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Abstract

CO$_2$-fertilization affects plant growth, which modifies surface physical properties, altering the surface albedo, and fluxes of sensible and latent heat. We investigate how such CO$_2$-fertilization effects on vegetation and surface properties would affect the climate system. Using a global three-dimensional climate-carbon model that simulates vegetation dynamics, we compare two multi-century simulations: a "Control" simulation with no emissions, and a "Physiol-noGHG" simulation where physiological changes

occur as a result of prescribed CO_2 emissions, but where CO_2-induced greenhouse warming is not included. In our simulations, CO_2-fertilization produces warming; we obtain an annual- and global-mean warming of about 0.65 K (and land-only warming of 1.4 K) after 430 years. This century-scale warming is mostly due to a decreased surface albedo associated with the expansion of the Northern Hemisphere boreal forests. On decadal time scales, the CO_2 uptake by afforestation should produce a cooling effect that exceeds this albedo-based warming; but if the forests remain in place, the CO_2-enhanced-greenhouse effect would diminish as the ocean equilibrates with the atmosphere, whereas the albedo effect would persist. Thus, on century time scales, there is the prospect for net warming from CO_2-fertilization of the land biosphere. Further study is needed to confirm and better quantify our results.

Introduction

When the atmospheric CO_2 concentrations increase, the land biosphere can take up more CO_2 per unit of water loss. This makes more carbon available for building woody material, favoring forested ecosystems over grasslands. In northern latitudes, these effects would tend to select boreal forests over tundra, since the former absorb more solar radiation and warm the surface (Bonan et al, 1992). This warming would also tend to make conditions less harsh, thus favoring the spread of boreal forests (Foley et al., 1994, Grace et al. 2002). Alterations in evapotranspiration due to CO_2-fertilization-induced changes in stomatal conductance and the amount of leaf area could also impact the climate. Here, we investigate the climate effects of such CO_2-induced changes in vegetation physiology, structure and distribution.

When water and nutrients are available, photosynthesis by land plants is expected to increase with atmospheric CO_2 content via the so-called CO_2-fertilization effect (Owensby et al. 1999), leading to increased carbon uptake. Higher CO_2 concentrations promote water-use efficiency of plants, leading to stimulated plant growth (Polley et al.

1993). However, increased global temperatures also result in increased heterotrophic soil respiration rates (Lloyd and Taylor 1994), diminishing or even reversing the net CO_2 flux from the atmosphere to the land biosphere (Cox et al. 2000; Friedlingstein et al. 2001; Thompson et al. 2004; Zeng et al. 2004; Govindasamy et al. 2005; Mathews et al. 2005).

Thompson et al. (2004) showed that CO_2-fertilization could strongly damp the global warming but nutrient limitations could weaken this effect. They performed a "Fertilization" simulation in which the land biosphere used the model-predicted atmospheric CO_2 concentration and a "Saturation" simulation in which the effects of CO_2-fertilization were assumed to be saturated at year 2000 atmospheric CO_2 levels. The Saturation simulation represented the possibility that nutrients or other factors could limit the CO_2-fertilization effect. The land biosphere was a very strong sink of carbon in the Fertilization simulation through the year 2100, but it became a source of carbon to the atmosphere in the Saturation case.

However, CO_2-fertilization can have climate consequences in three other ways. First, it has the potential to modify the surface properties by enhancing plant growth, changing the vegetation distribution and altering the surface albedo (Levis et al. 1999; 2000). Second, CO_2 fertilization promotes water use efficiency by inducing stomatal closure (Polley et al. 1993; Owensby et al. 1999), producing warming via a reduction of evapotranspiration (Henderson-Sellers et al. 1995; Sellers et al. 1996), and possibly also an increase in runoff (Gedney et al. 2005). Finally, CO_2-fertilization may enhance leaf area and increase evapotranspiration (Betts et al. 1997), which would have a cooling effect. We refer to the collective impacts of these vegetation changes on the physical climate as the "biogeophysical" effects of CO_2-fertilization. In contrast, we refer to the potential amelioration of the greenhouse radiative impact of CO_2 resulting from its sequestration by vegetation as the "biogeochemical" effects of CO_2-fertilization.

How large are the biogeophysical effects of CO_2 fertilization? In equilibrium experiments (run over a few decades) using a mixed layer ocean model coupled to

atmospheric and dynamic vegetation models, Levis et al. (1999) showed that vegetation changes due to a doubling of CO_2 could produce a warming of about 1.5 ^{0}C in summer and spring in the high latitudes. The warming was mainly brought about by reduction in surface albedo; however, there was no global mean warming in their decadal simulations (Levis et al. 2000). In a simulation that did not include the radiative effects of increasing CO_2 concentrations, Cox et al. (2000) obtained a slight warming over land by the year 2100 due to CO_2-induced changes in stomatal conductance and vegetation distribution. In their coupled climate-carbon model simulation, Notaro et al. (2005) obtained a global mean warming of about 0.1 K during the historical period (pre-industrial to present-day) due to the physiological effects of increasing atmospheric CO_2. The warming signals in Levis et al. (2000), Cox et al. (2000), and Notaro et al. (2005) were small because the simulations were run for only a few decades or a century. However, dynamic changes in vegetation distribution could take many centuries.

Using a coupled climate-carbon cycle model that simulates vegetation dynamics (Govindasamy et al. 2005; Bala et al. 2005, hereafter BA05), we compare two *multi-century* simulations in this study: a "Control" simulation with no emissions, and a "Physiol-noGHG" simulation with prescribed emissions. The Physiol-noGHG simulation allows CO_2-fertilization and other physiological changes, but has no direct CO_2-induced greenhouse climate change. Our simulations indicate that CO_2-fertilization could produce a biogeophysical climatic effect of warming over a timescale of a few centuries that would offset at least part of the biogeochemical effect of cooling. We find that this warming is mostly associated with an albedo decrease resulting from the expansion of the Northern Hemisphere boreal forests.

Model

To investigate the CO_2-fertilization impacts of climate change due to anthropogenic emissions, we use INCCA (INtegrated Climate and CArbon), a coupled climate and carbon cycle model (Thompson et al. 2004; Govindasamy et al. 2005; BA05)

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developed at Lawrence Livermore National Laboratory (LLNL). The physical ocean-atmosphere model is the NCAR/DOE PCTM model (Meehl et al. 2004), which is a version of the NCAR CCM 3.2 model (Kiehl et al. 1996) coupled to the Los Alamos National Laboratory's POP ocean model (Maltrud et al. 1998). The climate model is coupled to a terrestrial biosphere model, the Integrated Biosphere Simulator version 2 or IBIS2 (Foley et al. 1996; Kucharik et. al. 2000) and to a prognostic ocean biogeochemistry model based on the diagnostic Ocean Carbon-cycle Model Intercomparison Project (OCMIP) Biotic protocols (Najjar and Orr 1999). The horizontal resolution of the land and atmosphere models is approximately 2.8° in latitude and 2.8° in longitude, while the ocean model has a horizontal resolution of (2/3)°. The atmosphere and ocean models have 18 and 40 vertical levels, respectively. Flux adjustments are not applied to the physical climate model.

Experiments

We perform two model simulations starting from year-1870 pre-industrial initial conditions created by more than 200 years of spin up: a Control case with no CO_2 emissions for the period 1870-2300, and a Physiol-noGHG case in which emission rates are prescribed out to year 2300, as discussed in BA05. In the Control case, climate drift for the period 1900-2300 is a –0.62 K (~ -0.15 per century) change in mean surface temperature, and a 3.8 ppmv increase in atmospheric CO_2 concentration. The drift in sea ice area and volume for this 400-year period are +15.3 % and + 40 % respectively.

In the Physiol-noGHG case, the land and ocean carbon cycle models use the predicted atmospheric CO_2 content associated with the prescribed emissions, but the radiation calculation uses the pre-industrial concentrations of CO_2 and other greenhouse gases: there is no change in greenhouse gas radiative forcing for the period 1870-2300, and hence there is no CO_2-enhanced greenhouse climate change. This is the (perhaps misnamed) "constant-climate" simulation as discussed in several coupled climate-carbon

cycle modeling studies (e.g. Cox et al. 2000; Friedlingstein et al. 2001; Govindasamy et al. 2005). The results from the corresponding fully interactive climate-carbon experiment are discussed in BA05.

As described in BA05, for our Physiol-noGHG case, CO_2 emissions are specified at historical levels for the period 1870-2000 (Marland et al. 2002) and at SRES A2 levels for the period 2000-2100 (IPCC 2001). For the years 2100-2300, emission rates follow a logistic function for the burning of the remaining conventional (coal, oil and natural gas) fossil-fuel resources (assuming 5270 gigatons of carbon (Gt C) in 1750; Metz et al. 2001). Non-CO_2 greenhouse gas concentrations are specified at historical levels for 1870-2000 and at SRES A2 levels for 2000-2100 (IPCC 2001) and are fixed at 2100 levels thereafter. Land use emissions are taken from Houghton (2003) for the historical period, and from the SRES A2 scenario for the years 2000-2100, and are set to zero thereafter. There is no change in aerosol forcing. The cumulative emission for the entire period 1870-2300 is 5404 Gt C. Increasing from the present day values of 8 Gt C per year, the emission rate peaks at about 30 PgC per year in the 22nd century around year 2120, and then it declines sharply in the 23rd century to almost zero by year 2300 (see BA05).

Results

Here, we focus our discussion on differences between the experiment (Physiol-noGHG) and Control simulations. Compared to the Control, the global- and annual-mean temperature in the Physiol-noGHG simulation begins to rise after year 2050 (Fig. 1a). This global-mean temperature increase coincides with the sharp rise in atmospheric CO_2 concentrations after the year 2050 (Fig. 1b). The warming is as high as 1K in the 22nd century; thus, CO_2-fertilization has the potential to change the global climate on centennial time scales. The global- and annual-mean warming in the last decade of our simulation (2291-2300) is 0.65 K (1.4 K over land) which will be considered as the INCCA model's estimate of biogeophysical impact of CO_2-fertilization in the rest of our

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discussion. The global mean temperature change 0f 0.65 K is only a little larger than the residual drift in the Control, but it is opposite in sign. Hence, there is no prima facie reason to assume that the drift in the control affects our basic conclusions.

In the Physiol-noGHG case, the atmospheric CO_2 concentration increases to1166 ppmv from its pre-industrial value of 289 ppmv (Fig. 1). When this coupled climate and carbon cycle model also included the simulation of the radiative effects of greenhouse gases, it predicted 1423 ppmv and an 8 K warming for the same emissions scenario (BA05). This result suggests that the radiatively driven CO_2-induced greenhouse warming (i.e. the climate-carbon cycle feedback) in this model contributes 257 ppmv to the atmosphere by 2300 through reduced land and ocean carbon uptake. The CO_2 concentration is decreasing (Fig. 1) near the end of the simulation (year 2300) due to the continued uptake of about ~2 GtC per year by both land and oceans (Fig. 2) when the emissions rates are almost zero (see BA05). Fig.2 suggests that the response time scale for uptake by the land to increasing atmospheric CO_2 is much shorter (decades) than that by the oceans (centuries).

The global- and annual-mean change evinced by different climate variables (Physiol-noGHG case minus control in the decade of 2291-2300) is shown in Table 1. For example, while the changes in global mean precipitation are negligible (less than 1%), the magnitude of change in ice volume is about 37 %. The change in the net short wave absorption at the top of the atmosphere (TOA) is 0.88 Wm^{-2}. Assuming a radiative forcing of 3.4 Wm^{-2} for a doubling of CO_2 and an equilibrium climate sensitivity of 2.1 K for this model (IPCC, 2001), this implies an equilibrium warming of about 0.55 K, which is in close agreement with the simulated transient warming of 0.65 K.

The vegetation distributions in the two experiments are listed in Table 2. IBIS2 simulates only the natural potential vegetation so that land-use changes are not included. There is an increase in the tropical, temperate and boreal forests due to CO_2-fertilization (Fig.3). The total forested land fraction increases by 13 % of land area, with tropical,

temperate, and boreal forests expanding by about 4, 5, and 4 % respectively (Table 2). During the same period, deserts, polar deserts, grasslands, and savanna decline by a similar fraction. There is a northward expansion of boreal forests at the expense of tundra (Fig. 3), and tundra also moves northward into the land ice area. Grasslands, savanna, and deserts are replaced by temperate forests in Australia, South Africa, and North and South America. Tropical forests in Africa expand slightly to the north at the expense of savanna and grasslands. The CO_2-induced physiological changes in vegetation presumably help to counter effects of desertification that may occur as a result of climate change (Table 2). Because these changes to natural vegetation take many centuries to unfold, only models (such as the one used in this study) that include vegetation dynamics, and that can be run for multiple centuries, can adequately simulate the vegetation distribution changes.

However, our study also has some inherent limitations. We do not take into account the constraint that land use imposes on the development of natural ecosystems. In our Physiol-noGHG simulation, we only account for the global carbon emissions due to land use by taking Houghton's (2003) estimate for the historical period, the SRES A2 scenario for the period 2000-2100, and zero thereafter (BA05). Because the IBIS model allows only natural vegetation to grow, we do not simulate the effects of agricultural crops. Thus, forests would not actually be able to grow as simulated here because a large part of the land surface would be under some form of cultivation (Ramankutty et al. 2002). This would limit carbon uptake by the land biosphere because most cultivated ecosystems do not accumulate biomass, and under current management practices, very small amounts of litter enter the soil. In summary, Fig. 3 shows only the potential vegetation distributions, since it does not include the anthropogenic land use changes.

The expansion of forests and shrinking of grasslands and deserts decreases the albedo (Fig. 4a) and warms the surface (Fig. 4b) via increased absorption of solar radiation. The surface warming, in turn, leads to further forest expansion and changes in the spatial distributions of other vegetation types (Table 2). The local changes in albedo

and surface temperature are as large as -25 % and +8 K, respectively, in Siberia and Canada. The land-only mean temperature change in the Northern Hemisphere high latitudes (50^0N to 90^0N) is 2.6 K (Table 3) and the corresponding increase in the net surface absorbed shortwave flux is 12.5 Wm^{-2}. This warming is in agreement with previous studies (Hansen et al. 1997; Levis et al. 1999; Betts 2000; Govindasamy et al. 2001; Gibbard et al. 2005) which implied that increased temperate and boreal forest cover lead to warming through a decreased surface albedo, especially at high latitudes (since snow-covered vegetation has much lower albedo than snow-covered bare ground). Local warming of lesser magnitude (2-4 K) is simulated over Europe, Australia, South Africa, and eastern parts of North America and South America (Fig. 4b).

The average land-only annual mean warming in the Northern Hemisphere mid-latitudes (20^0N to 50^0N), Tropics (20^0S to 20^0N), and Southern Hemisphere mid-latitudes (50^0S to 20^0S) are 1.5, 0.9 and 1.1 K, respectively (Table 3). As in NH high-latitudes, albedo effects dominate in the NH mid-latitudes with associated increases in net surface solar absorption. In the Tropics and SH mid-latitudes, the albedo decrease is small (Fig. 4a; Table 3), and instead there are increases in the cloudiness, atmospheric column water vapor, and downwelling surface long wave radiation. We caution that the multivariate climate change indicated by Table 3 does not distinguish "causes" from "effects" because it includes the results of all the feedbacks. A detailed feedback analysis is beyond the scope of this paper. Because the increased cloudiness has resulted in reduced downwelling solar radiation and solar absorption at the surface, shortwave cloud feedbacks are unlikely to be the primary source of the warming. We instead infer that the net tropical and SH mid-latitude warming is probably due primarily to cloud/water vapor longwave feedbacks that are initiated by physiological changes to the vegetation.

Because the Leaf Area Index (LAI) and latent heat flux increase in all land areas (Table 3), it is likely that the increases in evapotranspiration (and hence latent heat flux) due to increased leaf area (cooling effect; Betts et al. 1997) dominates over the decrease

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due to reduced stomatal conductance (warming effect; Sellers et al., 1996). The net surface longwave heat loss also increases in all land areas (Table 3) because of the increase in the Leaf Area Index (LAI) and the associated increase in the surface emissivity (Levis et al. 1999). Cooling associated with increases in both the fluxes of latent heat and of the net surface longwave radiation tends to offset part of the warming.

Discussion

This study demonstrates that the biogeophysical effects of CO_2-fertilization could have non-negligible effect on centennial time scales. To some extent, these results may depend on the chosen model and/or experimental setup. Our terrestrial biosphere model, IBIS2, exhibits higher uptake in comparison to other models (Fig. 2; Prentice et al., 2001), and thus, our estimate of the biogeophysical climate impact of CO_2-fertilization is perhaps closer to the upper limit of possible outcomes. The higher sensitivity of IBIS2 to CO_2-fertilization may be associated with the lack of nutrient cycles (Govindasamy et al., 2005). In addition, we have not prescribed realistic future land use change, so as to correctly account for the different effects of crops and natural ecosystems which might produce overall cooling. For example, Govindasamy et al. (2001) and Brovkin et al. (1999) simulated a global cooling of –0.25 K and –0.35 K, respectively, due to changes from potential natural vegetation to the present-day vegetation distribution. Therefore, our simulated biogeophysical effect of climate warming due to CO_2-fertilization could be partially offset by human-induced deforestation which would produce cooling via increases in the surface albedo. On the other hand, the biogeophysical warming effects could also be enhanced by an anthropogenic land cover change such as afforestation that might be promoted for the purpose of sequestering atmospheric carbon dioxide.

The relative importance of the biogeophysical and the biogeochemical effects of CO_2-fertilization can be assessed by estimating the possible climate consequences of CO_2-fertilization on atmospheric CO_2 concentrations. At the end of our simulations, the

land biosphere contains ~2500 Pg more carbon in the Physiol-noGHG case than in the Control case. If half of this 2500 PgC were to remain in the atmosphere in the absence of CO_2-fertilization, then the atmosphere would contain an additional 1250 PgC, or about another 590 ppmv CO_2. If this 590 ppmv were added to the 1166 ppmv in the atmosphere at the end of the Physiol-noGHG simulation, it would produce an equilibrium warming of ~1.2 K, given our model's estimated climate sensitivity of 2.1 K, and a radiative forcing of 3.4 Wm^{-2} per CO_2-doubling.

This suggests that for our model and emission scenario, the biogeochemical climate effects of CO_2-fertilization would produce 1.2 K global cooling, but the biogeophysical effects would produce a global warming of 0.65 K. These effects are opposite in sign, but are of the same order of magnitude. Moreover, the change in atmospheric CO_2 resulting from a land cover change will diminish over time, whereas the dominant biogeophysical effect of reduced albedo will persist; the perturbation to atmospheric CO_2 content would be damped by equilibration with the ocean and ultimately with the rock cycles. Therefore, we infer that the biogeophysical climatic effects of land-cover change induced by CO_2-fertilization could cancel most of the biogeochemical effects, especially regionally and on centenial time scales. Additional research is needed to better quantify these effects.

Acknowledgements

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Figure captions

Figure 1. Evolution of drift-corrected global- and annual-mean surface temperature (upper panel) and atmospheric CO_2 concentration (lower panel) in the Physiol-noGHG simulation. Mean surface temperature warming in the Physiol-noGHG case in the last

decade (years 2291-2300) is about 0.65 K relative to Control, and the atmospheric CO_2 concentration is 1166 ppmv at year 2300.

Figure 2. Evolution of global- and annual-mean carbon fluxes between atmosphere and land (upper panel), and between atmosphere and ocean (lower panel) in the control and Physiol-noGHG simulations. Negative fluxes indicate uptakes of carbon by both land and oceans.

Figure 3 Dominant vegetation distributions in Control (top panel) and Physiol-noGHG (bottom panel) cases in the last 30 years of the simulations (2271-2300). Antarctica is not shown. There is an expansion of boreal forests into tundra regions in the Physiol-noGHG case, and the tundra vegetation moves into the land ice area of the control. The expansion of temperate forests also can be seen in Australia, South Africa, and North and South America.

Figure 4 Difference in surface albedo (top panel) and surface temperature (bottom panel) between the Physiol-noGHG and Control cases. The differences are computed over the decade 2291-2300. The expansion of forests at the expense of grasslands and savanna due to CO_2-fertilization leads to a decrease in the surface albedo and an increase in surface temperature over Northern high-latitude land areas.

Table 1: Changes in global- and annual-means of climate variables in the Physiol-noGHG case (decade of 2291-2300 minus 1891-1900). The changes are corrected for the drift in the Control simulation.

Experiment	Surface Temp. (K)	Precip. (%)	Sea ice extent (%)	Sea ice volume (%)	Net short wave flux at TOA (Wm^{-2})	Net flux at TOA (Wm^{-2})
Physiol-noGHG	0.65	0.4	-5.3	-36.8	0.88	0.01

Table 2: Fraction of land area occupied by dominant vegetation types at the end of simulations in the Control and Physiol-noGHG cases (dominant vegetation during 2271-2300). In the Physiol-noGHG case, the forested and tundra fractions have increased, and those of grasslands and deserts have decreased.

Vegetation type	Control	Fertilized
Tropical forests	22.1	26.3
Temperate forests	19.2	24.1
Boreal forests	6.4	10.4
Savanna, Grasslands & Shrub lands	12.4	7.0
Tundra	6.2	8.3
Desert	16.7	13.4
Polar desert	17.0	10.5

Table 3: Changes in average land-only annual means of climate variables by latitude belt in the Physiol-noGHG case (decade of 2291-2300 minus 1891-1900). The changes are corrected for the drift in the Control simulation. Decrease in the surface albedo appears to be the main source of warming in the NH high- and mid-latitudes. In the Tropics and SH mid-latitudes, there are increases in cloudiness, column water vapor, and downwelling long wave radiation, suggesting that cloud/water vapor feedback is the source of the surface warming.

Variable	All land	SH mid-lat 50^0S to 20^0S	Tropics 20^0S to 20^0N	NH mid-lat 20^0N to 50^0N	NH high-lat 50^0N to 90^0N
Sfc.Temp. (K)	1.41	1.06	0.94	1.52	2.56
Albedo (%)	-2.3	-1	-0.5	-1.5	-7.7
LAI (m^2/m^2)	4.06	5.88	6.91	2.87	2.10
Sfc. net SW flux (Wm^{-2})	2.33	0.42	-3.54	2.73	12.52
Sfc. net LW flux (Wm^{-2})	2.52	2.53	0.06	2.54	7.50
Sensible heat flux (Wm^{-2})	-0.98	-4.04	-3.88	-0.85	4.06
Latent heat flux (Wm^{-2})	0.65	2.1	0.13	0.52	1.00
Total cloud (%)	0.19	0.75	1.65	-0.08	-3.53
Low cloud (%)	-0.97	0.80	1.77	-1.14	-6.24
Sfc SW down (Wm^{-2})	-0.74	-2.17	-5.60	-0.06	6.22
Sfc. LW down (Wm^{-2})	4.47	3.25	5.38	5.23	3.86
Column water vapor (mm)	0.58	0.06	0.77	0.65	0.72

Figure 1

Figure 2

Figure 3

Dominant Vegetation Types (2271-2300)

Control

Physiol-noGHG

Tropical Temperate Boreal Forests   Savana, Grasslands   Tundra Desert Ice

Figure 4

Surface Albedo change

Surface Temperature change (K)