Holocene carbon cycle dynamics

Thomas Kleinen,[1] Victor Brovkin,[1,2] Werner von Bloh,[2] David Archer,[3] and Guy Munhoven[4]

Received 16 October 2009; revised 1 December 2009; accepted 16 December 2009; published 21 January 2010.

[1] We are investigating the late Holocene rise in CO_2 by performing four experiments with the climate-carbon-cycle model CLIMBER2-LPJ. Apart from the deep sea sediments, important carbon cycle processes considered are carbon uptake or release by the vegetation, carbon uptake by peatlands, and CO_2 release due to shallow water sedimentation of $CaCO_3$. Ice core data of atmospheric CO_2 between 8 ka BP and preindustrial climate can only be reproduced if CO_2 outgassing due to shallow water sedimentation of $CaCO_3$ is considered. In this case the model displays an increase of nearly 20 ppmv CO_2 between 8 ka BP and present day. Model configurations that do not contain this forcing show a slight decrease in atmospheric CO_2. We can therefore explain the late Holocene rise in CO_2 by invoking natural forcing factors only, and anthropogenic forcing is not required to understand preindustrial CO_2 dynamics. **Citation:** Kleinen, T., V. Brovkin, W. von Bloh, D. Archer, and G. Munhoven (2010), Holocene carbon cycle dynamics, *Geophys. Res. Lett.*, *37*, L02705, doi:10.1029/2009GL041391.

1. Introduction

[2] During the last 8000 years (8 ka) the Antarctic ice core records [*Indermhle et al.*, 1999; *Monnin et al.*, 2004] show an increase in atmospheric CO_2 by about 20 ppmv that has so far proven difficult to explain. Possible explanations of the rise in atmospheric CO_2 include a decrease in terrestrial carbon storage [*Indermhle et al.*, 1999; *Brovkin et al.*, 2002], increases in sea surface temperatures [*Indermhle et al.*, 1999; *Joos et al.*, 2004], coral reef regrowth [*Ridgwell et al.*, 2003], and carbonate compensation [*Broecker et al.*, 1999; *Elsig et al.*, 2009]. *Broecker et al.* [1999] suggested that reconstructed reduction in the carbonate ion content of the deep sea was caused by carbonate compensation to the early Holocene forest regrowth. *Elsig et al.* [2009] estimated that 15 ppmv growth of CO_2 during the Holocene could be explained by carbonate compensation to a land biosphere uptake of 700 and 200 GtC prior and during the Holocene, respectively. The remaining CO_2 increase they attributed to coral reef growth and other mechanisms. Last but not least, *Ruddiman* [2003] proposed that the CO_2 growth was caused by human activities including CO_2 emissions due to slash and burn agriculture.

[3] The magnitude of emissions necessary to increase CO_2 by 20 ppmv (ca. 200 GtC) due to agricultural changes is not supported by estimates of landuse emissions based on population estimates prior to 1700 AD [*Joos et al.*, 2004; *Pongratz et al.*, 2009], though assuming a fixed relation between population and agricultural area may severely underestimate deforestation [*Kaplan et al.*, 2009]. In addition an increase in land surface albedo offsets the warming due to the CO_2 increase. Explaining the rise in CO_2 by a decrease in natural land carbon storage also appears unlikely because the terrestrial biosphere likely takes up more CO_2 under elevated CO_2 levels due to CO_2 fertilization [*Kaplan et al.*, 2002; *Joos et al.*, 2004]. Besides, substantial amounts of peat have accumulated in boreal wetlands during the Holocene [*Gajewski et al.*, 2001]. Therefore it is unlikely that land processes are an important contributor to CO_2 growth.

[4] In this paper we test the explanations for the Holocene rise in atmospheric CO_2 based on changes in the oceanic carbonate chemistry applied before the background of the other natural mechanisms, including vegetation dynamics, and peat accumulation.

2. Model Description and Experiments

2.1. CLIMBER2-LPJ

[5] To investigate the dynamics of Holocene CO_2 we are using CLIMBER2-LPJ in four experiments. CLIMBER2-LPJ consists of the earth system model of intermediate complexity (EMIC) CLIMBER2, coupled to the dynamic global vegetation model (DGVM) LPJ.

[6] CLIMBER2 [*Petoukhov et al.*, 2000] consists of a 2.5-dimensional statistical-dynamical atmosphere with a resolution of roughly 51° (longitude) by 10° (latitude), a zonally averaged ocean resolving three basins with a latitudinal resolution of 2.5°, and a sea ice model. CLIMBER2 also contains oceanic biogeochemistry, a model for marine biota, and a sediment model [*Archer*, 1996; *Brovkin et al.*, 2002, 2007]. Weathering rates scale to runoff from the land surface. In comparison to the earlier study by *Brovkin et al.* [2002], this setup allows to simulate carbonate compensation interactively. This is also an advantage compared to *Elsig et al.* [2009] who used a box diffusion model of the ocean uptake with a simple parameterization of carbonate compensation.

[7] To this EMIC we have coupled the DGVM LPJ [*Sitch et al.*, 2003] in order to investigate land surface processes at a significantly higher resolution of $0.5 \times 0.5°$. We also implemented carbon isotope fractionation according to *Scholze et al.* [2003]. Monthly anomalies from the climatology of the climate fields are passed to LPJ, where they are added to climate patterns based on the Climatic Research Unit CRU-TS climate data set [*New et al.*, 2000]. The carbon flux F_{AL} between atmosphere and land surface is determined from the annual change in the LPJ carbon pools, and employed in

[1]Max Planck Institute for Meteorology, Hamburg, Germany.
[2]Potsdam Institute for Climate Impact Research, Potsdam, Germany.
[3]Department of the Geophysical Sciences, University of Chicago, Chicago, Illinois, USA.
[4]LPAP, Institut d'Astrophysique et de Geophysique, Universit de Liège, Liège, Belgium.

0094-8276/10/2009GL041391$05.00

Table 1. Carbon Cycle Components Considered in the Experiments

Experiment	Vegetation	Peat	Shallow Water Sediment
AO	–	–	–
AOV	+	–	–
AOVP	+	+	–
AOVPC	+	+	+

CLIMBER2 to determine the CO_2 concentration. Biogeochemical feedbacks are thus determined by the combination of CLIMBER2 and LPJ, while biogeophysical effects are solely determined by CLIMBER2.

2.2. Model Initialization and Experiments

[8] On multi-millennial timescales the global carbon cycle is never in a complete equilibrium state due to small but persistent fluxes associated with weathering [e.g., *Munhoven*, 2002] paced by climate changes during glacial cycles. To get consistent non-equilibrium initial conditions, the simulations should start long before the early Holocene, preferably at the last glacial inception. Such a non-equilibrium approach is still beyond the computational efficiency of 2- or 3-dimensional Earth system models.

[9] To get the initial state of the carbonate system for 8 ka BP, the starting point of our experiments, we used the following two-step initialization approach:

[10] 1. The model was run with present-day equilibrium conditions with CO_2 at 280 ppmv. This simulation assumes carbonate sedimentation of 9 Tmol/a $CaCO_3$ in the deep ocean [*Brovkin et al.*, 2007] and 8.4 Tmol/a in the shelf areas. Volcanic outgassing is equal to 5.85 Tmol/a CO_2, added to the atmosphere, which is a steady state for the coupled system under present-day conditions. Using this setup, the model was run for 20,000 years.

[11] 2. Using the model state reached in step 1, we changed the orbital configuration to 8 ka BP and prescribed CO_2 to 260 ppmv. Ocean alkalinity was increased to get a carbonate sedimentation flux of 16 Tmol/a in the deep ocean and 2 Tmol/a on the shelves in order to simulate the maximum in $CaCO_3$ preservation in the deep sea before the onset of the Holocene, clearly seen in the deep sea $CaCO_3$ records [*Broecker et al.*, 1999]. Atmospheric $\delta^{13}CO_2$ was initialized at -6.4 ‰, as measured in ice cores for 8 ka BP [*Elsig et al.*, 2009]. The model was run with prescribed CO_2 for 5000 years to insure that the CO_2 distribution in the ocean stays in equilibrium, and finally for another 2000 years with CO_2 and $\delta^{13}CO_2$ exchanged interactively.

[12] This simulation setup ensures that the system is in equilibrium with the initial conditions, a mixture of weathering conditions for present-day and the Holocene $CaCO_3$ preservation maximum. LPJ was then spun up separately for 2000 years under 8 ka BP boundary conditions. Remains of the Laurentide ice sheet at 8 ka BP are neglected.

[13] In the transient simulations the climate model was driven by orbital forcing with interactive CO_2 and $\delta^{13}CO_2$. The specific forcings for the carbon cycle were:

[14] 1. Shallow water sedimentation (SWS). We prescribed a constant accumulation of carbonates on the tropical shelves of 15.5 Tmol/a, produced by coral reef growth and general accumulation on the shelves, for example through shelf bioherms. The magnitude of this forcing is in line with the estimate by *Milliman* [1993] and the model value by *Ridgwell et al.* [2003], but higher by about 5 Tmol/a than the estimate used by *Kleypas* [1997].

[15] 2. Peat accumulation forcing. *Gajewski et al.* [2001] estimate that peat deposits of about 450 GtC have accumulated over the last 21,000 years, though other estimates are substantially lower, e.g. 273 GtC estimated by *Turunen et al.* [2002]. The main reason for this difference is that Gajewski et al. use a higher estimate for the bulk density of peat. We prescribe peat accumulation with an accumulation of 100 GtC in 8 ka, about one third of the *Gajewski et al.* [2001] estimate for the last 8 ka.

[16] We performed the four model experiments shown in Table 1. In experiment AO (atmosphere-ocean), the atmospheric CO_2 concentration is determined solely by the marine carbon cycle in CLIMBER2, C fluxes F_{AL} between atmosphere and land surface are neglected. In AOV (atmosphere-ocean-vegetation) the model setup is as in AO, but fluxes F_{AL} are considered. In experiment AOVP (atmosphere-ocean-vegetation-peat), peat accumulation is considered in addition to setup AOV. Finally, in experiment AOVPC (atmosphere-ocean-vegetation-peat-corals) we are also considering CO_2 outgassing due to shallow water sedimentation of $CaCO_3$.

3. Results

[17] Figure 1a shows the atmospheric CO_2 concentration for the four experiments. In AO, CO_2 decreases from 259 ppmv at 8 ka BP to 257 ppmv at present day. This is the result of (i) a small drift in the carbonate system tuned to the present-day carbonate state, and (ii) the direct impact of the changes in orbital forcing between 8 ka BP and now, a (very small) uptake of carbon by the ocean. If biomass changes are considered in AOV there is no substantial change from AO.

[18] In AOVP, where peat uptake is considered, the decrease in CO_2 is slightly stronger: CO_2 decreases from 259 ppmv to 251 ppmv, and down to 243 ppmv in a sensitivity experiment with doubled uptake (not shown). Finally, in AOVPC, shown in black, where SWS is considered, CO_2 increases from 259 ppmv to about 278 ppmv due to the decreasing total alkalinity in response to $CaCO_3$ sedimentation. Comparing these experiments to the EPICA Dome C data [*Monnin et al.*, 2004], it becomes clear that experiment AOVPC matches measurement data closely, while no other experiment can reproduce the CO_2 increase.

[19] Figure 1b shows $\delta^{13}C$ of atmospheric CO_2 in our experiments. $\delta^{13}C$ allows conclusions with regard to the partitioning of carbon between the biosphere and other carbon pools, since plants preferentially assimilate C^{12} as opposed to C^{13}. In AO and AOV, $\delta^{13}C$ stays constant, as expected if the land biosphere doesn't take up carbon. In AOVP, $\delta^{13}C$ increases slightly, and in AOVPC it increases more strongly, the latter being a result of the increase in biomass through CO_2 fertilization. The spline fit through the EPICA Dome C record of $\delta^{13}C$ by *Elsig et al.* [2009] shows a strong increase in $\delta^{13}C$ between 11 ka BP and 6 ka BP, after which $\delta^{13}C$ slowly decreases. The error bars of the raw measurement data allow other temporal evolutions as well, including a slow increase in $\delta^{13}C$ between 8 ka BP and about 2.5 ka BP. Our results for experiment AOVPC are within the error bars of *Elsig et al.*'s [2009] $\delta^{13}C$ record, though they leave this range at about 2.5 ka BP. This divergence is no surprise, since

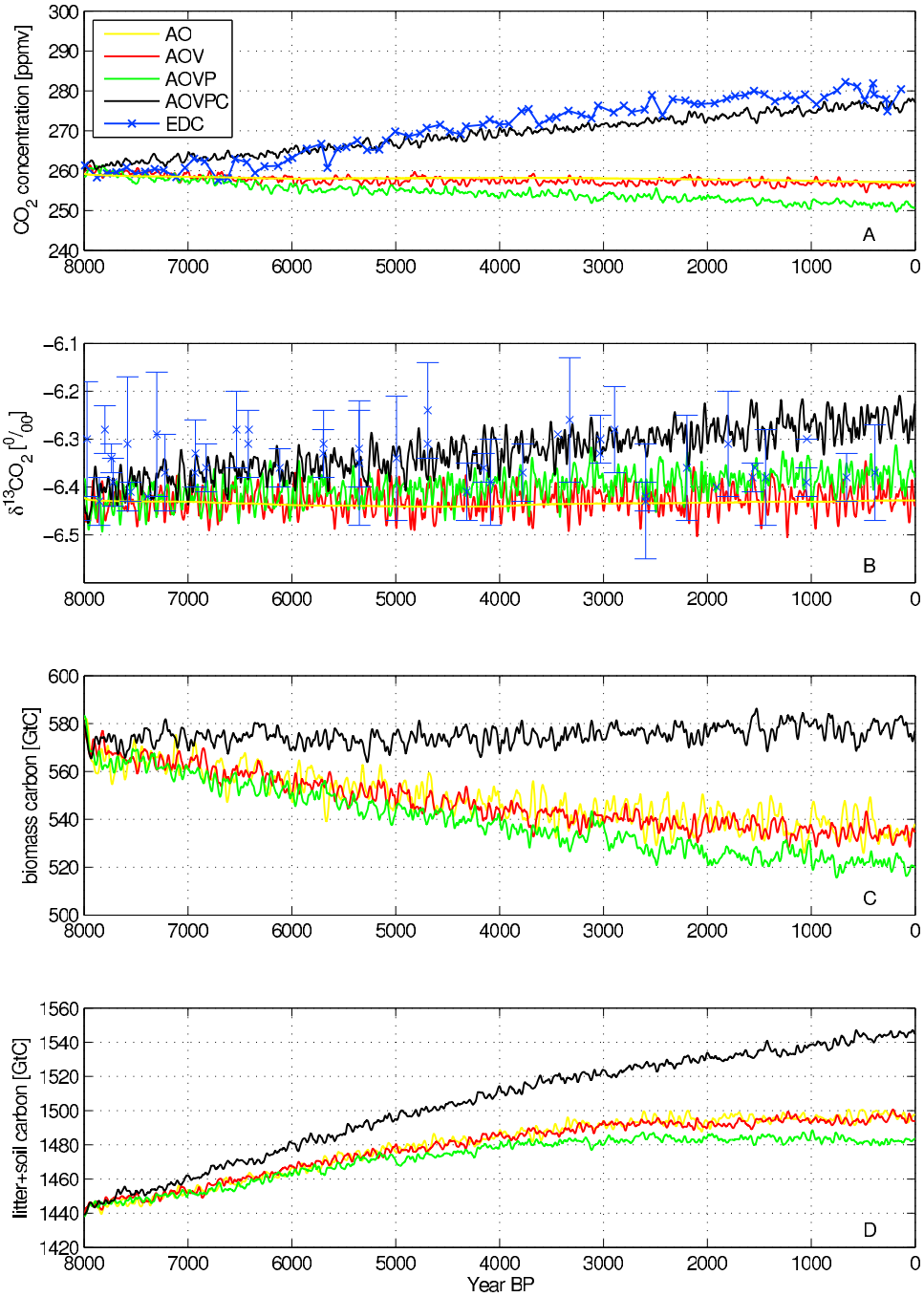

Figure 1. (a) CO_2 concentration [ppmv] and (b) $\delta^{13}CO_2$ [‰] in the model experiments, as well as data from EPICA Dome C [Monnin et al., 2004; Elsig et al., 2009]. Dynamics of (c) terrestrial biomass [GtC], and (d) litter and soil carbon [GtC] in the experiments. Model data are smoothed with a 50 year Gaussian weighted filter.

anthropogenic deforestation in not considered in our experiments, which would certainly lead to a decrease in $\delta^{13}C$.

[20] Total biomass carbon, shown in Figure 1c, decreases in all experiments but AOVPC due to the decrease in high latitude insolation and the corresponding reduction of boreal forest. In AOVPC there is a slight increase in biomass. Similarly, the litter and soil carbon (Figure 1d) increases in all experiments, with experiment AOVPC showing the largest increase, about 105 GtC. Experiments AO, AOV, and AOVP have very similar effects on modelled land vegetation: Biomass decreases, and the carbon stored in vegetation at 8 ka BP is shifted into litter and soil, making the land biosphere carbon neutral on the timescales considered.

[21] In experiment AOVPC the biomass increases due to CO_2 fertilization, and there is a corresponding increase in both litter and soil carbon, i.e., the land surface is a carbon sink. This is in disagreement with *Brovkin et al.* [2002], who show a decrease of about 90 GtC in terrestrial carbon storage in the VECODE model, mainly in subtropical areas and due to changes in monsoon, but very similar to *Kaplan et al.*

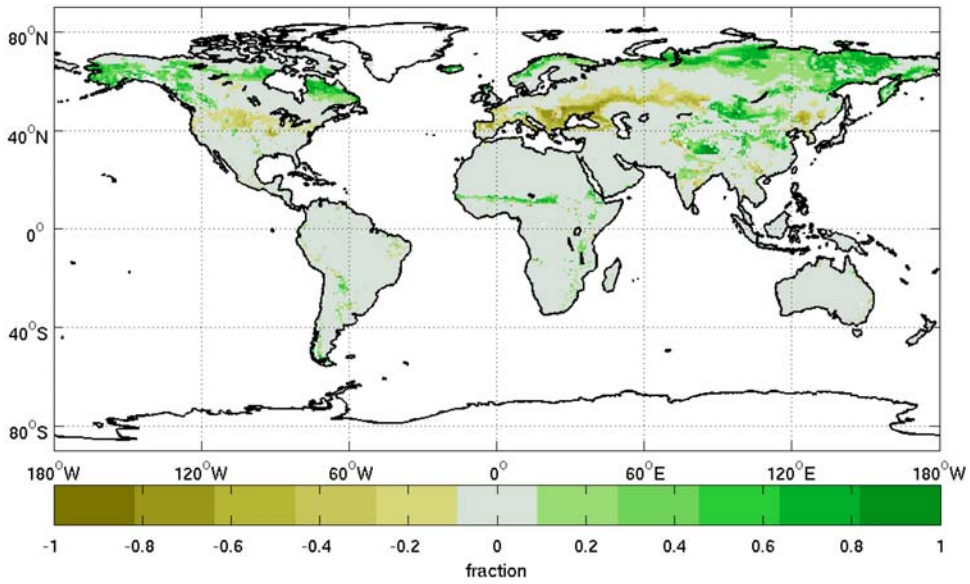

Figure 2. Tree cover change 8ka BP - present day in experiment AOV, expressed as grid cell fraction. Positive values imply higher tree cover at 8 ka BP than at present day.

[2002], who show an increase of about 100 GtC for the last 8 ka.

[22] The modelled carbonate ion concentration in the deep tropical ocean is shown in Figure S1 of the auxiliary material.[1] The CO_3^{2-} concentration declines by 11 and 18 μmol/kg in the Pacific and Atlantic basins, respectively. For the Atlantic, this is higher than the decrease by 10–12 μmol/kg reconstructed by *Broecker et al.* [1999]. This mismatch could be caused by (i) imperfectness of the coarse zonally-averaged ocean model in CLIMBER2, (ii) complications of CO_3^{2-} reconstruction [*Barker and Elderfield*, 2002], (iii) overestimation of SWS in the model experiment, or (iv) neglection of the non-equilibrium carbonate chemistry state in the initial conditions.

[23] Figure 2 shows the difference in tree cover between 8 ka BP and present day (potential natural vegetation) for experiment AOV. A northward shift of the boreal forest margin at 8 ka BP is especially pronounced in Siberia. These changes are in line with pollen-based reconstructions for western and central Siberia [*Bigelow et al.*, 2003]. In northeastern North America, the simulated northward shift in forest margin, as compared to the present, is in disagreement with pollen data [*Williams*, 2003], but this can be expected since we neglected the remains of the Laurentide ice sheet. Tree cover was reduced between 30°N and 50°N, but vegetation cover at the northern Sahel border is expanded northward for 8 ka BP, although this shift is less pronounced than in the reconstructions [*Jolly et al.*, 1998].

4. Summary and Conclusions

[24] We have performed four model experiments in order to investigate the evolution of the global carbon cycle between 8 ka BP and present day. All our experiments without shallow water sedimentation of $CaCO_3$ show a decrease in atmospheric CO_2 between 8 ka BP and present day, with the decrease in CO_2 strongest in AOVP where carbon uptake by peat is considered. *Frolking and Roulet* [2007] have assessed the climate forcing impact of peatlands by considering both CO_2 uptake and CH_4 release, and our assumption of sustained uptake of carbon by peat is in line with their conclusion of a net radiative cooling. Of our experiments, only AOVPC including SWS shows an increase in atmospheric CO_2 that is close to the reconstructions [*Monnin et al.*, 2004]. The model results suggest that the Holocene rise in CO_2 can be explained by a natural forcing, outgassing of CO_2 due to SWS. This is in line with the conclusions of *Ridgwell et al.* [2003], which we substantiate by using a coupled climate carbon cycle model.

[25] When it comes to the $\delta^{13}C$ of CO_2, our experiments are within the error bars of the latest measurement data [*Elsig et al.*, 2009], until anthropogenic deforestation can no longer be neglected at about 2.5 ka BP. While SWS in our experiment is at the higher end of estimates for the present day [*Milliman*, 1993; *Ridgwell et al.*, 2003], emissions may have been substantially higher between 8 ka BP and 4 ka BP, since sea level still rose by several meters between those times, likely leading to higher coral reef formation rates. At the same time, we are using a relatively low estimate of peat carbon uptake. While estimates vary, carbon uptake by peatlands may have been up to three times as large as considered here. This possible underestimation of the carbon uptake could easily be compensated if we assumed higher rates of $CaCO_3$ sedimentation or introduced a small contribution by carbonate compensation to the early Holocene vegetation regrowth, though the divergence in the $\delta^{13}C$ signal after 2.5 ka BP would increase at higher peat accumulation.

[26] There are limitations to our study. In particular, we did not account for a reorganization of the oceanic circulation during the transition, which should have some imprint on the early Holocene CO_2 dynamics. We also neglected land carbon changes before 8 ka BP. Accounting for them will lead to some increase of CO_2, as shown by *Elsig et al.* [2009]. However, glacial-interglacial changes in land carbon storage are not well constrained. While the Elsig et al. estimate of

[1] Auxiliary materials are available in the HTML. doi:10.1029/2009GL041391.

about 900 GtC of interglacial carbon buildup is within the envelope of uncertainties, recent studies provide evidence for enhanced glacial carbon storage in permafrost based on measurements [*Zimov et al.*, 2009] and isotopic analysis [*Ciais et al.*, 2009]. The latter study suggests less than 300 GtC transferred from ocean to land. Such a low transfer would invalidate carbonate compensation as the main explanation of the Holocene CO_2 growth. At the same time, the low carbon transfer is in line with the coral reef regrowth approach used in this paper. However, it is too early to rule out the biosphere regrowth hypothesis as observational constraints on it, as well as on the coral reef growth, are rather weak, and they both are in line with the decreasing CO_3^{2-} concentration in the deep ocean shown in sediment cores [*Broecker et al.*, 1999].

[27] **Acknowledgments.** We gratefully acknowledge support by the German Research Foundation (DFG) INTERDYNAMIK program. We would like to thank Matthias Cuntz and Marko Scholze for advice regarding the $\delta^{13}C$ implementation, and we would like to thank Fortunat Joos and Stefan Gerber for advice regarding the dependence of photosynthetically active radiation on orbital parameters. We further thank Joachim Segschneider for valuable comments on an earlier draft of this manuscript. Jed Kaplan, Pavel Tarasov, and two anonymous reviewers provided very helpful comments on the submitted manuscripts. Computations were performed at the Potsdam Institute for Climate Impact Research, and we gratefully acknowledge the use of their facilities.

References

Archer, D. (1996), A data-driven model of the global calcite lysocline, *Global Biogeochem. Cycles*, *10*, 511–526.

Barker, S., and H. Elderfield (2002), Foraminiferal calcification response to glacial-interglacial changes in atmospheric CO_2, *Science*, *297*, 833–836, doi:10.1126/science.1072815.

Bigelow, N. H., et al. (2003), Climate change and Arctic ecosystems: 1. Vegetation changes north of 55°N between the last glacial maximum, mid-Holocene, and present, *J. Geophys. Res.*, *108*(D19), 8170, doi:10.1029/2002JD002558.

Broecker, W. S., E. Clark, D. C. McCorkle, T.-H. Peng, I. Hajdas, and G. Bonani (1999), Evidence for a reduction in the carbonate ion content of the deep sea during the course of the Holocene, *Paleoceanography*, *14*, 744–752, doi:10.1029/1999PA900038.

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*, *22*, PA4202, doi:10.1029/2006PA001380.

Ciais, P., A. Tagliabue, M. Cuntz, L. Bopp, G. Hoffmann, M. Scholze, C. Prentice, J. Paris, and A. Lourantou (2009), ^{13}C and ^{18}O observations open a new window to understand the ice age carbon cycle, paper presented at the 8th International Carbon Dioxide Conference, Max-Planck-Inst. fr Biogeochm., Jena, Germany.

Elsig, J. , J. Schmitt, D. Leuenberger, R. Schneider, M. Eyer, M. Leuenberger, F. Joos, H. Fischer, and T. F. Stocker (2009), Stable isotope constraints on Holocene carbon cycle changes from an Antarctic ice core, *Nature*, *461*, 507–510, doi:10.1038/nature08393.

Frolking, S., and N. T. Roulet (2007), Holocene radiative forcing impact of northern peatland carbon accumulation and methane emissions, *Global Change Biol.*, *13*, 1079–1088, doi:10.1111/j.1365-2486.2007.01339.x.

Gajewski, K., A. Viau, M. Sawada, D. Atkinson, and S. Wilson (2001), Sphagnum peatland distribution in North America and Eurasia during the past 21,000 years, *Global Biogeochem. Cycles*, *15*, 297–310.

Indermhle, A., et al. (1999), Holocene carbon-cycle dynamics based on CO_2 trapped in ice at Taylor Dome, Antarctica, *Nature*, *398*, 121–126, doi:10.1038/18158.

Jolly, D., et al. (1998), Biome reconstruction from pollen and plant macrofossil data for Africa and the Arabian peninsula at 0 and 6000 years, *J. Biogeogr.*, *25*, 1007–1027.

Joos, F., S. Gerber, I. C. Prentice, B. L. Otto-Bliesner, and P. J. Valdes (2004), Transient simulations of Holocene atmospheric carbon dioxide and terrestrial carbon since the Last Glacial Maximum, *Global Biogeochem. Cycles*, *18*, GB2002, doi:10.1029/2003GB002156.

Kaplan, J. O., I. C. Prentice, W. Knorr, and P. J. Valdes (2002), Modeling the dynamics of terrestrial carbon storage since the Last Glacial Maximum, *Geophys. Res. Lett.*, *29*(22), 2074, doi:10.1029/2002GL015230.

Kaplan, J. O., K. M. Krumhardt, and N. Zimmermann (2009), The prehistoric and preindustrial deforestation of Europe, *Quat. Sci. Rev.*, *28*, 3016–3034, doi:10.1016/j.quascirev.2009.09.028.

Kleypas, J. A. (1997), Modeled estimates of global reef habitat and carbonate production since the Last Glacial Maximum, *Paleoceanography*, *12*, 533–545.

Milliman, J. D. (1993), roduction and accumulation of calcium carbonate in the ocean: Budget of a nonsteady state, *Global Biogeochem. Cycles*, *7*, 927–957.

Monnin, E., et al. (2004), Evidence for substantial accumulation rate variability in Antarctica during the Holocene, through synchronization of CO_2 in the Taylor Dome, Dome C and DML ice cores, *Earth Planet. Sci. Lett.*, *224*, 45–54, doi:10.1016/j.epsl.2004.05.007.

Munhoven, G. (2002), Glacial-interglacial changes of continental weathering: Estimates of the related CO_2 and HCO_3^- flux variations and their uncertainties, *Global Planet. Change*, *33*, 155–176, doi:10.1016/S0921-8181(02)00068-1.

New, M., M. Hulme, and P. Jones (2000), Representing twentieth-century space-time climate variability. Part II: Development of 1901–96 monthly grids of terrestrial surface climate, *J. Climate*, *13*(13), 2217–2238.

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.

Pongratz, J., C. H. Reick, T. Raddatz, and M. Claussen (2009), Effects of anthropogenic land cover change on the carbon cycle of the last millennium, *Global Biogeochem. Cycles*, *23*, GB4001, doi:10.1029/2009GB003488.

Ridgwell, A. J., A. J. Watson, M. A. Maslin, and J. O. Kaplan (2003), Implications of coral reef buildup for the controls on atmospheric CO_2 since the Last Glacial Maximum, *Paleoceanography*, *18*(4), 1083, doi:10.1029/2003PA000893.

Ruddiman, W. F. (2003), The anthropogenic greenhouse era began thousands of years ago, *Clim. Change*, *61*, 261–293, doi:10.1023/B:CLIM.0000004577.17928.fa.

Scholze, M., J. O. Kaplan, W. Knorr, and M. Heimann (2003), Climate and interannual variability of the atmosphere-biosphere $^{13}CO_2$ flux, *Geophys. Res. Lett.*, *30*(2), 1097, doi:10.1029/2002GL015631.

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.

Turunen, J., E. Tomppo, K. Tolonen, and A. Reinikainen (2002), Estimating carbon accumulation rates of undrained mires in Finland-Application to boreal and subarctic regions, *Holocene*, *12*, 69–80, doi:10.1191/0959683602hl522rp.

Williams, J. W. (2003), Variations in tree cover in North America since the Last Glacial Maximum, *Global and Planet. Change*, *35*, 1–23, doi:10.1016/S0921-8181(02)00088-7.

Zimov, N. S., S. A. Zimov, A. E. Zimova, G. M. Zimova, V. I. Chuprynin, and F. S. Chapin III (2009), Carbon storage in permafrost and soils of the mammoth tundra-steppe biome: Role in the global carbon budget, *Geophys. Res. Lett.*, *36*, L02502, doi:10.1029/2008GL036332.

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

V. Brovkin and T. Kleinen, Max Planck Institute for Meteorology, Bundesstr. 53, D-20146 Hamburg, Germany. (thomas.kleinen@zmaw.de)

G. Munhoven, LPAP, Institut d'Astrophysique et de Geophysique, Université de Liège, 17 Allée du 6-Août, B-4000 Liège, Belgium.

W. von Bloh, Potsdam Institute for Climate Impact Research, PO Box 60 12 03, D-14412 Potsdam, Germany.