Evaluating the carbon cycle of a coupled atmosphere-biosphere model

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[1] We investigate how well a coupled biosphere-atmosphere model, CCM3-IBIS, can simulate the functioning of the terrestrial biosphere and the carbon cycling through it. The simulated climate is compared to observations, while the vegetation cover and the carbon cycle are compared to an off-line version of the biosphere model IBIS forced with observed climatic variables. The simulated climate presents some local biases that strongly affect the vegetation (e.g., a misrepresentation of the African monsoon). Compared to the off-line model, the coupled model simulates well the globally averaged carbon fluxes and vegetation pools. The zonal mean carbon fluxes and the zonal mean seasonal cycle are also well represented except between 0° and 20° N due to the misrepresentation of the African monsoon. These results suggest that, despite regional biases in climate and ecosystem simulations, this coupled atmosphere-biosphere model can be used to explore geographic and temporal variations in the global carbon INDEX TERMS: 1615 Global Change: Biogeochemical processes (4805); 1620 Global Change: cycle. Climate dynamics (3309); 1818 Hydrology: Evapotranspiration; 3319 Meteorology and Atmospheric Dynamics: General circulation; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; KEYWORDS: carbon cycle in terrestrial biosphere, climate-vegetation interactions, vegetation dynamics, climate model, soil carbon

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1. Introduction

[2] Emissions of carbon dioxide (CO_2) from fossil fuel combustion and land use practices are increasing the atmospheric CO₂ concentration and affecting the Earth's radiation balance and climate. The continental biosphere and the oceans currently absorb about half of the anthropogenic CO₂ emissions, but this rate of uptake depends on the functioning of the biosphere and the oceans, which in turn depend on the climate and the atmospheric CO₂ concentration. To predict future changes in CO₂ concentration and climate, it is necessary to take into account the potential feedbacks between atmospheric CO₂ concentration, climate, and carbon exchange between the atmosphere, oceans, and biosphere. Therefore, climate models should include explicit representations of the global carbon cycle, including the interactions between atmospheric, oceanic, and terrestrial carbon pools.

[3] Two recent modeling studies [*Cox et al.*, 2000; *Friedlingstein et al.*, 2001], using very different model configurations, suggest that biospheric and oceanic carbon uptake could be significantly reduced by future climate

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change. However, their results differ in the magnitude of the biospheric signal. In the work of Cox et al. the biospheric uptake is reduced to zero and the biosphere becomes a net source of carbon, enhancing the CO_2 increase due to anthropogenic activity, while in the work of Friedlingstein et al. the uptake is reduced but the biosphere remains a net sink of carbon.

[4] These contrasting results raise the question of how well a terrestrial biosphere model coupled to a general circulation model (GCM) can simulate the carbon cycle in the first place. GCMs effectively simulate the large-scale average seasonal distribution of pressure, temperature, circulation, and extratropical precipitation but their skill at simulating regional climates is low [*Gates*, 1995]. Do those biases in the regional climate allow for a reasonable simulation of the functioning of the biosphere? Here, we try to answer this question with a coupled biosphere-atmosphere model: the terrestrial ecosystem/land-surface model IBIS [*Kucharik et al.*, 2000] coupled to the NCAR CCM3 [*Kiehl et al.*, 1998].

[5] There have been a number of modeling studies of the carbon exchange between the terrestrial biosphere and the atmosphere either with fluxes simulated by land surface models directly coupled to a GCM or advected with a transport model [e.g., *Fung et al.*, 1987; *Knorr and Hei*-

mann, 1995; *Denning et al.*, 1996; *Nemry et al.*, 1996] (see *Craig et al.* [1998] for an exhaustive list). These studies focused on the short-term exchange of carbon and used land surface models that did not represent vegetation dynamics or changing carbon pools. Our study builds on this research by representing the full dynamic carbon cycling in the vegetation and soil.

[6] We present here the mean carbon cycle simulated by the coupled biosphere-atmosphere model CCM3-IBIS at the end of a long (350 year) run with fixed sea-surface temperatures. The interannual variability and the long-term variability of the vegetation and the carbon cycle will be discussed in subsequent papers.

2. IBIS-2: Model Description

[7] In this study, we use an updated version of the Integrated Biosphere Simulator (IBIS) of *Foley et al.* [1996] and *Kucharik et al.* [2000]. IBIS (version 2) is a comprehensive model of terrestrial biospheric processes, and includes land-surface physics, canopy physiology, plant phenology, vegetation dynamics and competition, and carbon cycling.

[8] The IBIS land surface module simulates the energy, water, carbon, and momentum balance of the soil-vegetation-atmosphere system on a short time step consistent with GCMs (~20-60 min). The land surface module borrows much of its basic structure from the LSX land surface package [Thompson and Pollard, 1995a, 1995b]. The module includes two vegetation layers (i.e., "trees" and "grasses and shrubs") and six soil layers to simulate soil temperature, soil water, and soil ice content over a total depth of 4 m. Physiologically based formulations of C3 and C4 photosynthesis [Farquhar et al., 1980], stomatal conductance [Collatz et al., 1991, 1992] and respiration [Amthor, 1984] are used to simulate canopy gas exchange processes. This approach provides a mechanistic link between the exchange of energy, water, and CO_2 between vegetation canopies and the atmosphere. Budburst and senescence depend on climatic factors following the empirical algorithm presented by Botta et al. [2000].

[9] The annual carbon balance allows the vegetation dynamics submodel to predict the maximum leaf area index and biomass for 12 plant functional types (pfts), which compete for light and water. IBIS represents vegetation dynamics using very simple competition rules. The relative abundance of the 12 pfts in each grid cell changes in time according to their ability to photosynthesize and use water. For example, in a grid cell where trees and grasses coexist, grasses are shaded by trees and receive less light but their rooting profile allows them to withdraw water first as it infiltrates through the soil. In drought conditions, grasses will be favored, trees will accumulate less carbon, will grow less leaves and eventually wilt. Competition between grass types or between tree types result from different allocation, phenology, type of leaf, or type of photosynthesis leading to different annual carbon balances.

[10] IBIS simulates carbon cycling through the vegetation, litter and soil organic matter (Figure 1). The soil biogeochemistry module is similar to the CENTURY model [*Parton et al.*, 1993] and the biogeochemistry model of *Verberne et al.* [1990]. The total belowground carbon in the first meter of soil is divided into pools characterized by their residence time: from a few hours for the microbial biomass to more than 1000 years for stabilized organic matter. Decomposition rates of litter and soil carbon depend on soil temperature and soil moisture.

[11] IBIS has been extensively tested against site-specific biophysical measurements from flux towers [*Delire and Foley*, 1999] as well as spatially extensive ecological [*Kucharik et al.*, 2000] and hydrological data [*Lenters et al.*, 2000]. IBIS is explicitly designed to work within atmospheric models and has been extensively used in both uncoupled [e.g., *Foley et al.*, 1996; *Kucharik et al.*, 2000] and coupled modes [e.g., *Foley et al.*, 2000, 1998; *Levis et al.*, 1999a, 1999b, 1999c, 2000].

3. Simulations and Data Sets Used

[12] We performed two simulations: one in with IBIS coupled to the NCAR CCM3 atmospheric model [*Kiehl et al.*, 1998] and one with IBIS forced with observed climate data for the twentieth century. In this way, we can isolate the biases due to the terrestrial ecosystem model IBIS only, due to the atmospheric model CCM3 and due to the interaction between the vegetation and the atmosphere. Biases due to the off-line model IBIS alone have been described in an earlier paper [*Kucharik et al.*, 2000] and will not be emphasized here.

[13] In this study, the coupled model CCM3-IBIS is run for 300 years at a resolution of T31 (the spectral representation of the horizontal fields is truncated at the 31st wave number using a triangular truncation. Horizontal fields are converted to a $\sim 3.75^{\circ} \times 3.75^{\circ}$ grid). Fixed climatological sea surface temperatures (SSTs) are used. The T31 resolution offers an acceptable compromise between computing costs and accuracy of the simulated climate. To limit computing costs, we accelerate the vegetation dynamics module so that it achieves 150 years during 20 years of actual atmospheric model simulation.

[14] The off-line version of IBIS is run at the same spatial resolution over 500 years. We use the monthly data set of temperature, precipitation, relative humidity, and cloudiness from 1901 to 1995 compiled by *New et al.* [2000] (referred to as CRU05) as forcing to the off-line model. To run our simulation over 500 years, we interpolate the data at the GCM T31 resolution, construct a detrended 60-year climate record (from 1936 to 1995), and repeat this sequence eight times. We use this historical data set instead of a climatological data set to represent the interannual variability of the current climate.

[15] The soil texture data set used in the coupled model CCM3-IBIS and IBIS off-line is the IGBP-DIS global gridded texture database [*International Geosphere-Biosphere Programme (IGBP)*, 1999] interpolated at the T31 resolution. Both simulations are initialized with an "observed" potential vegetation map [*Ramankutty and Foley*, 1999] and run with a constant CO₂ concentration of 350 ppmv, characteristic of the mid-1980s. Because the soil carbon has a very long characteristic timescale, we



Figure 1. Schematic representation of the carbon cycle in IBIS.

accelerate the soil carbon module in both runs so that the soil carbon achieves 3500 years during the 100 first years of the run. In order to compare the results from the coupled model to the results of the off-line model, we average the last 60 years of each run.

4. Simulated Climate

[16] The vegetation distribution and the carbon balance simulated by the coupled model CCM3-IBIS depend strongly on the simulated climate. Therefore, we first compare the simulated climate to observations from the CRU05 data set. We then analyze the simulated vegetation structure and composition and the simulated carbon cycle by comparing results from the coupled model CCM3-IBIS with results from IBIS off-line and direct observations when available.

[17] Here, we focus on near surface air temperature and precipitation because they are the most important climate variables for the vegetation. The climate simulated by the coupled model CCM3-IBIS with a prescribed vegetation map is discussed in detail in an earlier paper [*Delire et al.*, 2002]. The climate simulated here is not identical to the simulated climate described by Delire et al. In the latter, the coupled model was run with a prescribed vegetation map so that the simulated climate did not affect the composition and structure of the vegetation. Here, the simulated climate affects vegetation structure and composition, which in turn influence the climate. The differences between the two simulated climates are therefore due to dynamic vegetation feedbacks. In the following discussion, we point out the

biases in the simulated climate and try to distinguish between purely atmospheric model causes and dynamic vegetation feedback causes.

[18] In boreal winter (December, January, February), the coupled model tends to underestimate surface temperature on the continents around the North Atlantic, and overestimate temperature both south of the Laurentian Lakes in Canada and in a narrow region stretching from the Caspian Sea to Northern Siberia (Figure 2a). This pattern of cold and warm bias in the Northern Hemisphere was also noted in studies where CCM3 was coupled to other land-surface models: LSM and a simple bucket type land-surface module [*Bonan*, 1998; *Delire et al.*, 2002]. Therefore, it is likely that the atmospheric model (or the sea-surface temperatures) is responsible for these biases. Temperatures are also too high in Argentina and along the Andes. The latter is mostly due to the flattened topography of the Andes in the model at low spatial resolution.

[19] In boreal summer (June, July, August), the coupled model simulates lower than observed temperatures in Alaska and Northern Siberia (Figure 2b). This cold bias is about 2°C stronger than in the work of Delire et al., indicating a feedback of the vegetation. The lower than observed temperatures simulated by the atmospheric model favors tundra at the expense of forests in these regions. Tundra has a higher albedo than boreal forest, especially in spring when snow is present, and absorbs less energy. Temperature is decreased in spring; melting is delayed, which results in colder summer conditions. Greenland, the United States and eastern Canada, the Andes, and a stretch of land North of the Mediterranean Sea from Spain to Lake a) Coupled model minus climatology, temperature, DJF

c) Coupled model minus climatology, precipitation (mm/day), DJF



b) Coupled model minus climatology, temperature, JJA

d) Coupled model minus climatology, precipitation (mm/day), JJA



Figure 2. Difference between temperatures simulated by the coupled model at the lowest GCM level (between surface and 65 m) and observed temperatures from the CRU05 climatology [*New et al.*, 2000] for (a) December, January, February, and (b) June, July, August. Difference between precipitation simulated by the coupled model and observed precipitation from the CRU05 climatology for (c) December, January, February, and (d) June, July, August. Shades of gray have temperature (precipitation) differences bigger than 1°C (1 mm/day). Contours are displayed every 4°C from -6° C for temperature and every 4 mm/day from -6 mm/day for precipitation.

Baikal are too warm. The lack of lakes, wetlands, and crops in CCM3-IBIS partly explains the warm bias in North America [*Delire et al.*, 2002]. The warm bias in the Andes is again due to the poor representation of the steep orography of the Andes in the fairly low-resolution coupled model. The high temperatures in Greenland are partly due to the snow module in CCM3-IBIS.

[20] GCMs have known biases in the precipitation fields, especially in the low latitudes, resulting from inaccurate parameterization of convection and low resolution. The coupled model CCM3-IBIS overestimates precipitation on the continents: 2.50 mm/day instead of 2.08 mm/day on average over all continents except Antarctica (the CRU05 data set does not cover Antarctica). To a lesser extent, this was also the case with the original land surface model LSM [Bonan, 1998] coupled to the CCM3. The largest errors occur in Africa and the Arabian Peninsula where the model fails to represent the location and intensity of the West African monsoon (Figure 2d). Instead of being located close to the Gulf of Guinea, the model simulates the African monsoon over Sudan, Ethiopia, and the Arabian Peninsula where it connects with the Indian monsoon. As a result, summers are too dry in West Africa and too wet in northeastern Sahel, and from the Arabian Peninsula to central India (JJA). This misrepresentation of the west-African monsoon was also obtained with CCM3 coupled to another land-surface model, LSM [*Bonan*, 1998] and is most likely related to the atmospheric model. The fairly low resolution used here accentuates the problem. The effect of vegetation dynamics on this bias is minor. In central and South Africa, maximum precipitation is too high during the wet season (DJF), and the dry season is too long.

[21] On the other continents, there are a variety of precipitation biases. The model overestimates precipitation over the Tibetan plateau all year round, which was also the case with LSM and is due to the atmospheric model. The model strongly underestimates precipitation in Indochina and southeast China during the wet season (JJA), while Northern China, Mongolia and southeast Siberia are too wet. The model tends to exaggerate the dry season in the Indonesian Archipelago. Seasonal precipitation reaches too far south into western Australia in austral summer. In the Amazon basin in South America, precipitation is underestimated during the wet season north of the equator (JJA), and the dry season is too long south of the equator. The Nordeste is too wet in the wet season. Southern Brazil and

							0,		
	Temperature								
Precipitation	-30°C/-20°C	$-20^{\circ}C/-10^{\circ}C$	$-10^{\circ}C/0^{\circ}C$	0°C/10°C	10°C/20°C	20°C/30°C	30°C/40°C	Total	
			CCM3-I	BIS					
0-2 mm/day	0	7.77	8.20	7.72	11.62	16.65	0.25	52.3	
2-4 mm/day	0	0.19	4.26	7.27	6.28	9.5	0.00	27.5	
4-6 mm/day	0	0.00	0.55	0.92	1.65	8.53	0.00	11.7	
6-8 mm/day	0	0.00	0.00	0.43	0.00	6.15	0.00	6.6	
8-10 mm/day	0	0.00	0.00	0.11	0.00	1.69	0.00	1.8	
10-12 mm/day	0	0.00	0.00	0.00	0.00	0.26	0.00	0.26	
Total	0	7.96	13.01	16.44	19.55	42.78	0.25	100.00	
			CRU0	5 ^a					
0-2 mm/day	0.56	6.55	12.13	14.71	10.22	17.96	0.00	62.13	
2-4 mm/day	0.12	0.29	1.46	5.61	5.65	12.73	0.00	23.76	
4-6 mm/day	0.00	0.00	0.00	1.16	1.16	7.41	0.00	8.94	
6-8 mm/day	0.00	0.00	0.00	0.00	0.00	3.74	0.00	3.74	
8-10 mm/day	0.00	0.00	0.00	0.00	0.00	1.30	0.00	1.30	
10-12 mm/day	0.00	0.00	0.00	0.00	0.00	0.13	0.00	0.13	
Total	0.68	6.84	13.59	18.59	17.03	43.27	0.00	100.00	

 Table 1. Percentage of Land Area That Falls Within a Certain Annual Temperature Range and Precipitation Range

 as Simulated by the Coupled Model CCM3-IBIS and as Observed From the CRU05 Climatology

^aCRU05 climatology data are from the work of *New et al.* [2000]. Antarctica is not included in the analysis because the CRU05 data set does not cover it.

the mouth of the Rio de la Plata are too dry all year round. In North America, Alaska, and the Rockies are too wet in winter and spring, central Canada from Lake Winnipeg too Hudson Bay is too dry in summer and fall, and the southeastern United States is too dry year round. Most of the biases in the precipitation fields were obtained with LSM coupled to the CCM3 and with CCM3-IBIS with fixed vegetation [*Delire et al.*, 2002]. They are most likely due to the GCM. Vegetation feedbacks play a role, but the amplitude of the feedback is smaller than the biases. In North America for instance, vegetation dynamics results in slightly drier conditions, reducing the wet bias in the West and enhancing the drought in the East.

[22] To analyze the simulated climate in a nonspatial way, we calculate the fraction of land area that falls within a certain annual temperature and precipitation range and compare it to observations from the CRU05 data set (Table 1). Early work in ecology [e.g., Holdridge, 1947] or on carbon cycling in vegetation and soils [Lieth and Whittaker, 1975; Post et al., 1982] showed that annual precipitation and annual temperature are key factors in determining the type of vegetation, net primary productivity (NPP) and soil respiration. Despite the important local biases in the simulated temperature and precipitation fields, the area of land experiencing a certain annual temperature and precipitation range is close to the observed. For instance, the coupled model simulates 9.5% of the land cover (minus Antarctica) with annual temperatures ranging from 20°C to 30°C and mean annual precipitation from 2 to 4 mm/day. According to the CRU05 data set, 12.7% of the total land cover (minus Antarctica) falls within this category. The coupled model does not necessarily simulate that given "climate" at the right location but does simulate it over an area comparable to the observations. GCMs are known to simulate effectively the general circulation of the atmosphere but not the exact placement of the particular atmospheric circulation patterns [Gates, 1995]. Here, while there are regional biases, CCM3 conserves the

land areas falling within a certain annual temperature and precipitation range.

[23] Differences in the temperature-precipitation distribution include, for instance, a higher simulated than observed fraction of land with temperature between -10 and 0°C and precipitation greater than 2 mm/day. This is also true for the areas with annual temperatures between 0 and 10°C. The Tibetan Plateau, where precipitation is overestimated, is mostly responsible for these differences. The GCM does not simulate climate regimes with annual temperatures below -20° C (except in Antarctica, which is not included here).

5. Simulated Vegetation Cover

[24] The performance of the off-line model in simulating the vegetation structure and composition and the carbon cycling in the biosphere was described in an earlier paper [Kucharik et al., 2000]. Here, we test the performance of the coupled model CCM3-IBIS by comparing its results primarily to the results of the off-line version of IBIS and secondarily to direct observations when available. There are several reasons for adopting this approach. First, very few global gridded observations of natural vegetation cover are available. Second, the model used in the off-line simulation is basically identical to the one coupled to the CCM3. By comparing both results, we can isolate biases due to the coupling to the climate model. Third, the results from the off-line model are necessarily better than the results of the coupled model because the off-line model is forced with observed climate data. The coupled model calculates its own climate and therefore compounds errors from both the land surface model and the climate model.

5.1. Leaf Area Index

[25] In IBIS, the vegetation cover is represented in terms of 12 plant functional types (pfts) competing for light and water. The maximum leaf area index (LAI) reached during the year



Figure 3. Distribution of potential leaf area index of evergreen trees, deciduous trees, and grasses and shrubs simulated by the off-line model forced with the CRU05 climate and simulated by the coupled model.

for each pft depends on the amount of carbon assimilated by this pft during the previous year. LAI in the model is thus linked to the carbon cycling and is therefore a good indicator of the behavior of the model. To ease the comparison between the coupled model and the off-line model, we grouped the 12 pfts into evergreen trees (tropical broadleaf evergreen, warmtemperate broadleaf evergreen, temperate conifer evergreen, and boreal conifer evergreen trees), deciduous trees (tropical broadleaf drought-deciduous, temperate broadleaf colddeciduous, boreal broadleaf cold-deciduous, and boreal conifer cold-deciduous trees), and grasses and shrubs (evergreen and cold-deciduous shrubs, C4 and C3 grasses). The total LAI for each group is simply the sum of the LAIs of the corresponding individual pfts.

[26] Except for a stretch of land southeast of Lake Chad, the coupled model simulates lower LAIs than the off-line version around the equator where evergreen tropical trees are simulated (Figures 3a, 3b, and 4). Deciduous trees gain from the competition with evergreen tropical trees in the Amazon basin, central Africa, and Indonesia (Figures 3c and 3d). This change in vegetation composition results from the underestimation of the precipitation during the dry season simulated by the GCM in the tropical regions. The LAI of evergreen trees (warm temperate) is increased in South Africa and from Saudi Arabia to northwest India at the expense of grasses and shrubs and deserts (Figures 3c-3f). Temperate and boreal evergreen trees have higher LAIs over the Tibetan plateau (at the expense of grasses and shrubs), in central Asia, Siberia, and in the central United States.

[27] Increased precipitation allows deciduous trees to colonize the Sahel and the south of the Arabian Peninsula. Higher spring to fall rainfall also favors temperate deciduous trees southeast of Lake Baikal. In the eastern United States and Canada, deciduous trees lose competition in favor of C4 grasses because the lower precipitation together with the higher summer temperatures result in lower available soil moisture. Cooler summers in Alaska and Siberia explain lower LAI of deciduous trees (and conifer) trees while warmer conditions West of Hudson Bay explain higher LAI for deciduous trees.

[28] In most of Africa, C4 grasses lose competition in favor of trees because of too wet conditions. Because of these wetter conditions, grasses are established further north in the Sahara. Wetter conditions also explain lower LAIs of



Figure 4. Distribution of potential vegetation types simulated by the off-line model forced with the CRU05 climate and simulated by the coupled model.

grasses (replaced by trees) in China while drier conditions explain the lower LAIs in eastern Australia.

5.2. Vegetation Types

[29] Vegetation type is only a diagnostic output in IBIS, but it allows easy comparison between model results. The vegetation type of a grid cell is calculated using simple rules based on the leaf area index (LAI) of the different plant functional types in that grid cell, and some overriding climatic rules. For example, a grid cell is classified as a temperate deciduous forest if temperate broadleaf cold deciduous trees have the highest LAI in that particular grid cell and if the total LAI of the upper canopy is greater than 1.5. If it is smaller than 1.5, the same grid cell is classified as a savanna. If the annual 5°C growing-degree-days (GDD5) is lower than 350, it is classified as tundra.

[30] The off-line model IBIS forced with climate observations represents well the major characteristics of today's potential vegetation, with tropical evergreen forests along the equator, surrounded by tropical deciduous forests and savannas [Kucharik et al., 2000]. Grasslands and shrublands mark the transition with the deserts around the Sahara or the Gobi desert. Grassland and shrublands are simulated in the western United States and Australia, South Africa, and South America. Temperate deciduous forests are simulated around the North Atlantic and the coast of China, boreal forests in Canada and north-central Eurasia, and tundra in the Arctic. However, the off-line model IBIS fails to represent extensive savannas because of poor representation of disturbance in the model [Foley et al., 1996; Kucharik et al., 2000]. Savannas are known to depend on disturbances as fire or extreme weather events, neither of which are well represented in the model [Botta and Foley, 2002]. Other discrepancies include too extensive temperate evergreen forests in Argentina and overly extensive tropical deciduous forests in Africa and South America, at the expense of savannas.

[31] The coupled model presents the same general vegetation characteristics as the off-line model, but some of the GCM biases in the precipitation and temperature fields together with vegetation climate feedbacks affect the simulated vegetation types. The most obvious differences are the forests simulated in the Arabian Peninsula and Iran and the boreal forest simulated on the Tibetan Plateau (Figure 4). The northern vegetation limit (polar desert/tundra) simulated by the coupled model CCM3-IBIS reaches higher latitudes northwest of Hudson Bay than with the off-line version but lower latitudes in northern Siberia because of the aforementioned warm bias in northern Canada and the cold bias everywhere else around the North Pole. The same pattern of warm and cold biases around the Arctic explains why the treeline is simulated further north in northern Canada, northwest of Hudson Bay, and is displaced further south in Eurasia and Alaska. In these regions, the model simulates spring-to-fall cooler than observed temperatures resulting in values of GDD5 (\sim 200) lower than the limit for existence of boreal trees in the model (>350). The warm bias in northwest Canada is more pronounced in spring than in the work of *Delire et al.* [2002] because the establishment of trees lowers the albedo of the surface, thereby increasing the available energy.

[32] The coupled model simulates a boreal forest over the Tibetan Plateau because of the overestimated precipitation rates. The transition between boreal and temperate forests is fairly well represented but the coupled model does not accurately represent the limit between boreal or temperate forests and grassland, shrubland, and desert. Grasslands expand too far east in the United States at the expense of

		Area (106 k	m ²)	Kappa Statistics		
Vegetation Classification	Coupled	Off-line	% Difference	Coupled versus Off-line	Degree of Agreement	
Tropical evergreen forest/woodland	17.6	15.1	17	0.62	good	
Tropical deciduous forest/woodland	12.7	13.0	-2	0.32	poor	
Temperate evergreen broadleaf forest/woodland	8.7	6.3	38	0.45	fair	
Temperate evergreen conifer forest/woodland	3.6	2.9	24	0.41	fair	
Temperate deciduous forest/woodland	12.6	10.9	16	0.56	good	
Boreal evergreen forest/woodland	12.3	12.0	3	0.34	poor	
Boreal deciduous forest/woodland	7.5	8.0	-6	0.40	fair	
Mixed forest/woodland	2.8	2.1	33	0.02	no	
Savanna	5.1	4.7	9	0.1	very poor	
Grassland steppe	12.9	20.5	-37	0.27	poor	
Dense shrubland	0.6	1.5	-60	0.	no	
Open shrubland	7.1	8.8	-19	0.17	very poor	
Tundra	7.9	8.8	-10	0.51	fair	
Desert	17.4	14.5	20	0.55	good	
Polar desert/rock/ice (without Antarctica)	17.2	16.9	2	0.74	very good	
Global average (without Antarctica)	145.9	145.9	0.0	0.42	fair	

Table 2. Area Covered by Each Vegetation Types as Simulated by the Coupled Model and the Off-line Model (Columns 2–3) and Kappa Statistics [*Monserud*, 1990] Measuring Spatial Agreement Between Simulated Vegetation Types by the Coupled Model and the Off-line Model (Column 4 Value and Column 5 Qualitative)

deciduous forest, and desert conditions are simulated around Lake Winnipeg as a downstream result of the poor representation of the orography of the Rockies. The grasslands in the eastern United States are due to the high summer temperatures simulated by the GCM reducing the water availability together with the lower precipitation (Figure 2d). On the other hand, higher precipitation rates result in temperate forests extending too far west in eastern China at the expense of grasslands. The GCM simulates well the Sahara and the Gobi deserts but simulates desert conditions in northeast Australia, north Burma, and Argentina.

[33] In the tropics, Africa is the least well represented by the coupled model. The overestimated precipitation rates result in too extensive tropical and temperate forests with very little grassland and savannas.

[34] We use the kappa statistic [*Monserud*, 1990] to evaluate the spatial agreement between the two vegetation distributions (Table 2). The most extreme environments, which are polar deserts, tropical evergreen forests, and deserts, have the best spatial agreements. Mixed forests, savannas, and shrublands have the poorest spatial agreement. The largest errors in simulated total areas covered by each vegetation type (Table 2) do not necessarily coincide with the poorest spatial agreement. For instance, there is very poor spatial agreement for savannas but their simulated areas are within 9% of each other. This relates to the similarity between simulated and observed land areas falling within a certain range of annual temperature and precipitation, as mentioned above.

6. Simulated Carbon Balance

[35] The carbon cycle in IBIS can be separated into two main components: the vegetation and the soils (Figure 1). In each grid cell, the vegetation cover is a combination of pfts, each characterized in terms of three biomass pools: leaves, stems, and fine roots. Those three biomass pools are the fundamental variables from which all the variables describing the vegetation are derived. For each plant functional type i, the rate of change in each biomass compartment j (leaf, stem or root) is given by

$$\frac{dCv_{i,j}}{dt} = a_{i,j}NPP_i - LF_{i,j} - D_{i,j}$$

where $a_{i,j}$ is the fraction of annual NPP allocated to each biomass compartment (leaf, wood, or root), LF_{ij} indicates litterfall from each compartment and D_{ij} the disturbance (fire, wind, ...). In each grid cell, the rate of change in total biomass Cv is simply the sum of the changes of each biomass compartment for each plant functional type, and the total fluxes are simply given by

$$\begin{split} NPP &= \sum_{i=1}^{npft} \sum_{j=1}^{3} a_{i,j} NPP_i \\ LF &= \sum_{i=1}^{npft} \sum_{j=1}^{3} LF_{i,j} \\ D &= \sum_{i=1}^{npft} \sum_{j=1}^{3} D_{i,j}. \end{split}$$

[36] Similarly, the evolution of the different soil carbon pools can be summarized as

$$\frac{dCs}{dt} = LF - HR - L$$

where Cs, the total carbon content in the soil is the sum of the different individual pools (microbial biomass, slow and recalcitrant carbon pools), HR represents heterotrophic respiration (microbial respiration) and L leaching of carbon through the soil column. As a whole, the carbon content in the biosphere changes according to

$$\frac{d(Cv+Cs)}{dt} = NEE - L$$

where *NEE*, the net ecosystem exchange represents the net flux of carbon between the biosphere and the atmosphere and is equal to NPP - HR - D. We chose to count as positive any

Biosphere as Simulated by the Coupled Model CCM3-IBIS and the Off-line Model IBIS						
	Off-line Mean	Off-line Std	Coupled Mean	Coupled Std		
Biomass, Gt C	634.11	1.63	621.10	1.12		
Net primary productivity, Gt C yr^{-1}	54.46	1.55	53.76	1.10		
Gross primary productivity, Gt C yr ⁻¹	112.95	2.28	108.86	1.50		
Litterfall, Gt C yr ⁻¹	49.09	0.71	47.57	0.35		
Disturbance, Gt C yr ⁻¹	6.4	0.02	6.21	0.01		
Soil carbon content, Gt C	1335.9	1.98	1516.30	1.23		
Heterotrophic respiration, Gt C yr^{-1}	48.73	0.68	47.26	0.30		
Soil CO2 flux, Gt C yr^{-1}	70.79	0.97	67.66	0.57		
Leaching, Gt C yr^{-1}	0.23	$4 imes 10^{-4}$	0.26	$2 imes 10^{-4}$		
Net ecosystem exchange, Gt C yr ⁻¹	0.27	1.1	0.30	1.1		

Table 3. Global Values of Carbon Pools and Fluxes in the

Biomass (kg C m⁻²), offline model



flux from the atmosphere to the biosphere. Equilibrium is reached when, averaged over several years, the carbon pools are constant, and NEE compensates exactly leaching.

[37] Global averages of carbon fluxes and pools are similar in the coupled model and in the off-line model (Table 3). With the exception of the NEE, the total soil carbon, and leaching, the fluxes and pools simulated by the two models are within 5% of each other, although the GCM has a slightly slower carbon cycle than the off-line model. Neither simulation is at equilibrium at the end of the run but the imbalances are very small: Biomass increases by 0.16 Gt C yr^{-1} (0.2%) and the total soil carbon content by 0.08 Gt C yr^{-1} (0.005%). We first discuss the geographical distribution of the carbon stocks in the vegetation and soils and then the carbon fluxes.

6.1. Biomass

[38] The distribution of living biomass is directly related to the distribution of trees and grasses, as trees accumulate more carbon than grasses. Therefore, the highest biomass values are found in the tropical and temperate forests, and the lowest are found in deserts, grasslands, and tundras. The coupled model overestimates biomass in comparison to the off-line model on a stretch of land from Lake Chad to the southeast of Lake Baykal because of higher spring to fall precipitation (Figure 5). Overestimated precipitation in comparison to observations also explains higher biomass values in South Africa, Argentina, northeast of Brazil, and in the Northern Plains of North America. Underestimated precipitation is responsible for the lower biomass in southeast China, Burma, West Africa, Uruguay, southeast of the United States (together with higher summer temperatures), and around Lake Winnipeg. Lower summer rainfall rates together with higher summer temperatures explain the lower biomass in the southeast United States and western Europe. The warm and cold summer bias around the Arctic explains higher values of biomass west of Hudson Bay and lower biomass in Alaska and Siberia.

6.2. Soil Carbon

[39] As expected, soil carbon content is highest in boreal region where degradation is slow and is lowest both in

Figure 5. Distribution of biomass (kg C m^{-2}), simulated by the off-line model and by the coupled model.

deserts where NPP is low and in tropical regions where degradation is fast (Figure 6). Simulated values range from 0 kg C m⁻² in deserts to a maximum of 51 kg C m⁻² in northern Canada with the coupled model and to a maximum of 44 kg C m⁻² in the off-line version. Observed values from the IGBP-DIS soil database [IGBP, 1999] range from 0 to 85 kg C m⁻². The coupled model simulates 20% more carbon in the soil than the off-line version. The coupled model simulates larger soil carbon contents in the Northern latitudes, on the Tibetan Plateau, in Africa, and from the Arabian Peninsula through central India. The higher values in the tropics, the Tibetan Plateau, and northern Canada are

Total Soil Carbon Content (kg C m⁻²), offline model



Figure 6. Distribution of soil carbon content (kg C m^{-2}) to a depth of 1 m, simulated by the off-line model and by the coupled model.



Figure 7. Soil carbon content (kg C m⁻²) averaged per vegetation type (±1 standard deviation) as observed [*IGBP*, 1999], simulated by the off-line model and by the coupled model. The standard deviations in the model results and observations represent variations within each vegetation type.

explained by a higher NPP not fully compensated by higher respiration rates. The higher values around the Arctic, which account for 60% of the difference in total soil carbon content, cannot be explained by the values of the carbon fluxes (NPP and heterotrophic respiration) over the last 60 years of the run. They partly result from the high (350 ppmv) atmospheric CO₂ concentration imposed for the length of the run and the acceleration technique used. Because of the acceleration of the soil carbon module, the soil carbon pool has actually experienced 3000 years with 350 ppmv instead of the 280 ppmv prevalent during the last 2000 years. The high CO₂ concentration is more likely to affect the arctic regions because the slow decomposition rates result in very long residence times. Lower soil carbon contents are simulated in eastern US and at the mouth of the Rio de la Plata.

[40] We compare biome averages and standard deviations of simulated and observed soil carbon from the IGBP-DIS data set (Figure 7). The averages and standard deviations for the observed data are obtained by assigning each grid cell to the corresponding off-line IBIS vegetation type for that grid cell. Soil carbon contents agree fairly well for most of the vegetation types. Mixed forests, open shrublands, tundra and polar deserts, and deserts present the largest discrepancies between models and observations. One possible explanation for the observed high values in the deserts might be the result of previous climate conditions (warmer and wetter conditions during the Holocene) not simulated with the model and could also be related to the small number of pedons available to construct the data set (six in Mali, two in Oman, and three in Australia). The high values simulated for the tundra are linked to the aforementioned acceleration technique and the 350 ppmv atmospheric CO_2 concentration imposed.

6.3. Net Primary Productivity

[41] Over the last 60 years of the run, the global annual average net primary productivity (NPP) estimated by the coupled model is 54.3 Gt C yr^{-1} with an interannual variability (estimated by the standard deviation) of 0.9 Gt C yr^{-1} . In the off-line version of the model, the 60-year average is 54.5 Gt C yr^{-1} with a standard deviation of 1.6 Gt C yr⁻¹. Both values are within 5% of each other and fall within the range of 44–66 Pg yr⁻¹ [*Cramer et al.*, 1999] obtained with other (off-line) models. Because we use fixed sea-surface temperatures, the interannual variability of temperature and precipitation simulated by the coupled model is smaller than the observed variability of the climate, explaining the reduced standard deviation in NPP. Locally, NPP varies from 0 in deserts to a maximum of 1.3 kg m⁻² yr⁻¹ in tropical forests. The differences in the geographical distribution of NPP simulated by the coupled model and the offline model (not shown) follows closely the biases in spring to fall precipitation except in the Arctic where the differences in NPP are related to the temperature biases.

[42] Total NPP as a function of latitude is at a maximum around the equator, and at a minimum around 20°N because of the Sahara desert (Figure 8). The coupled model simulates similar total NPPs, except for a lower maximum around the equator. This is due to slightly lower NPP of



Figure 8. Zonal totals of annual average net primary productivity, NPP (Gt C yr^{-1}), heterotrophic respiration, HR, and disturbance. The dotted lines are results from the off-line model and the solid lines are results from the coupled model.

the tropical evergreen forest caused by the exaggerated dry season and a higher minimum around 20°N explained by the greener Sahel and Arabian Peninsula. Cooler than observed summers explain the lower NPP at high northern latitudes.

[43] We compare the net primary productivity (aboveground and belowground) averaged per vegetation types

simulated by the coupled model and IBIS off-line with a compilation of field measurements (S.T. Gower unpublished data, 1999) (Figure 9). This data set combines data from Esser et al. [1997], Cannell [1982], and S.T. Gower (unpublished data, 1999). The 1882 points were assembled in the 15 IBIS vegetation types by Kucharik et al. [2000] to evaluate the off-line version of IBIS at a $1^{\circ} \times 1^{\circ}$ resolution. Belowground productivity is estimated using the Gower et al. [1999] ratio of belowground to aboveground NPP. For both the observations and the models, the error bars indicate the standard deviation among the different values of NPP for the same vegetation type. We do not expect model results to match exactly the observations. The model simulates a grid cell average of NPP while field studies are generally made on small plots. However, the model values need to fall within the range of measured values (shown by the error bars). The off-line IBIS and the coupled model CCM3-IBIS tend to underestimate tropical and temperate forest NPPs. Both models strongly underestimate grassland and shrubland NPP. There is much less scatter in the simulated values of NPP for each vegetation type than in the observations. This can be partly explained by the resolution used in the models: topography, soils and climate are homogenous over a 3.75° grid cell not representative of the spatial variability in reality.

6.4. Heterotrophic Respiration

[44] The global annual average heterotrophic respiration simulated by the coupled model over the last 60 years is 47.6 Gt C yr⁻¹, 71% of the total soil carbon flux (67.7 Gt C yr⁻¹). Root respiration accounts for 29% of the total soil carbon flux. The off-line model simulates slightly higher values of the heterotrophic respiration and the total soil carbon flux (48.7 and 70.8 Gt C yr⁻¹) coherent with the



Figure 9. Annual net primary productivity (NPP in kg C m⁻² yr⁻¹) averaged per vegetation type (±1 standard deviation) as observed (S.T. Gower, unpublished data, 1999), simulated by the off-line model and by the coupled model. The standard deviations in the model results and observations represent variations within each vegetation type.

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Figure 10. Average seasonal cycle of NPP (squares), heterotrophic respiration (circles), net ecosystem exchange (triangles) aggregated over 30° latitude bands. The dotted lines indicate results from the off-line model, and solid lines indicate results from the coupled model. The seasonal cycles are presented as anomalies from the annual means.

simulated higher productivity. Locally, total soil CO₂ fluxes vary from 0 to 1.55 kg C m⁻² yr⁻¹ in the coupled model (0 to 1.789 kg C m⁻² yr⁻¹ in the off-line model) and follow the general patterns of NPP.

[45] The latitudinal distribution of yearly heterotrophic respiration follows closely the patterns of NPP (Figure 8). This behavior is expected because both simulations are run toward equilibrium for which NPP - HR - D - L = 0 and because the remaining fluxes, disturbance (*D*) and leaching (*L*) are small.

6.5. Seasonality of the Carbon Fluxes

[46] Dargaville et al. [2002] tested the performance of the off-line IBIS and three other models in simulating the seasonal cycle of the carbon fluxes using an atmospheric transport model and observations of atmospheric CO₂. They showed that the off-line IBIS (and to a lesser extent the other models) tends to underestimate the amplitude of the seasonal cycle of the net ecosystem exchange in the Northern Hemisphere. The coupled model gives similar results as the off-line version in the northernmost latitudes $(60^{\circ}N-90^{\circ}N)$ and in the Southern Hemisphere between 0° and 30°S (Figures 10 and 11). Both models simulate similar seasonal cycles for NPP and heterotrophic respiration, which result in similar seasonality of NEE. In the northern midlatitudes, the coupled model simulates a stronger seasonal cycle than the off-line model due to a stronger seasonality in the NPP. NPP increases too fast in spring, reaches a maximum in May-June, and decreases too early (July). Too early and excessive growth in spring

in China and Siberia southeast of Lake Baikal due to wetter conditions explains the higher NPPs in spring. Excessive temperatures and limited water availability in summer in the eastern United States explain the early drop in NPP. In the low latitudes of the Northern Hemisphere $(0^{\circ}-30^{\circ}N)$, the coupled model underestimates the amplitude of the seasonal cycle of NEE. The misrepresentation of the West African monsoon connecting with the Indian monsoon is responsible for this discrepancy between the results of the two models. When the Arabian Peninsula, Ethiopia, Sudan, and Egypt are removed from the computation of total NEE, the seasonal cycle simulated by the GCM is within 10% of the seasonal cycle simulated by the off-line model.

7. Summary and Conclusions

[47] In this paper, we investigate how well a coupled biosphere-atmosphere model can simulate carbon cycling and the general functioning of the terrestrial biosphere. Therefore, we analyze the climate, vegetation cover and carbon cycle in the vegetation and soils simulated by a coupled atmosphere-biosphere model, CCM3-IBIS. The climate is compared to observations, while the vegetation cover and the carbon cycle are compared to an off-line version of IBIS forced with observed climatic variable. The comparison of the simulated climate to an earlier simulation with the coupled model where the vegetation is fixed [*Delire et al.*, 2002] shows that the biases in the simulated climate are mainly due to the atmospheric model. The



Figure 11. Average seasonal cycle of zonal mean (a) NPP, (b) heterotrophic respiration, and (c) net ecosystem exchange simulated by the off-line model (top panels), the coupled model (middle panels) and the difference between the coupled and the off-line model (bottom panels). The seasonal cycles are presented as anomalies from the annual means. The dotted (solid) contours indicate negative (positive) anomalies.

coupling with the vegetation dynamics enhances or reduces some of those biases, but the effect is small compared to the magnitude of the biases.

[48] As already mentioned, GCMs do not accurately simulate local and regional climates but their skill at simulating the large-scale distribution of pressure, temperature, circulation, and precipitation is high. This is clearly the case with CCM3-IBIS. The coupled model simulates a geographic distribution of the temperature and precipitation fields that presents important local biases like the high precipitation in the Arabian Peninsula, but the area of land experiencing certain ranges of annual temperature and precipitation are similar to observations. As a result, the simulated vegetation cover presents local anomalies like the deciduous forest in Saudi Arabia, but the total area covered by each vegetation type (except grasslands) simulated by the coupled model is similar to the area simulated by the off-line model. The geographic distribution of the carbon fluxes and pools presents important local differences but with the exception of the soil carbon content, the global averages are almost identical, indicating that the coupled model simulates a biosphere that functions, on the whole, in a manner similar to the off-line model forced with climatic observations.

[49] The zonal mean of the annual carbon fluxes is affected by the local biases in the simulated climate, but the differences do not exceed 0.5 Gt C year⁻¹, except between 10°N and 20°N (Sahel and Arabia) where they reach 1 Gt C yr ⁻¹. This means that the biases in the fluxes compensate each other zonally, except for the latitude band of the Sahel. The seasonality of the zonally averaged carbon fluxes is well represented in the coupled model in comparison to the off-line model, although there are major flaws between 0° and 30°N due to the misrepresentation of the African monsoon.

[50] Averaged per vegetation types, the carbon fluxes are very similar, indicating that the vegetation types simulated by the GCM function in the same way as the vegetation types simulated by the off-line model, even if their exact location is not always correct.

[51] The biases in the simulated regional climates do not affect the vegetation in the same way. Biases in winter temperatures, for example, play a less important role than biases during the growing season. On the other hand, the patterns of warm and cold biases in summer around the Arctic strongly influence the northern vegetation limit and the northern treeline. The warm summer bias together with the precipitation in the eastern United States affect vegetation cover and carbon fluxes and pools. This bias also affects the seasonality of the zonal mean flux by reducing the NPP too early in summer.

[52] The most important bias in the simulated precipitation field is the misrepresentation of the African monsoon extending to central India. It dramatically affects the vegetation cover in the Sahel, in the Arabian Peninsula, and from the Persian Gulf to India. It also strongly affects the seasonality of the zonal mean fluxes. The excessive precipitation over the Tibetan Plateau results in boreal forest instead of tundra, but this does not affect the zonal mean of the annual carbon fluxes and their seasonal cycle. On the other hand, the excess summer precipitation southeast of Lake Baykal affects strongly the seasonality of the carbon fluxes of that latitude band. The exaggerate dry season in the tropical forests is responsible for slightly lower NPPs and a higher percentage of tropical deciduous trees.

[53] In summary, despite some important biases in the simulated local climate and carbon fluxes and pools, the coupled model simulates a biosphere functioning as a whole in a similar way than the off-line model. Global values of carbon pools and fluxes are almost identical. Zonal averages and averages per vegetation types are very similar. The seasonality of the zonal carbon fluxes is also well reproduced, except for the $0^{\circ}-30^{\circ}N$ latitude band.

[54] In this study, the atmospheric CO_2 concentration was kept fixed to 350 ppmv, not allowing vegetation and soils to affect it. This limitation is justified by the absence of an interactive ocean model. CCM3-IBIS is now coupled to an ocean carbon cycling model in order to simulate the full carbon cycle and the effect of fossil fuel emissions on the functioning of the atmosphere-biosphereocean system.

[55] According to our study, coupled models can be used to explore the global, hemispheric and zonal-mean coupling of atmospheric and terrestrial carbon cycles. This is compatible with the level of observational detail now available from flask measurements and inverse modeling studies. Further regional detail (at particular regions) will require further improvements in the fidelity of climate simulations, as well as regional data, not yet available.

[56] This work also has some implications for the strategy of future coupled model development. Coupled models allow us to simulate the processes resulting from the interactions between the individual elements of the climate system (e.g., atmosphere and vegetation interactions). However, the results of the coupled model will often be farther away from the local reality than the individual component models forced by observations because of feedbacks between the components, which reinforce model biases. For example, forced by the observed climatology of the Arabian Peninsula, the off-line vegetation model simulates a desert. Similarly, when a desert is imposed in the peninsula, the atmospheric model simulates a limited monsoon over the region in fairly good agreement with the observations. However, when the vegetation and climate model are coupled, the vegetation responds to the limited monsoonal rain, resulting in a feedback to the atmosphere and more extensive rainfall over the peninsula, thus making the results of the coupled model further from reality than either of the models alone.

[57] Coupling models with each other is thus not likely to correct the biases of the individual models: The biases can be enhanced or damped by the coupling but in the latter case most likely not for the right reason. When coupling models with each other, modelers need to understand the origin of the individual biases in order to know what result they can trust in the coupled model. Following from these points, modelers should probably test the validity of a coupled model against the results of the individual components as well as against observations.

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