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Positive feedback between future climate change and the carbon cycle

Pierre Friedlingstein, Laurent Bopp, Philippe Ciais,

IPSL/LSCE, CE-Saclay, 91191, Gif sur Yvette, France

Jean-Louis Dufresne, Laurent Fairhead, Hervé LeTreut,

IPSL/LMD, Université Paris 6, 75252, Paris, France

Patrick Monfray, and James Orr

IPSL/LSCE, CE-Saclay, 91191, Gif sur Yvette, France

Abstract. Future climate change due to increased atmospheric CO_2 may affect land and ocean efficiency to absorb atmospheric CO_2. Here, using climate and carbon three-dimensional models forced by a 1% per year increase in atmospheric CO_2, we show that there is a positive feedback between the climate system and the carbon cycle. Climate change reduces land and ocean uptake of CO_2, respectively by 54% and 35% at $4 \times CO_2$. This negative impact implies that for prescribed anthropogenic CO_2 emissions, the atmospheric CO_2 would be higher than the level reached if climate change does not affect the carbon cycle. We estimate the gain of this climate-carbon cycle feedback to be 10% at $2 \times CO_2$ and 20% at $4 \times CO_2$. This translates into a 15% higher mean temperature increase.

Introduction

Atmospheric CO_2 is expected to increase in the coming decades due to emissions of CO_2 by fossil fuel burning and land use changes. The rate of increase depends on anthropogenic emissions and on the capacity of the oceans and the land biosphere to take up CO_2 [*Schimel et al.*, 1995]. Current climate models predict a mean temperature increase of 1 to 4.5°C compared to the present for a doubling of atmospheric CO_2 [*Kattenberg et al.*, 1996]. Recent carbon cycle studies suggest that such climate change may reduce the uptake of CO_2 by the ocean [*Maier-Reimer et al.*, 1996; *Sarmiento et al.*, 1998; *Matear and Hirst*, 1999] or the land biosphere [*Cao and Woodward*, 1998; *Meyer et al.*, 1999; *Cramer et al.*, 2000]. It is thus necessary to account for the climate impact on the carbon cycle when translating anthropogenic emissions into CO_2 concentrations.

Method

In this study, we used a model structure composed of a coupled ocean-atmosphere general circulation model (OAGCM), and models of land and ocean components of the carbon cycle, the carbon cycle models being forced by the climate fields of the OAGCM. Two climate simulations have been run with the OAGCM : the control run where the CO_2 is held constant at 350 ppmv and the transient climate run where the CO_2 increases at a rate of 1% per year from 350 ppmv up to 1400 ppmv (Figure 1a). We then performed two carbon simulations. In the "constant climate" simulation the carbon models are forced by a CO_2 increase of 1% per year and the control climate from the OAGCM. In the "climate change" simulation, the carbon models are forced by the same 1%/yr CO_2 increase as well as the climate from the transient climate run.

In this experiment, atmospheric CO_2 and monthly averaged climate fields from the IPSL OAGCM [*Braconnot et al.*, 2000] are used to drive both terrestrial and oceanic carbon cycle models. These two models allow one to translate anthropogenic CO_2 emissions into atmospheric CO_2 concentration trajectories and vice-versa. The terrestrial carbon model (SLAVE) [*Friedlingstein et al.*, 1995; *Ciais et al.*, 1999] is driven by surface air temperature, precipitation, and solar radiation, and calculates net primary productivity (NPP) following a light use efficiency formulation [*Field et al.*, 1995] that is a function of temperature and water stress. NPP increases with CO_2 under a Michaelis-Menten beta factor formulation [*Gifford*, 1992], which has a global value of 0.5, in the upper range of experimental data [*DeLucia et al.*, 1999], although, nitrogen limitation and deposition as well as vegetation dynamics and land use changes are ignored in this study. The ocean carbon model (IPSL-OCCM1) [*Aumont et al.*, 1999; *Le Quéré et al.*, 1999], based on the HAMOCC3 biogeochemical scheme [*Maier-Reimer*, 1993] is driven by monthly mean global fields of oceanic circulation, temperature, salinity, and surface fields of winds, sea ice and water fluxes all issued from the OAGCM. Both land and ocean carbon models have been applied successfully to study seasonal, interannual and decadal characteristics of the carbon cycle over the historical period [*Friedlingstein et al.*, 1995; *Ciais et al.*, 1999; *Aumont et al.*, 1999; *Le Quéré et al.*, 1999].

Table 1. Changes in cumulated carbon budget at $2{\times}CO_2$ and $4{\times}CO_2$

	$2{\times}CO_2$		$4{\times}CO_2$	
	Constant Clim.	Clim. Change	Constant Clim.	Clim. Change
Ocean uptake (GtC)	347	312 (−10%)	1002	800 (−20%)
Land uptake (GtC)	403	310 (−23%)	1195	808 (−32%)
Atmospheric Increase (GtC)	742	742	2226	2226
Anthropogenic emission (GtC)	1492	1364 (−8.5%)	4423	3834 (−13%)

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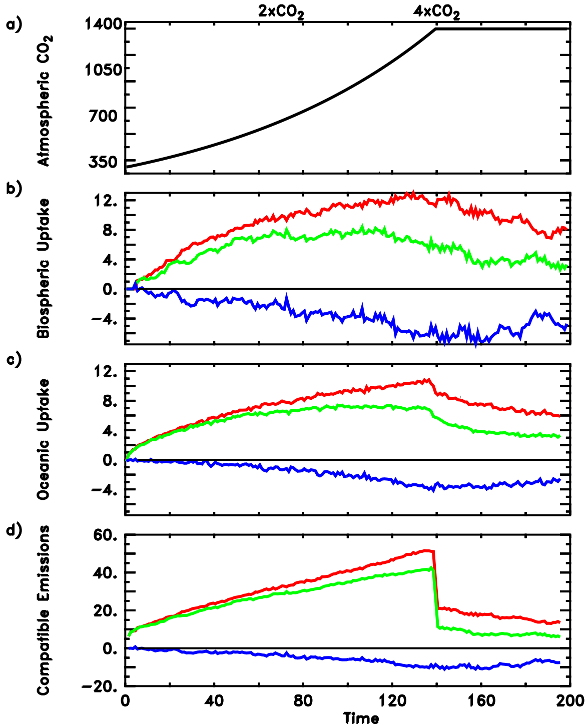

Figure 1. Carbon budget. a) Atmospheric CO_2 scenario used as a forcing for the climate model (in ppmv) (*12*). b) Simulated annual biospheric CO_2 uptake (GtC/yr) for the constant climate simulation (red line), the climate change simulation (green line) and the difference between the two simulations, showing the climate change impact on reduction biospheric carbon uptake (blue line). c) same as b), but for the ocean. d) Annual rate of compatible anthropogenic CO_2 emissions calculated as the sum of atmospheric CO_2 growth rate and land plus ocean carbon uptakes (GtC/yr). Lines colors follows the same convention as in b).

Climate Impact on Land Uptake

In the constant climate experiment, increasing CO_2 stimulates terrestrial NPP from 70 to 110 GtC/yr at $2 \times CO_2$, and to 150 GtC/yr at $4 \times CO_2$. These results fall within the range of previous model [*Cao and Woodward*, 1998; *Meyer et al.*, 1999; *Cramer et al.*, 2000]. The residence time of carbon in living and dead biomass induces a transient disequilibrium between NPP and the release due to oxidation of decaying material. A net biospheric uptake (NEP) grows as long as atmospheric CO_2 increases, reaching 9 GtC/yr at $2 \times CO_2$ and 12 GtC/yr at $4 \times CO_2$ (Figure 1b). When CO_2 stabilizes, so does the NPP, and the biosphere reaches a new equilibrium state. The climate change experiment, shows a much smaller NEP than the constant climate run (Figure 1b). Ten years before reaching $2 \times CO_2$, NEP saturates at around 7 GtC/yr, and starts to decrease after 120 years despite increasing atmospheric CO_2. When CO_2 reaches $4 \times CO_2$, NEP only amounts to 5.5 GtC/yr, less than half of the value found at the same CO_2 level in the constant climate run. The cumulative land uptake in the climate change run is 310 GtC at $2 \times CO_2$ and 808 GtC at $4 \times CO_2$, that is respectively 23% and 32% lower than in the constant climate simulation (Table 1).

The strong reduction of NEP induced by the climate change is mainly located in the subtropics (especially South America) and caused by increase in soil aridity, due to a larger increase in evaporative demand than in precipitation (Figure 2). Qualitatively similar findings were found previously [*Cao and Woodward*, 1998; *Cramer et al.*, 2000].

Climate Impact on Ocean Uptake

For the constant climate run, rising atmospheric CO_2 also increases the oceanic uptake. At $2 \times CO_2$, the ocean carbon sink reaches 7.5 GtC/yr and 10.5 GtC/yr at $4 \times CO_2$ (Figure 1c). After CO_2 stabilizes, the ocean uptake decreases as the ocean carbon tends toward a new equilibrium state. As for the land uptake, the oceanic uptake is always lower in the climate change simulation than under constant climate. After 80 years, oceanic uptake saturates around 7 GtC/yr, and shows a slight decrease during the last ten years of increasing atmospheric CO_2. At $4 \times CO_2$, the oceanic uptakes amounts to 5.7 GtC/yr, which is 35% lower than in the constant climate run. When cumulated, the climate induced decrease of oceanic uptake is 10% at $2 \times CO_2$, and 20% at $4 \times CO_2$ (Table 1). The effect of the global warming scenario in reducing oceanic CO_2 uptake is, at $2 \times CO_2$ of the same order as that was previously found [*Maier-Reimer et al.*, 1996; *Sarmiento et al.*, 1998; *Matear and Hirst*, 1999]. As shown on Figure 3, the reduction in the oceanic uptake of carbon, as discussed in earlier studies, results from the combination of three effects: impact of increased sea-surface temperature on CO_2 solubility, impact of reduced vertical mixing on CO_2 transport from the surface to the deep ocean and impact of changes in the biogeochemical cycle of CO_2. The combination of those three climatic feedbacks lead to a reduced oceanic uptake of CO_2, principally located at high latitudes, and for its main part in the Southern Ocean. However, this effect might be over-evaluated in our experiments due to abnormally strong oceanic convection in the Southern Ocean.

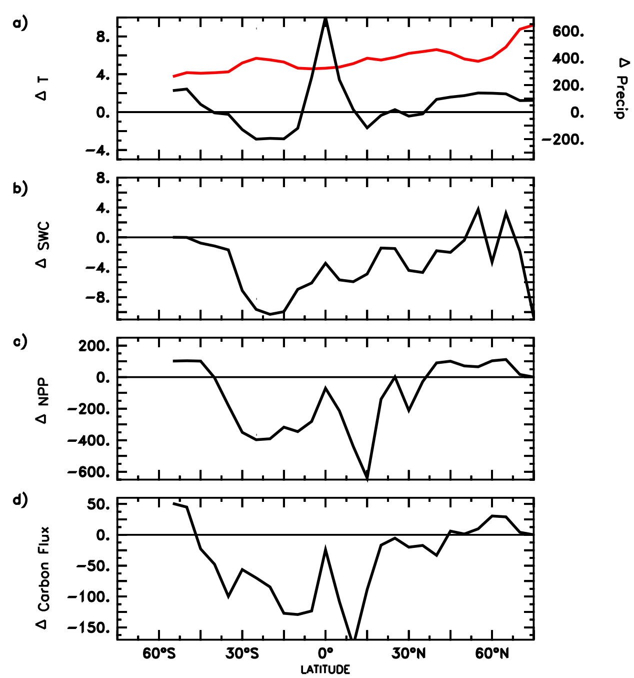

Figure 2. Zonal mean difference between the climate change and the constant climate simulations at the time of $4 \times CO_2$ of a) annual surface land temperature (°C) (red line) and precipitation (mm/yr) (blue line), b) soil water content (mm), c) Net Primary Productivity (gC/m^2/yr), and d) net carbon uptake (gC/m^2/yr).

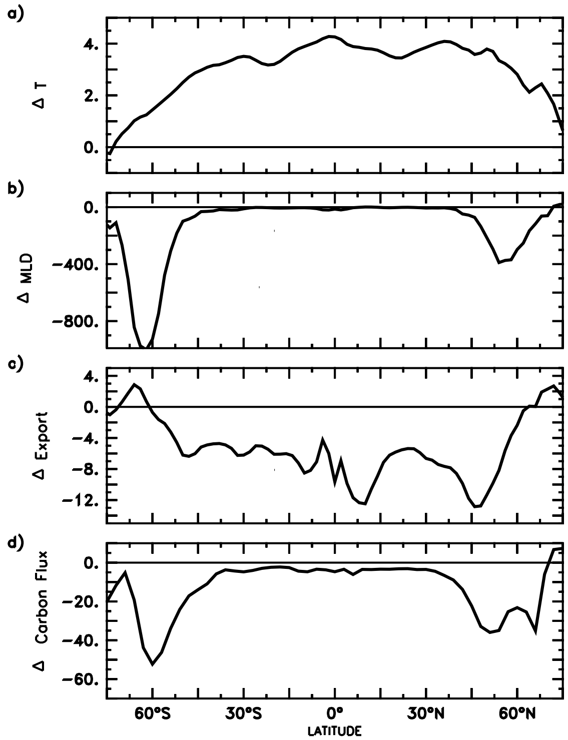

Figure 3. Zonal mean difference between the climate change and the constant climate simulations at the time of $4xCO_2$ of a) sea surface temperature (°C), b) depth of the mixed layer (m), c) export production ($gC/m^2/yr$), and d) net carbon uptake ($gC/m^2/yr$).

In the following, we provide the first estimate of the magnitude of this positive feedback. Using a classical approach [*Hansen et al.*, 1984], we define the gain of the climate system carbon cycle feedback, g, as $\partial^* T/\partial C \times \partial^* C/\partial T$ where the first term represents the overall physical sensitivity of temperature to atmospheric CO_2, and the second term represents the overall sensitivity of atmospheric CO_2 to temperature. In our climate change simulation, the sensitivity of temperature to CO_2 gradually decreases from 0.007 K/ppmv at $2 \times CO_2$ to 0.003 K/ppmv at $4 \times CO_2$ (Figure 4a). The CO_2 sensitivity to temperature can be inferred from Figure 1, showing the impact of climate change on the carbon fluxes, and from the calculated airborne fraction. The CO_2 sensitivity to temperature increases strongly from 20 ppmv/K at $2 \times CO_2$ to 60 ppmv/K at $4 \times CO_2$ (Figure 4b). The gain, g, defined above, amounts to 0.11 at $2 \times CO_2$ and reaches 0.19 at $4 \times CO_2$ (Figure 4c). The net feedback, f, which is the global warming amplification, defined as $1/(1-g)$, reaches 1.12 and 1.23 at $2 \times CO_2$ and $4 \times CO_2$ respectively (Figure 4d). Assuming that future emissions follow a trajectory compatible with today's climate (Figure 1d, constant climate simulation), one can approximate the CO_2 levels and the climate change that would occur in a coupled climate-carbon cycle configuration. This analytical calculation gives a 5.2°C warming at a CO_2 level of 1560 ppmv after 140 years instead of a 4.6°C warming at 1400 ppmv, as given by the uncoupled simulation.

Impact on Derived Emissions

Our two estimates of both terrestrial and oceanic carbon uptakes allow us to determine the compatible anthropogenic emissions with and without accounting for the climate change. In the constant climate simulation, in order to sustain a 1%/yr increase in atmospheric CO_2, the compatible emissions have to peak at 50 GtC/yr at the time of $4 \times CO_2$, whereas they would be lowered by 10 GtC/yr in the climate change run (Figure 1d). When cumulated, the compatible emissions are respectively reduced by 8% and 13% at $2 \times CO_2$ and $4 \times CO_2$ when climate change is accounted for (Table 1). Thus, to achieve a given atmospheric CO_2 trajectory any economic CO_2 emission scenario needs to prescribe lower emissions if the climate impact on the carbon cycle is accounted for. Furthermore, the response of the carbon cycle to the warming being non-linear, reductions in emissions will then have to be increasingly stronger with time.

Climate System – Carbon Cycle Feedback

In a more consistent study where the carbon cycle is forced by anthropogenic emissions, as in the real world, our results would translate into a faster atmospheric CO_2 buildup as land and ocean efficiencies to sequester carbon decrease with time. That in turn would feed back in a more rapid climate change and may have further adverse impacts on terrestrial and oceanic processes and on the CO_2 concentration.

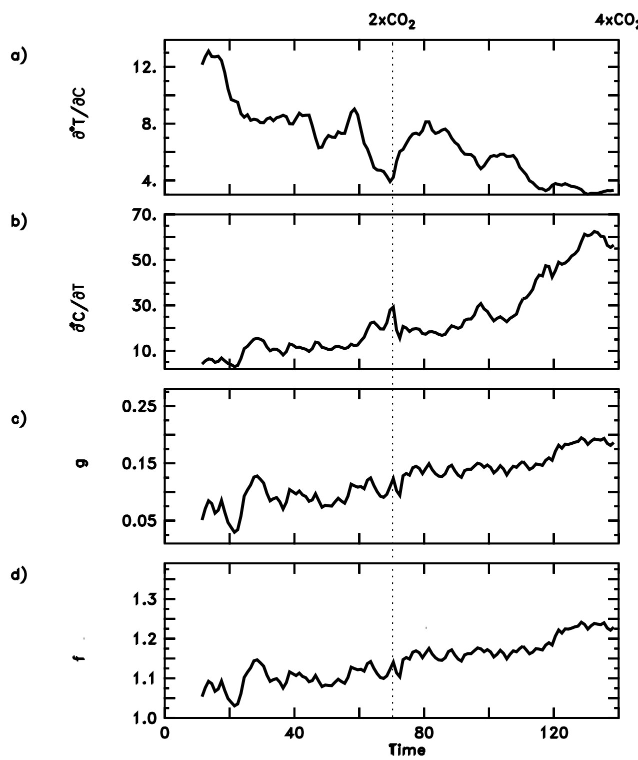

Figure 4. Time evolution of a) $\partial^* T/\partial C$, the overall sensitivity of surface temperature to the atmospheric CO_2 (10^3 K/ppmv), b) $\partial^* C/\partial T$, the overall sensitivity of atmospheric CO_2 to surface temperature (ppmv/K), c) g, the gain of the climate system-carbon cycle feedback calculated as $g = \partial^* T/\partial C \times \partial^* C/\partial T$, and d) f, the global warming amplification calculated as $f = 1/(1-g)$.

Conclusions

Our results suggest that the future climate change impact on the carbon cycle can be large, with a risk of seeing both ocean and biospheric capacity to absorb anthropogenic CO_2 significantly reduced as the Earth warms up, leaving larger CO_2 fraction in the atmosphere and therefore enhancing the climate change. In order to further explore these effects, it should be given high priority to develop comprehensive models where physical climate system and carbon cycle are explicitly coupled.

This study is a first attempt to quantify the climate–carbon feedback under elevated CO_2. To help reduce uncertainties, and to identify the key processes controlling CO_2 and climate requires a better understanding of the observed historical trends. In future scenarios, one should also specifically account for changes in non–CO_2 greenhouse gases, in future land use and land cover, in vegetation-climate feedbacks controlled by stomatal conductance and canopy development, as well as for alterations in land and ocean ecosystem distribution, and in the cycling of nutrients. In addition, non-linear changes in the ocean-atmosphere dynamics [*Manabe et al.*, 1992], could affect the magnitude of the feedback we have calculated here.

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P. Friedlingstein, L. Bopp, P. Ciais, P. Monfray, and J. Orr, IPSL/LSCE, CE-Saclay, 91191, Gif sur Yvette, France. (email: pierre@lsce.saclay.cea.fr; bopp@lsce.saclay.cea.fr; ciais@cea.fr; monfray@cea.fr; orr@cea.fr)

J.-L. Dufresne, L. Fairhead, and H. LeTreut, IPSL/LMD, Université Paris 6, 75252, Paris, France. (email: Jean-Louis.Dufresne @lmd.jussieu.fr; Laurent.Fairhead@lmd.jussieu.fr; Herve.Letreut @lmd.jussieu.fr)

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