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Notes du Pôle de Modélisation

Positive feedback of the carbon cycle on future climate change

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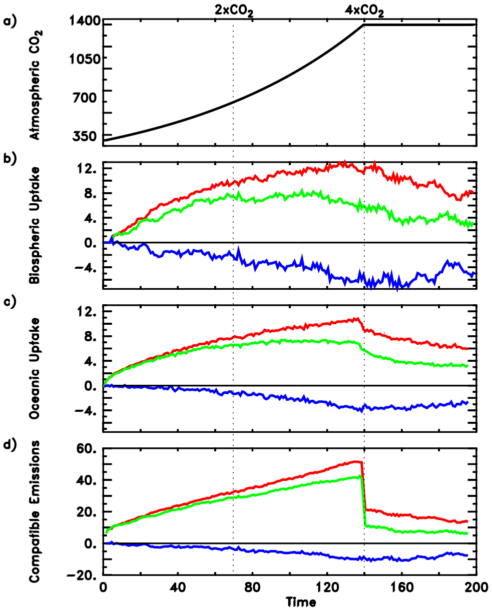

Octobre 2000 **Note n° 19**

I P S L

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Future climate change may affect land and ocean efficiency to absorb atmospheric CO_2. Here, using a Ocean-Atmosphere General Circulation Model and an Land and Ocean carbon three-dimensional model forced by a 1% per year increase in atmospheric CO_2, we model that climate change reduces land and ocean uptake of CO_2. There is therefore a positive feedback between the climate system and the carbon cycle. The gain of this positive feedback is estimated to be 0.1 at $2xCO_2$ and 0.2 at $4xCO_2$, which implies that for prescribed anthropogenic CO_2 emissions, the increase in global mean temperature could be up to 15% higher, if the climate change impact on the carbon cycle is accounted for.

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Abstract

Future climate change may affect land and ocean efficiency to absorb atmospheric CO_2. Here, using a Ocean-Atmosphere General Circulation Model and an Land and Ocean carbon three-dimensional model forced by a 1% per year increase in atmospheric CO_2, we model that climate change reduces land and ocean uptake of CO_2. There is therefore a positive feedback between the climate system and the carbon cycle. The gain of this positive feedback is estimated to be 0.1 at $2xCO_2$ and 0.2 at $4xCO_2$, which implies that for prescribed anthropogenic CO_2 emissions, the increase in global mean temperature could be up to 15% higher, if the climate change impact on the carbon cycle is accounted for.

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 terrestrial biosphere to take up CO_2 [*Schimel et al.*, 1995; *Keeling et al.*, 1996]. 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 and Le Quéré*, 1996; *Sarmiento et al.*, 1998; *Matear and Hirst*, 1999] or the land biosphere [*Joos et al.*, 1999; *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. This has not been done so far [*Kattenberg et al.*, 1996] .

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 CO_2 is held constant at 350 ppmv and the transient climate run where CO_2 increases at a rate of 1% per year [*Meehl et al.*, 1999] 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. Doing so allows to separate the climate impact from the geochemical impact of increasing CO_2 on the carbon cycle, to estimate the emissions compatible with a given CO_2 scenario under the climate change, and to calculate the sign and the strength of the feedback between the climate system and the carbon cycle.

In practice, 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. 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 [*Monteith*, 1973; *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 [*Wullschleger et al.*, 1995; *DeLucia et al.*, 1999]. 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].

The OAGCM control simulation is at quasi-equilibrium and produces a realistic geographic pattern of sea surface temperature [*Barthelet et al.*, 1998], without any flux correction being necessary. Its main deficiency in simulating today's climate is an incorrect fresh water flux at southern high latitudes, inducing an excessive convection in these regions, which hinders the geographic extent of sea ice. The results from the climate change simulation, which will be included in the Third Assessment Report of IPCC, are within the range of the current AOGCMs [*Kattenberg et al.*, 1996]. At $2xCO_2$ (year 70 of the simulation), global surface air temperature increases by 2 °C with a higher increase over the continents and in the high latitudes. Precipitation streghtens in the tropics and high latitudes but weakens in the subtropical region (15° to 40° N and S). South America experiences a general decrease in precipitation. At $4xCO_2$ (year 140 of the simulation), global warming reaches 4.6 °C.

Climate Impact on Land Uptake

In the constant climate experiment, increasing CO_2 stimulates terrestrial NPP from 70 GtC/yr to 110 GtC/yr at $2xCO_2$, and to 150 GtC/yr at $4xCO_2$. These results fall within the range of previous model estimates [*Cao and Woodward*, 1998; *Meyer et al.*, 1999; *Cramer et al.*, 2000], or experimental studies [*Wullschleger et al.*, 1995; *DeLucia et al.*, 1999] . The residence time of carbon in living and dead biomass induces a transient disequilibrium between NPP and the carbon 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 $2xCO_2$ and 12 GtC/yr at $4xCO_2$ (Figure 1b). When CO_2 stabilizes, so does 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 $2xCO_2$, NEP saturates at around 7 GtC/yr, and starts to decrease after 120 years despite increasing atmospheric CO_2. When CO_2 reaches $4xCO_2$, NEP only amounts to 5.5 GtC/yr, i.e. 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 $2xCO_2$ and 808 GtC at $4xCO_2$, which is respectively 23 % and 32 % lower than in the constant climate simulation (Table 1). The strong reduction of NEP in the climate change simulation (Figure 2) is mainly caused by a large decrease in tropical soil moisture (especially in South America). Qualitatively similar findings were reported previously. Here we show that the climate effect on the terrestrial carbon uptake is non-linear, the relative reduction in NEP at $4xCO_2$ (-32%) being larger than at $2xCO_2$ (-23%).

Climate Impact on Ocean Uptake

For the constant climate run, rising atmospheric CO_2 also increases the oceanic carbon uptake. At $2xCO_2$, the ocean carbon sink reaches 7.5 GTC/yr and 10.5 GtC/yr at $4xCO_2$ (Figure 1c). After atmospheric CO_2 stabilizes, the ocean uptake decreases as the ocean carbon tends toward a new equilibrium state. As for lands, the oceanic uptake is 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 $4xCO_2$, the oceanic uptake 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 $2xCO_2$, and 20% at $4xCO_2$ (Table 1). The reduction of oceanic CO_2 uptake under global warming is, at $2xCO_2$ similar to what previously found [*Maier-Reimer et al.*, 1996; *Sarmiento and Le Quéré*, 1996; *Sarmiento et al.*, 1998; *Matear and Hirst*, 1999]. The ocean uptake of CO_2 also responds non-linearly to the climate change, the climate impact on oceanic uptake being twice as large at 4xCO_2 than at 2xCO_2. 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 leads 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.

Impact on Derived Emissions

Our two estimates of both terrestrial and oceanic carbon uptakes allow us to determine the anthropogenic emissions compatible with the atmospheric CO_2 growth rate and a carbon cycle seing either a "constant climate" or a "changing climate". 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 4xCO_2, whereas they would peak at 40 GtC/yr in the climate change run (Figure 1d). When cumulated, the compatible emissions are respectively reduced by 8% and 13% at 2xCO_2 and 4xCO_2 when climate change is accounted for (Table 1). Thus, to achieve a given atmospheric CO_2 trajectory any economic CO_2 emission scenarii need to prescribe lower emissions to account for the climate impact on the carbon cycle. Furthermore, the response of the carbon cycle to the warming being non-linear, reductions in emissions will have to be increasingly stronger with time.

Climate System – Carbon Cycle Feedback

In the real world, where the carbon cycle is forced by anthropogenic emissions, our results would translate into a faster atmospheric CO_2 buildup (as land and ocean efficiencies to sequester carbon decrease with time), resulting in a more rapid climate change that may have further adverse impacts on terrestrial and oceanic processes and on atmospheric CO_2 concentration. In the following, we provide the first estimate of the magnitude of this positive feedback.

In a classical approach [*Hansen et al.*, 1984] (see Appendix 1), the gain of the climate system carbon cycle feedback, g, can be estimated 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 $2xCO_2$ to 0.003 K/ppmv at $4xCO_2$ (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 $2xCO_2$ to 60 ppmv/K at $4xCO_2$ (Figure 4b). The gain, g, defined in Appendix 1, amounts to 0.11 at $2xCO_2$ and reaches 0.19 at $4xCO_2$ (Figure 4c). The net feedback, f, which is the global warming amplification, also defined in Appendix 1, reaches 1.25 and 1.55 at $2xCO_2$ and $4xCO_2$ respectively (Figure 4d). Assuming that future emissions follow a trajectory compatible with today's climate (Figure 1d, constant climate simulation), one can use equations 6 and 7 from Appendix 1, to 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.

Conclusions

Our results suggest a large impact of future climate change on the carbon cycle, As the Earth warms up, there is a risk of seeing both ocean and biospheric capacity to absorb anthropogenic CO_2 significantly reduced, leaving larger CO_2 fraction in the atmosphere which in turn would enhance the climate change. In order to further explore these feedbacks, 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. A better understanding of the observed historical trends will help reduce uncertainties, and identify the key processes controlling CO_2 and climate requires. 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 [*Foley et al.*, 1996], and in the cycling of nutrients. Typically, one can expect deforestation to further enhance the positive feedback we calculated here. In that sense our estimate should be seen as a rather optimistic case. In addition, non-linear changes in the ocean-atmosphere dynamics [*Manabe et al.*, 1992; *Joos et al.*, 1999], could affect the magnitude of the feedback we have calculated here.

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Appendix 1. Climate-Carbon Cycle Feedback Calculation

If one considers the climate system, assuming conditions close to steady-state, the temperature change can be defined as :

$$dT = \frac{\partial^* T}{\partial C} dC + dT_{ind} \qquad \text{(A1)}$$

where $\partial^* T/\partial C$ is the overall temperature sensitivity to atmospheric CO_2 (through changes in the radiative forcing, accounting for direct and indirect physical effects) and dT_{ind} is the temperature change independent of CO_2 (eg. as a response to solar forcing). We introduce the ∂^* notation to highlight the fact that this definition of climate sensitivity, classically called total derivative as it includes all known physical feedbacks, is still a partial derivative in the present context as it omits the climate-carbon cycle feedback. Similarly, for the carbon cycle, one can define the atmospheric CO_2 change as:

$$dC = \frac{\partial^* C}{\partial T} dT + dC_{ind} \qquad \text{(A2)}$$

with $\partial^* C/\partial T$ being the overall atmospheric CO_2 sensitivity to temperature (through changes in the carbon cycle due to changes in climate and circulation) and dC_{ind} the atmospheric CO_2 change independent of temperature change (eg. due to anthropogenic emissions).

Now, in the coupled climate–carbon cycle system one can combine these two equations into :

$$dT = \frac{\partial^* T}{\partial C}\frac{\partial^* C}{\partial T} dT + \frac{\partial^* T}{\partial C} dC_{ind} + dT_{ind} \qquad \text{(A3)}$$

which can be rewritten as :

$$dT = \frac{1}{1-g} dT_{unc} = f dT_{unc} \qquad \text{(A4)}$$

Similarly :

$$dC = \frac{1}{1-g} dC_{unc} = f dC_{unc} \qquad \text{(A5)}$$

where g is the gain of the climate system–carbon cycle feedback, defined as $\partial^* T/\partial C \times \partial^* C/\partial T$, dT_{unc}, and dC_{unc} are, respectively, the temperature and the CO_2 changes in an uncoupled system deduced from equations 1 and 2. $f = 1/(1-g)$, the net feedback factor, is the global warming amplification due to the climate–carbon feedback loop. Equations 4. and 5. can be integrated :

$$\Delta T(t) = \int_0^t f dT_{unc} \qquad \text{(A6)}$$

$$\Delta C(t) = \int_0^t f dC_{unc} \qquad \text{(A7)}$$

$\partial^* T/\partial C$, $\partial^* C/\partial T$, g, f, and therefore $\Delta T(t)$ and $\Delta C(t)$ can be easily approximated from our two simulations.

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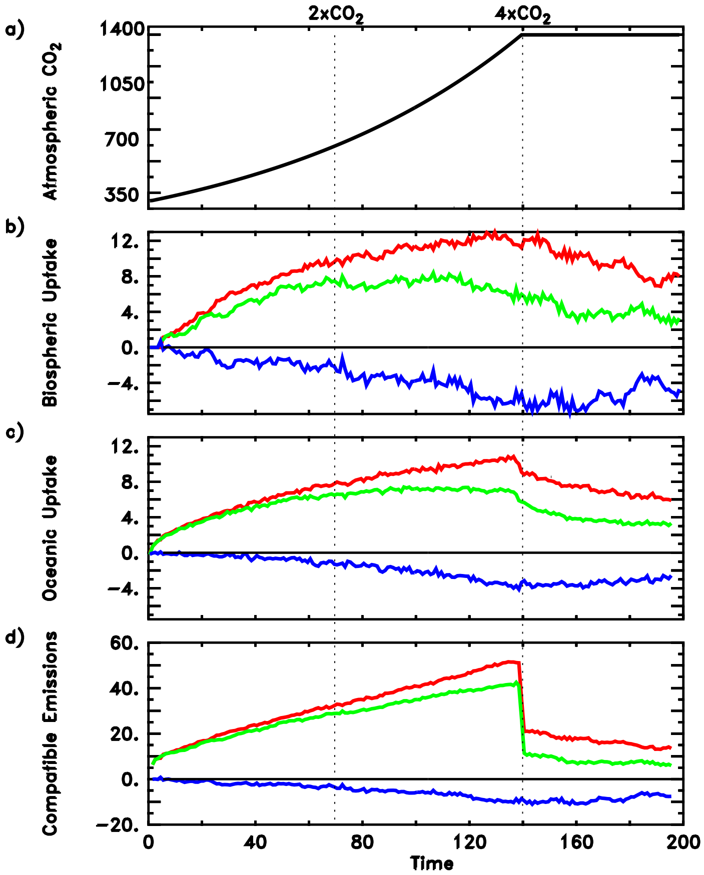

Figure 1. *Carbon budget. a) Atmospheric* CO_2 *scenario used as a forcing for the climate model (in ppmv) (12). b) Simulated annual biospheric* CO_2 *uptake (GtC/yr) for the constant climate simulation (red line), the climate change simulation (green line) and the difference between the two simulations, 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).*

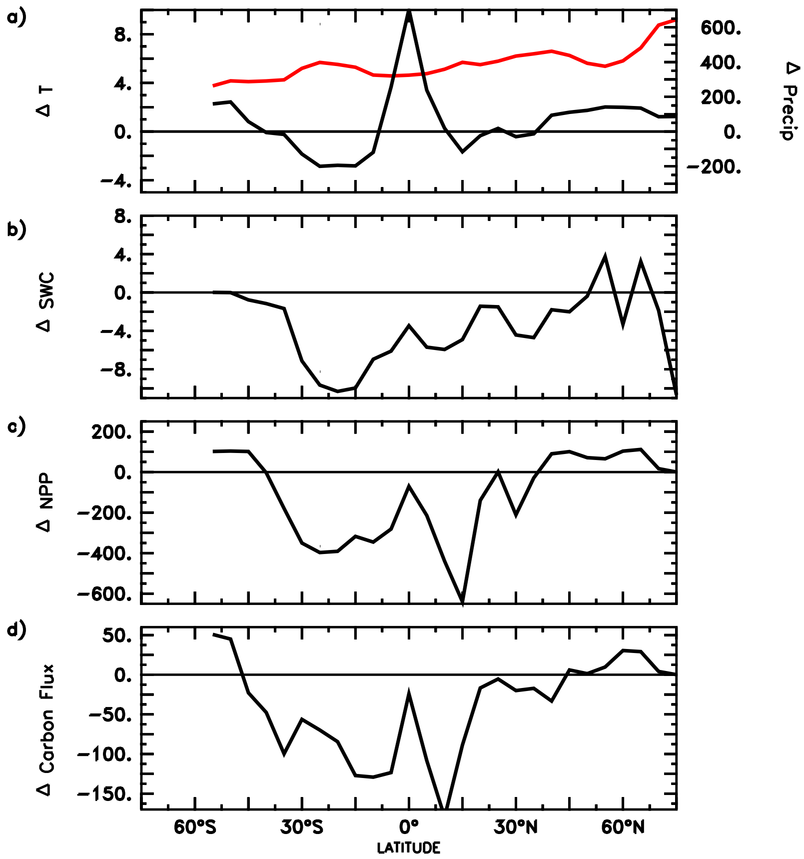

Figure 2. Zonal mean difference between the climate change and the constant climate simulations at the time of 4xCO_2 of a) annual surface land temperature (°C) (red line) and precipitation (mm/yr) (blue line), b) soil water content (mm), c) Net Primary Productivity (gC/m^2/yr), and d) net carbon uptake (gC/m^2/yr).

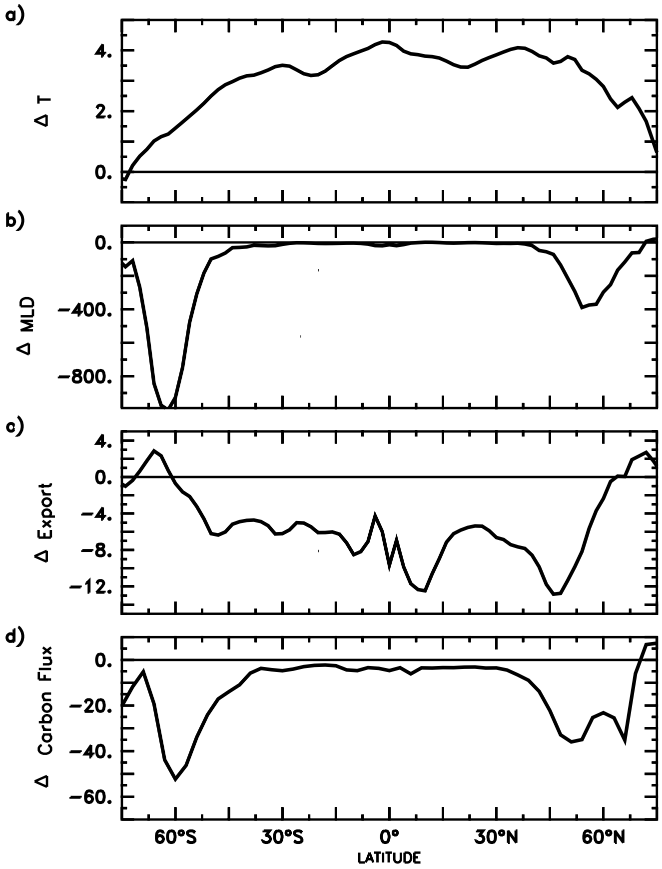

Figure 3. Zonal mean difference between the climate change and the constant climate simulations at the time of $4xCO_2$ of a) sea surface temperature (°C), b) depth of the mixed layer (m), c) export production ($gC/m^2/yr$), and d) net carbon uptake ($gC/m^2/yr$).

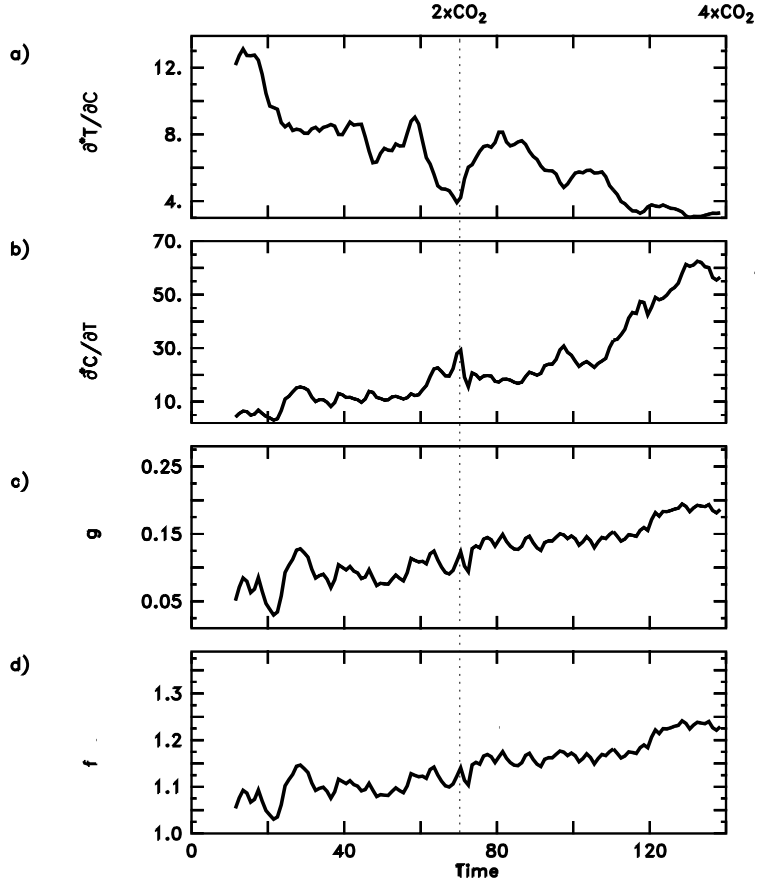

Figure 4. Time evolution of a) $\partial^*T/\partial C$, the overall sensitivity of surface temperature to the atmospheric CO$_2$ (10^3 K/ppmv), b) $\partial^*C/\partial T$, the overall sensitivity of atmospheric CO$_2$ to surface temperature (ppmv/K), c) g, the gain of the climate system-carbon cycle feedback calculated as $\partial^*T/\partial C \times \partial^*C/\partial T$, and d) f, the global warming amplification calculated as $f = 1/(1-g)$ (see Appendix 1).

Table 1. Changes in cumulated carbon budget at $2xCO_2$ and $4xCO_2$

	$2xCO_2$		$4xCO_2$	
	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%)

Déjà paru :

6 : Mai 1998 Emmanuelle Cohen-Solal and Hervé Le Treut, *Long term climate drift of a coupled surface ocean-atmosphere model : role of ocean heat transport and cloud radiative feedbacks*

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http://www.ipsl.jussieu.fr/modelisation/liste-notes.html.