaver, A. J., et al. (2001), The UVic Earth System Climate Model: Model description, climatology and application to past, present and future climates, *Atmos. Ocean*, *39*(4), 361–428.

D. Archer, Department of the Geophysical Sciences, University of Chicago, 5734 S. Ellis Avenue, Chicago, IL 60637, USA.

V. Brovkin, Potsdam Institute for Climate Impact Research, P.O. Box 60 12 03, D-14412 Potsdam, Germany.

M. Eby, A. Montenegro, and A. J. Weaver, Department of Earth and Ocean Science, University of Victoria, P.O. Box 3055, Stn CSC, Victoria, BC, Canada V8W 3P6. (alvaro@uvic.ca)