On the magnitude of positive feedback between future climate change and the carbon cycle

J.-L. Dufresne

IPSL/LMD, Paris, France

P. Friedlingstein, M. Berthelot, L. Bopp, P. Ciais

IPSL/LSCE , Gif sur Yvette, France

L. Fairhead, H. Le Treut

IPSL/LMD, Paris, France

P. Monfray

IPSL/LSCE , Gif sur Yvette, France

Abstract. We use an ocean-atmosphere general circulation model coupled to land and ocean carbon models to simulate the evolution of climate and atmospheric CO_2 from 1860 to 2100. Our model reproduces the observed global mean temperature changes and the growth rate of atmospheric CO_2 for the period 1860-2000. For the future, we simulate that the climate change due to CO_2 increase will reduce the land carbon uptake, leaving a larger fraction of anthropogenic CO_2 in the atmosphere. By 2100, we estimate that atmospheric CO_2 will be 18% higher due to the climate change impact on the carbon cycle. Such a positive feedback has also been found by [*Cox et al.*, 2000]. However, the amplitude of our feedback is three times smaller than the one they simulated. We show that the partitioning between carbon stored in the living biomass or in the soil, and their respective sensitivity to increased CO_2 and climate change largely explain this discrepancy.

Introduction

Atmospheric CO_2 has increased by 80 ppmv over the last two centuries as a result of fossil fuel burning and land use changes [*Schimel et al.*, 1995]. These anthropogenic emissions are expected to continue in the coming decades. The corresponding atmospheric CO_2 concentration, needed to estimate future climate change, is not straightforward to predict since both the land and the ocean play a role in controlling atmospheric CO_2. Rising atmospheric CO_2 is known to increase plant photosynthesis [*DeLucia et al.*, 1999] and carbon dissolution in seawater [*Oeschger et al.*, 1975]. On the other hand, future climate change is believed to reduce the ocean carbon uptake [*Sarmiento et al.*, 1998; *Joos et al.*, 1999] and the terrestrial carbon uptake [*Cramer et al.*, 2001]. These reductions may constitute a positive feedback which has been estimated in a previous analysis based on off-line simulations [*Friedlingstein et al.*, 2001]. Recently, fully coupled climate-carbon simulations were performed by [*Cox et al.*, 2000] using the Hadley Center HadCM3 climate model coupled to a carbon cycle model. They found a very large negative impact of climate change on land carbon cycle with a decline of tropical forest and a widespread climate-driven loss of soil carbon leading to large CO_2 losses to the atmosphere. In this paper we present results obtained with our climate-carbon coupled model and we compare our results with [*Cox et al.*, 2000], highlighting the main reasons for discrepancies.

Models and Runs Description

Our climate model is a coupled ocean-atmosphere general circulation mode, the IPSL-CM2 model [*Khodri et al.*, 2001]. The carbon model is composed of the SLAVE code [*Friedlingstein et al.*, 1995] for the terrestrial part and of the IPSL-OCCM1 code [*Aumont et al.*, 1999] for the ocean part. Simulated monthly mean climate fields and annual mean atmospheric CO_2 concentration drive both carbon models, the annual change in CO_2 concentration being calculated as the balance between prescribed emissions and the land and ocean sinks.

Starting from an initial state where the coupled climate carbon model is near equilibrium[1], two 240 year-long simulations have been performed without any restoring term or flux correction. The first one is a control simulation without any anthropogenic CO_2 sources. The second is a scenario simulation where CO_2 emissions are prescribed, using historical emissions from fossil and land use change from 1860 to 1990 [*Andres et al.*, 1996], and the IPCC SRES98-A2 emission scenario from 1990 to 2100 [*Nakicenovic et al.*, 2000]. CO_2 sources are the only change between the two runs. The control run results are realistic and display no significant drift[1] in both global mean surface temperature and atmospheric CO_2 (Fig. 1).

Recent Historical Period and Future Period

Over the period 1860-2000, we simulate a rise in global mean surface temperature of 0.6°C (Fig. 1), consistent with observations [*Jones*, 1994]. The simulated atmospheric CO_2 concentration matches measurements [*Etheridge et al.*, 1996] within a few ppmv (Fig. 1). Over the 1980s, the mean land and ocean uptake amounts to 1.9 and 2.1 $GtCy^{-1}$ respectively, consistent with recent estimates [*Keeling et al.*, 1996; *Battle et al.*, 2000]. Also, CO_2 interannual variability is realistic in term of amplitude.

In the scenario run we obtain, by 2100, an atmospheric CO_2 increase of 484 ppmv (from 286 to 770 ppmv), a global temperature increase of 3°C (4.4°C over the continents) and a small global mean precipitation increase (4%). The oceanic circulation shows a small but significant reduction of the thermohaline circulation and of the deep convection at high latitudes.

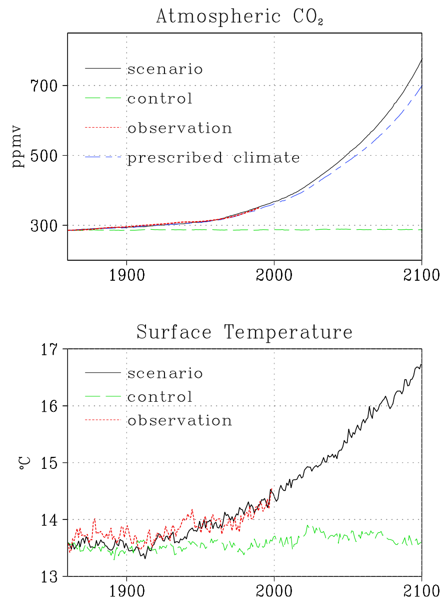

Figure 1. Atmospheric CO_2 and global mean surface temperature time series for the control run for the scenario run and comparison with observation. The atmospheric CO_2 for the prescribed climate run is also represented.

Carbon Climate Feedbacks

To quantify the climate carbon feedback, we perform an additional carbon cycle simulation called "prescribed climate", where the CO_2 emissions are the same as in the "scenario" run but where the associated climate change is not considered. The biospheric and oceanic carbon models are therefore driven by the climate of the control run. The prescribed climate simulation shows a lower atmospheric CO_2 increase than does the scenario simulation (Fig. 1). By 2100, CO_2 reaches only 695 ppmv. That means that interactions between climate and carbon cycle increase by 18% the accumulation of CO_2 in the atmosphere.

This results confirm the findings of [*Cox et al.*, 2000] and we both attribute this positive feedback to the terrestrial biosphere. However magnitude of their climate impact is approximately 3 times larger than ours. While we simulate a 75 ppmv additional atmospheric CO_2 increase, [*Cox et al.*, 2000] simulate a 250 ppmv increase because of climate change impact on the biosphere. In our simulations, the cumulated land uptake over the 1860-2100 period amounts to 680 GtC in the prescribed climate run and is reduced to 480 GtC in the scenario run. *Cox et al.*, [2000] simulate a 630 GtC uptake in their prescribe climate run (called the offline experiment in their paper) but a 100 GtC source in their scenario run (called the coupled run in their paper).

Regarding the ocean uptake, both studies agree that despite a stratification and a warming of its surface, the ocean uptake slightly increase under the climate change. In our study, cumulated ocean uptake amounts to 670 GtC in the prescribed climate run, to be compared to 700 in the scenario run (Hadley corresponding numbers being 370 and 490 respectively). This seems to be in contradiction with previous ocean only model studies [*Sarmiento et al.*, 1998; *Joos et al.*, 1999; *Bopp et al.*, 2001] which all simulated that climate change reduces ocean carbon uptake. Additional simulations show that the climate-induced reduction of biospheric uptake induces a larger atmospheric CO_2 rise, leading to a larger geochemical air-sea exchange. That latter compensates the reduction of ocean carbon uptake due to climate change alone. In other words, the reduced land uptake, through enhanced atmospheric CO_2 leads to a increased ocean uptake, a feedback obviously missed in ocean only studies. Note that the IPSL ocean uptake, despite a lower atmospheric CO_2, is much larger than the Hadley ocean uptake.

Discussion

On the continents, a combination of several mechanisms explain the large difference between the Hadley Center and the IPSL results. One obvious difference between our study and the one from the Hadley Center is that this latter uses a terrestrial model that accounts for vegetation dynamics, a process we did not include in our simulations. However, the following analysis clearly shows that this can not be the only reason.

Figure 2 shows the allocation of carbon to vegetation and soil for the runs of both Hadley and IPSL. The major difference between the two study lies in the CO_2 and climate relative impacts on vegetation and soil carbon.

First, we predict that increased CO_2 alone induces a larger increase of carbon in vegetation (400 GtC) than in soil (200 GtC), whereas the [*Cox et al.*, 2000] study shows the partitioning tendency to be in the opposite sense (200 GtC increase in vegetation and 400 GtC increase in soils) (Fig. 2). We compared these results to the ones from [*Cramer et al.*, 2001], who intercompares 6 different dynamic global vegetation models (DGVMs) under CO_2 and climate change scenario. Out of these 6 DGVMs, 3 simulate a larger storage in vegetation than in soil as we do, 2 simulate similar storage in vegetation and soil, while only one (the model of *Cox et al.*, [2000]) simulates the opposite, a larger storage in the soil than in the vegetation (Table 1).

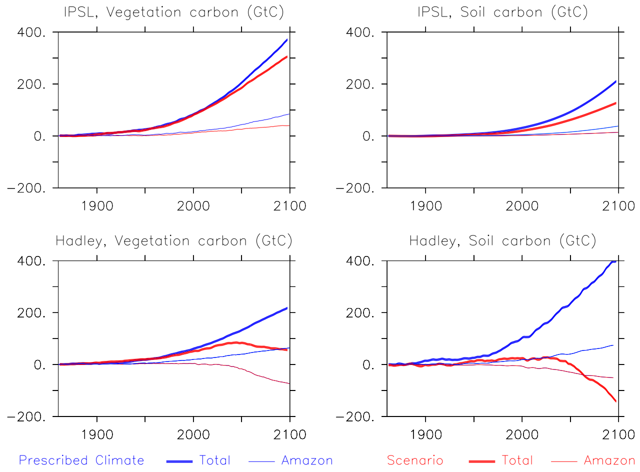

Figure 2. Partitioning of terrestrial carbon pools changes between living vegetation (top left) and soil (top right). Blue line shows the carbon changes in the prescribed climate run for total lands (thick line) and Amazon region only (thin line), red lines are for the scenario run. The bottom two panels show the same quantities from the [*Cox et al.*, 2000] study.

Table 1. Carbon allocated to vegetation (Δ Veg), carbon allocated to soil (Δ Soil) and allocation ratio for the models from the [*Cramer et al.*, 2001] DGVM intercomparison study, the [*Cox et al.*, 2000] coupled simulation and this present work. All numbers are taken for the simulation with CO_2 increase alone, at the time where atmospheric CO_2 reaches 700 ppmv.

	Δ Veg (GtC)	Δ Soil (GtC)	Δ Veg/Δ Soil
HYBRID	410	75	5.47
IBIS	460	180	2.55
SDGVM	260	120	2.17
VECODE	180	160	1.13
LPJ	250	270	0.93
TRIFFID	400	520	0.77
Cox et al.	220	410	0.54
This study	380	220	1.73

Second, the climate impact on the terrestrial reservoirs is also quite different in the two model runs. Spatially, our carbon decrease is mainly located in the tropics, and affects equally the vegetation and the soil, whereas in [*Cox et al.*, 2000], vegetation carbon decrease is primarily tropical, but soil carbon release, which is responsible for their large positive feedback, is essentially occuring in the extra-tropics (Fig. 2). The contribution of the Amazon forest dieback to the positive feedback is secondary.

When focussing on the processes responsible for the reduced land uptake in both studies, we also see different responses. In the IPSL scenario run, the warming is associated to a soil drying in the tropics. The drying leads to a severe reduction of net primary production (NPP) (-25% in the Amazon basin), as well as a reduction of soil respiration rate (SRR). The warming tends to increase SRR, but the drying effect dominates and SRR decreases in the tropics in our coupled simulation. The Hadley run also simulates an NPP reduction in the Amazon basin, but simulates a very large increase of SSR because of higher temperature. We note that both models parameterize the soil respiration response to temperature with a Q_{10} formulation, the value of the Q_{10} parameter being set to 2 in the two models. However, [*Cox et al.*, 2000] simulate a much larger negative climate impact on soil carbon (-550 GtC) than we do (-100 GtC) (Fig. 2). This is in part due to the large soil pool they build through CO_2 fertilization, but also due to the worldwide SRR increase consequent to the warming simulated by their climate model.

Finally, we are also aware that we do not account for the actual change of land cover due to deforestation. Therefore, our simulation with a fixed land cover is likely to overestimate CO_2 fertilization in regions where deforestation may occur in the future. In order to evaluate this bias, we performed an additional offline simulation with an imposed 1%/yr conversion of tropical forest area to savannas. This reduces the biospheric and enhances the positive feedback by an other 50 ppmv by 2100. The Hadley simulation does not account for direct land cover changes either. However, they simulate the climate impact on land cover with the dieback of the Amazon forest starting around 2050. Although this may seems realistic, it introduces an inconsistency insofar as their input emission scenario already accounts for tropical deforestation, and they thus account twice for the associated CO_2 fluxes.

Conclusion

In summary, our results confirm the [*Cox et al.*, 2000] study that there is a positive feedback between climate and carbon cycle, due to the climate impact on the terrestrial biosphere. However, the effects of climate change on atmospheric CO_2 and on the carbon cycle is subtle and still poorly constrained by manipulative experiments. Better representation of future carbon cycle and its control on the climate system requires a clear understanding at the regional scale of the relative sensitivities of the land and ocean carbon cycle to atmospheric CO_2 and climate change. As highlighted by the comparison with the study of [*Cox et al.*, 2000], the question of whether the additional storage of carbon occurs in the vegetation or in the soil seems to be crucial, as these two compartments have drastically different sensitivities to climate change. Also, the response of soil respiration to warming and drying needs to be better constrained in terrestrial models. Finally, a clear methodology needs to be pursued in order to account for both the direct (through deforestation) and indirect (through climate change) impacts of human on the vegetation in vegetation models avoiding model inconsistencies.

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J.-L. Dufresne, L. Fairhead and H. Le Treut, LMD/IPSL, Université Paris 6, boite 99, F-75252 Paris cedex 05, France. (e-mail: dufresne@lmd.jussieu.fr ; fairhead@lmd.jussieu.fr ; letreut@lmd.jussieu.fr)

M. Berthelot, L. Bopp, P. Ciais, P. Friedlingstein and P. Monfray, LSCE/IPSL, Bat. 709, CE L'Orme des Merisiers, F-91191 Gif sur Yvette, France. (e-mail: mberthe@lsce.saclay.cea.fr ; bopp@lsce.saclay.cea.fr; ciais@lsce.saclay.cea.fr; pierre@lsce.saclay.cea.fr ; monfray@lsce.saclay.cea.fr)

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