Importance of vegetation feedbacks in doubled-CO₂ climate experiments

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Abstract. The rising atmospheric concentration of carbon dioxide resulting from the burning of fossil fuels and deforestation is likely to provoke significant climate perturbations, while having far-reaching consequences for the terrestrial biosphere. Some plants could maintain the same intake of CO₂ for photosynthesis by reducing their stomatal openings, thus limiting the transpiration and providing a positive feedback to the projected surface warming. Other plants could benefit from the higher CO₂ level and the warmer climate to increase their productivity, which would on the contrary promote the transpiration. The relevance of these feedbacks has been investigated with the Météo-France atmospheric general circulation model. The model has been run at the T31 spectral truncation with 19 vertical levels and is forced with sea surface temperature and sea ice anomalies provided by a transient simulation performed with the Hadley Centre coupled ocean-atmosphere model. Besides a reference doubled-CO₂ experiment with no modification of the vegetation properties, two other experiments have been performed to explore the impact of changes in the physiology (stomatal resistance) and structure (leaf area index) of plants. Globally and annually averaged, the radiative impact of the CO₂ doubling leads to a 2°C surface warming and a 6% precipitation increase, in keeping with previous similar experiments. The vegetation feedbacks do not greatly modify the model response on the global scale. The increase in stomatal resistance does not systematically lead to higher near-surface temperatures due to changes in the soil wetness annual cycle and the atmospheric circulation. However, both physiological and structural vegetation feedbacks are evident on the regional scale. They are liable to modify the CO₂ impact on the hydrological cycle, as illustrated for the case of the European summertime climate and the Asian summer monsoon. The strong sensitivity of the climate in these areas emphasizes the large uncertainties of climate change predictions for some of the most populated regions of the world and argues for the need to include more interactive land surface processes in current generation climate models.

1. Introduction

Increases in anthropogenic greenhouse gases resulting from the burning of fossil fuels and deforestation have altered the composition of the atmosphere, resulting in an increase in the amount of heat energy trapped near the Earth’s surface. Numerical and observational studies suggest that this enhanced greenhouse effect is perhaps responsible for the recent increase seen in global surface temperatures. This warming is expected to accelerate during the next century, possibly provoking significant climate modifications especially in the hydrological cycle [Melillo et al., 1995]. The perturbations at the Earth’s surface may also have far-reaching consequences for the terrestrial biosphere. The rate of change in surface air temperature is likely to be the highest since the Younger Dryas episode (about 11,000 B.P.), and many plant species may not be able to disperse rapidly enough to track areas of suitable climate. Other species might be sustainable but constrained to some adaptations in order to face the new environmental conditions.

Carbon dioxide is the greenhouse gas that is most influenced by human activities. Besides the possible climatic perturbations, the rising concentration of CO₂ is also likely to exert a direct influence on the terrestrial ecosystems [Woodward, 1987]. One response of vegetation may be a widespread increase in stomatal resistance [Field et al., 1995]. In a CO₂-enriched atmosphere, plants could maintain the same intake of CO₂ for photosynthesis by reducing their stomatal openings. Such a response would increase the water-use efficiency and reduce the transpiration, which has the potential of causing significant modifications in climate on the regional and global scales. The main effect may be a positive feedback onto the surface warming due to the decrease in evaporation. However, the atmospheric response could be fairly complex since the
stomatal conductance will respond to changes both in the CO$_2$ concentration and in the surface climate.

Increased stomatal resistance may not be the only response of vegetation to elevated CO$_2$ levels. On longer timescales, CO$_2$ fertilization might also cause an increase in net primary productivity in areas with enough water and nutrients [Rosenberg, 1982]. This could lead to a larger leaf area index (LAI), which would, on the contrary, increase the transpiration and, possibly, counteract the stomatal response. As the stomates, the photosynthesis will be affected not only by the CO$_2$ concentration but also by the associated climate change. The visible and near-infrared measurements of the advanced very high resolution radiometer (AVHRR) suggest that the Northern Hemisphere middle- and high-latitude forests could have already experienced an increased growth between 1981 and 1991, which has been attributed to the warming trend recorded over that period [Myhnen et al., 1997]. Longer series of satellite observations are necessary to confirm this hypothesis. However, these preliminary results may represent a first glimpse of possible vegetation response to climate change on the global scale.

Until recently, vegetation feedbacks have been ignored in climate models. It was not until the mid-1990s that the consequences of increased stomatal resistance were first investigated in global simulations, both for present-day climate and doubled-CO$_2$ conditions. Pollard and Thompson [1995] doubled the minimum stomatal resistance for each vegetation type in the GENESIS climate model linked to the LSX land surface scheme. They obtained a decrease in evapotranspiration leading to a surface warming, especially in tropical and boreal forested areas. Henderson-Sellers et al. [1995] performed the same sensitivity experiment with the NCAR GCM and the BATS land surface scheme (for both single and double CO$_2$ concentration) and found similar results. They noted that the combined effect of increases in CO$_2$ and stomatal resistance was to increase soil moisture content over large regions of northern midlatitudes, whereas the CO$_2$ increase alone led to a drying of these areas. Sellers et al. [1996] investigated the same issue by using the SiB2 land surface scheme incorporating a coupled photosynthesis-conductance submodel. The global reduction in the canopy conductance was between 25 and 35%, suggesting that the previous studies might have somewhat exaggerated the stomatal effect. Nevertheless, this effect was found to amplify significantly the surface warming due to the radiative CO$_2$ impact over the tropical continents.

While more physical vegetation schemes have been included in general circulation models (GCMs), even more sophisticated off-line vegetation models have been developed [Woodward et al., 1995; Kergoat, 1998; C. Jones et al. (A coupled climate carbon cycle project, submitted to Journal of Geophysical Research, 1999)]. These models are driven by monthly mean solar radiation, surface temperature, and precipitation and are able to estimate the vegetation net primary productivity on the global scale. They may also include a CO$_2$ fertilization effect and are therefore likely to predict the sensitivity of plants to changes in both CO$_2$ concentration and surface climate. Betts et al. [1997] were the first to use such a model in conjunction with an atmospheric GCM in order to analyze both physiological and structural vegetation feedbacks in climate change simulations. They used the Hadley Centre GCM iteratively coupled to the Sheffield University vegetation model [Woodward et al., 1995]. They showed that changes in vegetation density due to increased productivity could partially offset the stomatal effect and that both physiological and structural feedbacks could provide significant regional-scale effects. More recently, the Hadley Centre developed its own vegetation model (J99) which was then coupled to the MOSES land surface scheme in the HadAM3 atmospheric GCM [Cox et al., 1999]. A comparison of the response to CO$_2$ doubling with the previous study of Betts et al. [1997] suggests that the vegetation feedbacks are very uncertain on the regional scale, depending on the simulated climate change and on the vegetation model. Levis et al. [1999] also used a fully coupled climate-vegetation model to examine the potential effects of vegetation feedbacks on simulations of CO$_2$-induced climate change and found an enhanced surface warming in the northern high latitudes during spring and summer, mainly due to changes in surface albedo.

In the present study, three 10-year doubled-CO$_2$ experiments are performed with the ARPEGE-Climat model as part of the LPSCR European project [Polcher et al., 1999]. The aim of the project is to assess the uncertainties in climate change predictions which are linked to our ability to model the land surface processes. In the first experiment we only consider the radiative impact of doubling the CO$_2$ concentration. The following simulations investigate the influence of various vegetation feedbacks. The ARPEGE-Climat model does not yet include a direct effect of carbon dioxide on plants. Before incorporating a more interactive vegetation scheme [Calvet et al., 1998], this study is a preliminary attempt to evaluate the relative importance of two possible feedbacks: the physiological feedback related to a change in the minimum stomatal resistance ($R_{\text{min}}$) and the structural feedback related to a change in the leaf area index (LAI). The separation of the two feedbacks may be justified by their different timescales, physiological responses being faster than structural ones [Rosenberg, 1982]. The strong uncertainties in the LAI response and the rather simple way it affects the transpiration of plants in the ARPEGE-Climat model is another reason for considering the feedbacks separately.

In section 2 the experiment design will be described. The global results of the three doubled-CO$_2$ experiments will be compared in section 3. Special attention will be devoted to the response at the Earth’s surface. Section 4 will focus on the regional impacts obtained over Europe and southern Asia, where the atmospheric response to the CO$_2$ doubling was found to be particularly sensitive to the vegetation feedbacks. Finally, conclusions about the importance of the physiological and structural vegetation feedbacks for regional climate change predictions will be presented in section 5.

2. Experiment Design

2.1. Atmospheric Model and Land Surface Scheme

Version 2 of the ARPEGE-Climat spectral model [Déqué et al., 1994] is used with a T31 triangular truncation (horizontal resolution of about 3.8°) and 19 vertical levels. The main novelty compared to version 1 is the introduction of an improved radiation scheme [Moretette, 1989, 1990]. For the shortwave radiation the two-stream formulation is employed together with the photon path distribution method in two spectral intervals. For the longwave radiation the scheme uses the broadband flux emissivity method with six intervals. Gas absorption is described explicitly, so that the carbon dioxide concentration is a control parameter of the scheme that can easily be modified.
At the Earth’s surface the ISBA (interactions between soil biosphere and atmosphere) scheme is used to provide a boundary condition to temperature and moisture [Noilhan and Planton, 1989; Manzi and Planton, 1994; Mahfouf et al., 1995; Douville and Royer, 1996]. In version 1 the ARPEGE-Climat model had only two temperature levels in the soil. The deep soil temperature was restored toward climatological values in order to avoid spurious drift at high latitudes during the polar night. This method was rather restrictive for climate change experiments, since it did not allow the atmospheric GCM to exhibit freely its own internal feedbacks. In version 2 the deep soil temperature relaxation has been replaced by a four-layer heat diffusion scheme [Déqué et al., 1994].

Despite containing the basic physics of land surface processes, the ISBA scheme is parsimonious and needs only a few parameters, depending on the types of soil and vegetation. The vegetation types are taken from the global classification of Wilson and Henderson-Sellers [1985] which was simplified by Manzi and Planton [1994]. Heat and water transfers in the ground are based on the force-restore method [Deardorf, 1978]. The treatment of the canopy layer has been simplified to avoid the numerical resolution of a specific foliage temperature. A single surface temperature is computed, which is representative of the whole soil-canopy system. Four reservoirs are used for moisture: the reservoir of rain intercepted by the canopy, the surface volumetric water content, the total volumetric water content, and the snow layer. A bottom runoff by gravitational drainage has been introduced in the prognostic equation for the deep water content [Mahfouf et al., 1995]. Finally, the treatment of snow-covered surfaces has been improved by adding snow density and snow albedo as prognostic variables, thereby allowing the description of snow aging processes [Douville et al., 1995].

Besides its influence on the thermal and radiative surface properties, the vegetation has two major effects. Firstly, rainfall and dew are intercepted by the foliage and evaporate in the air at a potential rate from the fraction \( \delta \) of the foliage covered with a film of water. Secondly, the remaining part of the leaves transpires at the rate

\[
E_{o} = \text{veg} \rho_{o} \frac{1 - \delta}{R_{s} + R_{s}'} (q_{s}(T_{s}) - q_{s}),
\]

where \( \text{veg} \) is the fraction of vegetation cover, \( \rho_{o} \) is the air density, \( T_{s} \) is the surface temperature, \( q_{s}(T_{s}) \) is the saturated specific humidity at temperature \( T_{s} \), and \( q_{s} \) is the surface air specific humidity. The surface resistance \( R_{s} \) adds to the aerodynamic resistance \( R_{s}' \) and is computed as

\[
R_{s} = \frac{R_{s}'}{\text{LAI}} F_{o},
\]

where the minimum stomatal resistance \( R_{s}'' \) and LAI are prescribed for each vegetation type. In keeping with the Jarvis-type approach [Jarvis, 1976], the factor \( F_{o} \) represents the additional dependence on the photosynthetically active radiation, the vapor pressure deficit of the atmosphere, the air temperature, and the soil water stress [Noilhan and Planton, 1989].

Equation (2) indicates that the surface conductance \( G_{s} = 1/R_{s} \) is proportional to the ratio \( \text{LAI}/R_{s}'' \). As explained in the introduction, many species of plants could reduce their stomatal openings in a \( \text{CO}_{2} \)-enriched atmosphere, which may be represented by an increase in \( R_{s}'' \). However, \( \text{CO}_{2} \) fertilization and climate change could also reinforce the productivity of plants and lead to increased LAI. The two responses have opposite impacts on the transpiration. In the control doubled-\( \text{CO}_{2} \) experiment, none of these effects are considered. The impact of the stomatal response alone is investigated in a second doubled-\( \text{CO}_{2} \) experiment. The combined \( R_{s}'' \) and LAI effects are analyzed in a third simulation. Note that some observations suggest that the surface conductance saturates at high values of LAI. This saturation is not considered in the ISBA scheme, and the two last experiments represent extreme cases in which the LAI response is either ignored or likely overestimated.

### 2.2. Description of Experiments

The present study is based on the commonly used “timeslice” technique, which avoids the use of an expensive coupled ocean-atmosphere model [Stephenson and Held, 1993]. The atmospheric GCM is forced by sea surface temperature (SST) and sea-ice anomalies extracted from a transient simulation made with a global coupled model. In the real climate system, the vegetation is fully interactive with both the ocean and the atmosphere. Here we assume that the terrestrial vegetation feedbacks have a limited impact on the global scale, so that the timeslice technique is sufficient for a preliminary assessment of their role in climate change simulations. The results will confirm that this hypothesis is reasonable and that the vegetation feedbacks are relevant on the regional scale but do not have a dramatic influence on the general circulation.

#### 2.2.1. Experiment CO1, present-day climate.

Before doubling the \( \text{CO}_{2} \) atmospheric concentration, a first 10-year simulation, CO1, was performed for present-day climate. The \( \text{CO}_{2} \) concentration is taken to be 353 ppmv, and the model is forced by observed monthly mean SSTs and sea-ice extents provided by the AMIP climatology for the 1979–1988 period [Gates, 1992]. The results of this control experiment have been briefly discussed by Douville et al. [1997], showing that the ARPEGE model is able to capture the main features of the observed climate. The most obvious defect is an overestimation of the air surface temperatures in the middle and high latitudes of the Northern Hemisphere. However, the large-scale circulation is generally well simulated, and the results are reasonable in the two regions of particular interest: southern Asia and Europe.

#### 2.2.2. Experiment CO2, \( \text{CO}_{2} \) doubling with no vegetation feedback.

Experiment CO2 only considers the radiative impact of \( \text{CO}_{2} \) doubling, without any modification of the vegetation properties. As in the following time-slice experiments, the SST and sea-ice anomalies were obtained from the “GHG” transient experiment performed with the HadCM2 coupled model of the Hadley Centre [Johns et al., 1997]. The reference period was chosen to be the years 1980 to 2000, the values for a climate with a doubled-\( \text{CO}_{2} \) concentration being obtained by averaging the years 2050 to 2070. Figure 1 shows the SST anomalies prescribed in January and July. The maximum warming appears in the tropical oceans as well as along the coasts in the northern midlatitudes. In the high latitudes the ocean warming is limited by the melting of sea ice.

The climate change simulated in experiment CO2 is in many aspects close to the projections of previous similar studies performed at Météo-France or elsewhere: greater surface warming of the land than of the sea, maximum surface warming in high northern latitudes in winter, little surface warming over the Arctic in summer, an enhanced global mean hydrological cycle, and increased precipitation and soil moisture in...
high latitudes in winter. In global and annual means, the surface air warming is 2°C, and the precipitation increase is about 6%. Generally speaking, the rainfall increases over the Intertropical Convergence Zone (ITCZ) and decreases in the subtropics, in keeping with the stronger Hadley circulation. In the extratropics the anomalies are positive, and this increase is accompanied by a poleward shift of the midlatitude precipitation belt. A more detailed analysis of experiment CO2 may be found in the work of Douville et al. [1997].

2.2.3. Experiment RO2, CO2 doubling with increased \( R_{\text{min}} \). Experiment RO2 is exactly the same as CO2 except for the modification of the minimum stomatal resistance \( R_{\text{min}} \). This modification was inferred from similar time-slice experiments (same SST and sea-ice anomalies) performed at the Hadley Centre with the HadAM3 atmospheric model incorporating the MOSES land surface scheme [Cox et al., 1999]. In MOSES the stomatal conductance is linked to the photosynthesis and is therefore directly dependent on the atmospheric CO2 level. The Hadley Centre performed two doubled-CO2 experiments, one in which the increased CO2 concentration is seen by both the atmosphere and the vegetation and another in which it is seen only by the atmosphere. Comparing the changes in surface conductance allowed them to isolate the direct CO2 effect from the indirect effect of climate change. As expected, the results indicate that the surface conductance decreases in a doubled-CO2 climate. It is due to the direct CO2 impact which dominates the influence, either positive or negative, of the other environmental parameters (radiation, temperature, humidity, and soil moisture) involved in the calculation of the stomatal conductance.

We are aware that the CO2 impact on stomatal conductance depends on several climate parameters and cannot be simply introduced in a Jarvis-type conductance model. Despite this strong limitation and the obvious differences in the climates
Figure 2. Schematic diagram of the experiment design.

simulated by the Météo-France and Hadley Centre models, the relative variations of the annual mean surface conductance between the two doubled-CO$_2$ experiments performed at Hadley Centre were applied on the maximum stomatal conductance prescribed in ARPEGE-Climat (Figure 2). Note that the ARPEGE-Climat model uses the same vegetation classification [Wilson and Henderson-Sellers, 1985] as the Hadley Centre model and that the prescribed changes in $R_{\text{min}}$, derived from the Hadley Centre experiments do not show a large spread when looking at all the land grid points corresponding to the same vegetation type (not shown). This result suggests that the stomatal response to CO$_2$ doubling is not too variable for a given vegetation type and that ignoring the interaction with the other climate parameters is a reasonable approximation on the annual timescale.

Globally averaged, the prescribed decrease in maximum stomatal conductance is about 34%, which is less than the reduction prescribed by Pollard and Thompson [1995] and Henderson-Sellers et al. [1996], and more consistent with the results of Sellers et al. [1996]. The corresponding increase in $R_{\text{min}}$ is generally between 50 and 100% except in the arid areas where the vegetation is anyway very sparse. Plate 1 shows the global distribution of the seasonal mean ratio LAI/$R_{\text{min}}$ in experiment RO2 for DJF (December-January-February) and JJA (June-July-August), as well as the difference between RO2 and CO2. Because of the strong annual cycle of the leaf area index in the middle and high latitudes the ratio LAI/$R_{\text{min}}$ is much higher in summer. The increase in $R_{\text{min}}$ introduced in experiment RO2, and thereby the decrease in LAI/$R_{\text{min}}$ is therefore more obvious in summer than in winter in the Northern Hemisphere midlatitudes. Other significant changes appear over China and in the tropics. On the other hand, the ratio is not greatly modified over the equatorial rain forests, which have already large values of LAI in present-day climate, so they cannot sustain much higher values due to the constraint on available light. The vegetation response to CO$_2$ doubling is therefore stopped as soon as this critical water stress is reached. Secondly, an increase in LAI also increases self-shading in vegetation canopy. It is assumed that LAI growth stops as soon as the annual carbon balance of the lowest canopy leaf layer becomes negative. Thirdly, an increase in LAI requires proportional carbon investments in the leaves, whereas the carbon gains (photosynthesis) saturates for high LAI. This again results in an upper limit for LAI, which also depends on the local environmental stresses affecting photosynthesis. The comparison of modeled LAI and water balance with satellite-derived LAI and river discharge data shows that the combination of the three criteria produces reasonable estimates for present-day climate (K99).

The relative variations of the maximum LAI predicted by the vegetation model were linearly interpolated onto the ARPEGE global grid and applied to the present-day monthly mean LAI used in the model. Globally averaged over land grid points, the relative increase is about 21%. As the decrease in the stomatal conductance is specified as in RO2 (34% in global mean), both effects are expected to partly cancel one another in experiment LO2. Plate 1 shows the spatial distribution of the differences in the ratio LAI/$R_{\text{min}}$ between experiments LO2 and RO2. The maximum differences appear in the Northern Hemisphere midlatitudes (including Europe), over China and West Africa, where the CO$_2$ fertilization effect is not inhibited by any unfavorable climate change. The differences found in the Northern Hemisphere midlatitudes are obviously more apparent in summer than in winter due to the annual cycle of the leaf area index. The differences are weak in the equatorial rain forests, which have already large values of LAI in present-day climate, so they cannot sustain much higher values due to the constraint on available light. The vegetation response to CO$_2$ doubling is described and analyzed in more detail by K99.

2.3. Statistical Testing of Results

In each of the three time-slice experiments the ARPEGE model has been integrated for 12 years. The first 2 years have been used for spin-up, and the climate anomalies have been averaged over the last 10 years, which is a minimum period necessary for detecting climate signals of weak magnitude. The regional atmospheric response to the CO$_2$ doubling can be difficult to detect, and the additive effects of the vegetation feedbacks can be even weaker. Although not presented in the present study, the one-tailed Student’s t-test [Chervin and Schneider, 1976] has been performed at all the grid points for many different fields, in order to assess the statistical significance of the climate anomalies. The main anomalies discussed in the rest of this article were found to be significantly different from zero at more than 90%. As it will be discussed in sections
3 and 4, further confidence in the vegetation impacts is suggested by the opposite responses obtained in experiments RO2 and LO2. More elaborate significance tests can be imagined but are outside the main focus of this investigation.

3. Global Response

3.1. Radiative Impact of CO₂ Doubling at the Land Surface

The control time-slice experiment CO2 gives an annual and global mean warming of 2°C. The spatial distribution of the DJF and JJA surface air temperature anomalies CO2-CO1 is described in Plate 2. In keeping with similar studies the maximum warming occurs in the high latitudes of the winter hemisphere. Also noticeable is the fact that the warming is clearly stronger over land than over sea. Although the SSTs used here were prescribed from the output of the HadCM2 coupled ocean-atmosphere model as discussed above, the greater warming over land is also seen in this model and others. The contrast is partly due to the fact that the heat capacity is much higher for the ocean than for the land surface. As a result, the land temperature responds more to anomalous heat fluxes from the atmosphere. The surface turbulent heat fluxes are also very different between land and sea. Some land areas are
Plate 2. Global distribution of DJF and JJA anomalies CO2-CO1; (left) surface air temperature (degrees), (right) total soil water content (mm).
Plate 3. Global distribution of JJA anomalies of surface air temperature (degrees).
fairly dry, so a large proportion of the radiative heating is used to warm the surface rather than to increase evaporation.

Plate 2 also shows the global distribution of the DJF and JJA soil water content anomalies generated by the ARPEGE-Climat model in response to CO2 doubling. In the middle and high latitudes the wetter winter conditions are a robust feature of the global warming already found in many GCM predictions. The summertime results are more uncertain, since the positive winter soil wetness anomalies may be offset by a larger spring and summer evaporation [e.g., Roads et al., 1996]. In the present study, only the strongest positive anomalies persist from winter to summer. They are mainly located in the central and eastern parts of the North American and Eurasian continents, while negative summer anomalies are found over western Europe and the west of the United States. In the tropics, experiment CO2 indicates a general drying, especially in JJA. The only exceptions are equatorial Africa, limited areas in Amazonia and Australia, Southeast Asia, and the southern part of the Indian peninsula.

A careful examination of Plate 2 indicates that there is a close relationship between the soil wetness anomalies and the predicted surface warming on the seasonal timescale. At a given latitude the maximum warming is found over areas where the soil moisture anomalies are negative, while positive soil moisture anomalies are associated with a weaker increase in surface temperatures. This result illustrates the influence of the surface hydrology in determining the surface warming due to CO2 doubling and suggests that the predicted temperature anomalies are strongly dependent on the treatment of the land surface. This is not a new result since the relevance of soil moisture for climate change numerical experiments was already emphasised by Meehl and Washington [1988] at the National Center for Atmospheric Research (NCAR). Although they used nearly the same low-resolution model as Manabe and Wetherald [1987] at Geophysical Fluid Dynamics Laboratory (GFDL), they did not obtain the same summer drying in the interior of the Northern Hemisphere continents in a doubled-CO2 climate. They found that their lack of the summer drying follows from their soil moisture values being much lower than those of Manabe and Wetherald [1987] in the control simulation. They concluded that the soil moisture amount, particularly during spring, is the critical factor determining the magnitude of the summer anomalies.

The present study confirms that soil moisture is a critical factor for determining climate change at the land surface and suggests that the regional patterns of climate change remain uncertain as long as soil, snow, and vegetation are not satisfactorily described in the GCMs.

### 3.2. Influence of Vegetation Feedbacks

The global warming obtained in experiment CO2 is only slightly modified on the global scale in experiments RO2 and LO2 (see Table 1). Several negative feedbacks limit the direct and global impact of the vegetation changes. The response of the transpiration of plants is partly offset by an opposite response of the bare ground evaporation in relatively wet areas (not shown). Moreover, as it will be further discussed, the local response of evaportranspiration can be counterbalanced by changes in the large-scale circulation and thereby in the hydrological cycle.

However, the vegetation feedbacks have a significant impact on the regional scale. Plate 3 compares the spatial distribution of the JJA surface air temperature anomalies between the various experiments. The difference RO2-CO2 shows that the surface warming is strengthened by the increase in $R_{s,m}$ over the equator and the vegetated areas of the summer hemisphere, where the net radiation is strong and the surface energy budget is sensitive to a reduction in the local transpiration. Experiment LO2 includes the additive effect of the modified LAI. As expected, the anomalies LO2-RO2 show roughly an opposite pattern to the anomalies RO2-CO2 over land. In other words, the net effect of the combined $R_{s,m}$ and LAI vegetation feedbacks is rather small. The surface warming is slightly stronger in LO2 than in CO2 over the equatorial and Northern Hemisphere midlatitude land regions, in keeping with the magnitude of the changes in LAI/$R_{s,m}$ shown in Plate 1. Experiment RO2 shows a significant and unexpected cooling relative to experiment CO2 over India and a north-south dipole over Europe, which are also partly offset by the modified LAI in experiment LO2. As it will be discussed in section 4, these regional impacts cannot be explained by a direct response to the modification of the local vegetation properties and are probably related to changes in the atmospheric circulation.

Figure 3 compares the JJA sea level pressure anomalies. Patterns of change in sea level pressure in doubled-CO2 experiments are known to be strongly model dependent [Mellilo et al., 1995]. They are not completely determined by the direct effect of surface temperature anomalies, even if reductions in pressure may be expected over the areas of strong surface warming (heat lows). Experiment CO2 shows a decrease in sea level pressure in the polar regions and an increase in the midlatitudes, in keeping with the latitudinal distribution of the global warming. Additional anomalies are found in experiment RO2, especially over Eurasia where the sea level pressure is reduced except in the very high latitudes. The reverse pattern appears in LO2-RO2, with anomalies of similar magnitude but opposite sign. As for temperature, the net effect of the physiological and structural vegetation feedbacks is rather weak.

Less expected is the relative antisymmetry between the CO2-CO1 and RO2-CO2 patterns (especially over Eurasia, but also over nonvegetated areas such as North Atlantic or Antarctica). This result can be explained by the fact that the impact of the increase in $R_{s,m}$ is mainly found in areas where the soil moisture stress is weak. As discussed in section 3.1, there is a strong relationship between the soil moisture and the temperature anomalies in experiment CO2 over the Northern Hemisphere continents in summer, with positive soil moisture anomalies associated with a weak warming and vice versa (Plate 2). The increase in $R_{s,m}$ has a specific impact over the wet regions where the warming was moderate in CO2 and becomes stronger in RO2. This is particularly obvious over the boreal forests of the Eurasian continent. This differential warming may contribute to a contrasted response of the plan-

<table>
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<th>Field</th>
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Figure 3. Global distribution of JJA anomalies of sea level pressure (hPa).
etary-scale circulation, which appears in the distribution of the sea level pressure anomalies.

Plate 4 compares the global distribution of the JJA mean anomalies of total precipitation among the various experiments. In keeping with the results of similar numerical studies, experiment CO2 leads to positive rainfall anomalies over the ITCZ and negative anomalies in the subtropics, due to a stronger Hadley circulation, as well as to a strengthening and poleward shift of the midlatitude precipitation belt in the winter hemisphere. The anomalies associated with the vegetation feedbacks often have a low magnitude, except over some limited areas such as northern India and some parts of Europe (compare section 4). Once again, the impact of the increase in $R_{\text{min}}$ is partly offset by the impact of the increase in LAI, so the LO2 climate is intermediate between the CO2 and the RO2 climates.

As for sea level pressure, there is an antisymmetry between RO2-CO2 and CO2-CO1, despite the expected reinforcement of the surface warming due to the increase in $R_{\text{min}}$. As previously discussed, the extra surface warming in RO2 is preferentially induced over wet and vegetated areas, where the temperature anomalies in CO2 may be fairly weak. Furthermore, the transpiration is favored by the CO2 radiative effect, while it is inhibited by the increase in $R_{\text{min}}$ if the soil moisture and the other environmental control parameters are not greatly modified. This negative feedback may dominate the positive feedback on the surface temperature. The presence of opposite patterns, not only over land but also over oceanic regions (for example, over the Atlantic), shows that the anomalies are related to changes in the atmospheric circulation rather than to local precipitation-evaporation feedbacks. This result confirms the influence that the land surface evaporation, and therefore the vegetation, can exert on the large-scale circulation.

The temperature, sea level pressure, and precipitation anomalies were only shown for boreal summer. Yet, other seasons lead to similar conclusions. The regional impacts of the vegetation feedbacks appear in all seasons and are not only found in the summer hemisphere (where the surface evaporation can be strong) but also in the winter hemisphere due to the remote influence of the regional perturbations in the surface hydrology. In all seasons and for each variable, the $R_{\text{min}}$ feedback in experiment RO2 is partly offset by the LAI feedback in experiment LO2. This result suggests that the atmospheric response to the modification of the vegetation properties is statistically robust. It does not mean that the vegetation feedbacks should be ignored but that each individual feedback is liable to modulate the CO2 radiative impact and should be assessed more carefully in order to obtain more reliable predictions of climate change on the regional scale.

### 4. Regional Responses

This section focuses on two specific regions, that are strongly influenced by both the CO2 radiative effect and the vegetation feedbacks. We recognize that the experiments have been performed at a rather low resolution and that we cannot claim to reproduce the regional climates with high accuracy, but we believe that the simulation of present-day climate is realistic enough for investigating the CO2 impact on the continental scale. Moreover, the main objective is not to precisely predict what will be the regional CO2 impact but to investigate the uncertainties in such predictions. For the regions to be considered, Europe and southern Asia, there is still little consensus about the possible climatic conditions that these large populations may face in the 21st century. Other regions (North America, Amazonia, and many other forested areas) also show significant impacts of the vegetation feedbacks, but Europe and southern Asia have been selected because they illustrate both the high sensitivity and the complexity of the climate response, which is not only related to the direct effects of the vegetation changes but also to the indirect effects, i.e., the changes in the atmospheric circulation. As for the global response, the focus is on boreal summer (JJA). After describing the regional impacts of the vegetation feedbacks over Europe and southern Asia, the local versus remote origin of these impacts will be briefly discussed, and the response of the Asian monsoon will be further investigated by analyzing changes in the moist static energy distribution.

#### 4.1. Europe

The radiative CO2 impact (experiment CO2) obtained in summer over Europe, described by Douville et al. [1997], can be summarized as follows: The maximum surface warming appears over southern Europe, because of the significant drying of the soil. The precipitation anomalies are negative over western Europe but positive in the northeast (Figure 4). The vegetation feedbacks considered in experiments RO2 and LO2 generate additional anomalies to the previous ones, which are locally of similar magnitude. As for the global anomaly patterns, there is an obvious antisymmetry between RO2-CO2 and CO2-CO1 on one hand and LO2-RO2 and RO2-CO2 on the other hand.

One of the main concerns for the Northern Hemisphere midlatitudes and especially for the European continent is the possible occurrence of more severe summer droughts in a doubled-CO2 world. Such a risk actually appears over western Europe in experiment CO2, although the anomalies remain rather small compared to the total soil water content (Figure 5). This negative anomaly is offset by the increase in $R_{\text{min}}$ prescribed in RO2 and is partly regenerated by the increase in LAI added in LO2. It is therefore very difficult to make predictions of the CO2 impact on the European water resources in summer, since a small perturbation, such as those introduced in the vegetation properties is liable to reverse the response given by the ARPEGE-Climat model.

Figure 6 shows the annual cycle of the monthly mean anomalies relative to the control experiment CO1 averaged over western Europe (36°N-55°N/10°E-20°W) for the canopy conductance and various features of the hydrological cycle. The negative anomalies of canopy conductance in experiments RO2 and LO2 exhibit some seasonal variations with a quick decrease in October at the end of the Northern Hemisphere growing season. In experiment CO2, soil moisture anomalies are positive from November to May and negative from June to October. The increase in the winter precipitation is offset by an increase in the runoff and in the evaporation, so the soil water budget shows a deficit in summer. This deficit is sustained by, but may be also partly responsible for, a reduction in the summer precipitation. In experiment RO2 no soil moisture deficit appears in summer. Despite the increase in $R_{\text{min}}$, the relative evaporation anomalies are often greater than in CO2, which may be interpreted as a consequence of the wetter surface conditions, especially in summer. From February to October the relative precipitation anomalies are greater in RO2.
than in CO2, which is due to a stronger atmospheric moisture convergence and, to lesser extent, to the stronger surface evaporation in summer. The anomalies LO2-CO1 are generally intermediate between the anomalies CO2-CO1 and RO2-CO1, but LO2 is closer to CO2 than to RO2.

Figure 6 suggests that the role of soil moisture and surface evaporation is not straightforward. The precipitation anomalies cannot be interpreted only in terms of water recycling. There is probably a more relevant feedback whereby the precipitation efficiency is governed by the state of the soil. In experiment RO2 the increase in $R_{\text{min}}^s$ allows more water to be stored in the soil, so a larger fraction of the atmospheric moisture that transits over Europe contributes to rainfall. The relevance of this mechanism for summertime rainfall over Europe was shown by Schär et al. [1999], who suggested that an increase in moisture convergence could be the consequence
Plate 4. Global distribution of JJA anomalies of total precipitation (mm/d).

more than the reason for increased precipitation, which could be attributed to wetter surface conditions.

4.2. Asian Summer Monsoon

The impact of greenhouse gases on the Asian summer monsoon is still a matter of debate [Zhao and Kellogg, 1988; Melillo et al., 1995; Royer et al., 1996; Kitoh, 1997]. Comparing the response to CO₂ doubling of five atmospheric models coupled to a slab ocean model, Zhao and Kellogg [1988] suggested that wetter summer conditions were likely to occur over both India and Southeast Asia. However, the results remained uncertain since three models indicated an increase in soil moisture, one produced a strong decrease, and the last one showed both positive and negative anomalies. Moreover, the Asian mon-

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**Figure 6.** Annual cycle of the anomalies over western Europe (36°N-55°N/10°W-20°E) for canopy conductance (mm/s), total soil moisture (mm), total precipitation (mm/d), total runoff (mm/d), total evaporation (mm/d), and precipitation minus evaporation (mm/d); months are displayed from January to December on the abscissa.
soon behavior can be rather complex, with contrasted responses of the Indian and Chinese subcontinents (see, for instance, Kitoh [1997]). In the present study we concentrate on the Indian peninsula, which is the most sensitive to the vegetation feedbacks, and where no evidence has been found in the observational record to suggest that the summer monsoon rainfall has increased systematically over the last decades [Pant et al., 1993].

The response of the Indian summer monsoon in experiment CO2 was examined by Douville et al. [1997]. To summarize, the surface warming over the Indian peninsula exhibits a marked latitudinal gradient, with stronger warming in the north than in
the south. This is true both in winter and in summer, but the summer warming is stronger and is associated with a decrease in the Indian monsoon precipitation except over the southern tip of the peninsula (Figure 7). These precipitation changes lead to significant anomalies in the total soil water content (Figure 8). The weakening of the Indian monsoon is noticeable not only in the precipitation field but also in the large-scale circulation, as indicated by a westward anomaly in the 850-hPa wind field over the Arabian Sea (not shown). The dynamical response of the Asian monsoon to CO₂ doubling will be analyzed in more depth in a forthcoming study.

Douvillé et al. [1997] suggested that it is not only the temperature anomalies but also the changes in surface humidity which may be relevant for understanding the CO₂ impact on the monsoon. Figure 9 shows the annual cycle of the soil moisture anomalies averaged over northern India (15°N-30°N/68°W-92°W). Before the onset of the monsoon, the Indian peninsula is very dry and is drier in CO₂ than in CO1. This drying could be responsible for the weakening of the monsoon convective activity in June and July, as shown by the annual cycle of the rainfall anomalies. This hypothesis is confirmed by the results of experiment RO2. The reduction of the canopy conductance causes more water to be stored in the soil during the dry season, so a positive soil water content anomaly appears over northern India before the onset of the monsoon (Figure 9). This moistening favors the convection over the continent and offsets the CO₂ radiative impact on the summer precipitation (Figure 7). In experiment LO2 the canopy conductance is intermediate between the values prescribed in CO2 and RO2. As over Europe, the total soil moisture finds a new equilibrium, which also lies between the results of CO2 and RO2.

4.3. Remote Versus Local Effects

It has been demonstrated that the vegetation feedbacks are liable to modulate the response of the ARPEGE atmospheric model to CO₂ doubling and even to reverse the regional hydrological response over Europe and southern Asia in summer. These sensitive regions correspond to areas where the perturbation of the vegetation properties is quite strong compared to many other parts of the globe (compare Plate 1). It is therefore tempting to assume that these anomalies have a local rather than a remote origin, even if this origin is not straightforward due to the competitive effects of soil moisture and canopy resistance on the surface evaporation (sections 4.1 and 4.2). However, this local origin is difficult to demonstrate without performing further experiments in which the vegetation properties would be modified only over limited areas, and the possible impact of global teleconnections is not excluded.

In the tropics and subtropics the role of vegetation in sustaining large-scale atmospheric circulations has been already discussed. Charney [1975] was the first to emphasize the relevance of vegetation feedbacks in the global climate system, in order to explain the dynamics of deserts in the Sahel. More recently, Eltahir [1996] analyzed the atmospheric response to tropical deforestation and proposed a simple mechanism for describing how the tropical forests influence the circulation. Eltahir [1996] assumed that a moist tropical atmosphere satisfies a quasi-equilibrium between moist convection and radiative forcing, so the temperature profile is related to the boundary layer entropy. A deforestation reduces the boundary layer entropy, cools the upper troposphere, which leads to a subsidence and weakens the low-level convergence. In keeping with our own hypothesis, this mechanism shows that it is necessary to consider both surface temperature and humidity in order to understand the influence of the land surface on the large-scale circulation (via changes in the boundary layer entropy).

Figure 10 shows JJA anomalies in the vertical gradient of moist static energy, \( c_p T + L q + g Z \), between 850 and 200 hPa. This gradient is related to the gradient of entropy and is a simplified version of the gross moist stability used by Neelin and Held [1987] to model the tropical convergence. In a warmer climate (difference CO2-CO1) the moist static energy is expected to increase in the whole troposphere due to the warming and the moistening of the air. However, the difference between 850 and 200 hPa shows a large negative anomaly centered over northern India. Despite the stronger surface warming simulated over land than over sea, the vertical gradient of moist static energy has been reduced over India, in keeping with the weaker convective activity and the lower precipitation. This indicates that the moist static energy anomalies are mainly driven by changes in specific humidity. The low-level specific humidity increases much more over the ocean, in response to the SST anomalies, than over the Indian continent, which is fairly dry before the onset of the monsoon. Therefore the combined effect of the large SST increase over the Indian Ocean and the low humidity over the Indian peninsula before the monsoon might be responsible for the monsoon’s response in experiment CO2. This result is consistent with the conclusions of Neelin and Held [1987], stating that warmer SSTs create mean upward motion by increasing the low-level moisture and thereby increasing the instability of the flow to moist convection and that precipitation in the favored regions is enhanced by convergent moisture transport at the expense of the less favored regions.

In the difference between RO2 and CO2 the vertical gradient of moist static energy shows a positive anomaly over India, in keeping with the increased monsoon precipitation in this area. A strong positive anomaly also appears over the temperate and boreal forests of the Eurasian continent, corresponding to the surface warming related to the increased canopy resistance. The difference between LO2 and RO2 shows an antisymmetric pattern in the middle and high latitudes. The negative anomaly over northern India is rather weak, despite the clear negative anomaly in the monsoon precipitation. This result suggests that the sensitivity of the tropical atmosphere to the vegetation feedbacks could be partly explained by the surface anomalies found in the higher latitudes. This hypothesis is consistent with the results of Douville and Roey [1996] concerning the consequences of a temperate and boreal deforestation on the Northern Hemisphere climate. A more detailed analysis of the dynamics would be necessary to identify the mechanism which could be responsible for the teleconnection between the extratropics and the tropics.

Figure 11 shows the stationary wave (eddy) component of the geopotential response over the Eurasian region. The stationary wave component is obtained by subtracting out the zonal mean geopotential height response from the total geopotential response. The CO2-CO1 response consists of a positive (anticyclonic) anomaly over western Europe and the North Atlantic and a negative (cyclonic) anomaly over western Siberia. These large-scale anticyclonic and cyclonic conditions are consistent with the decreased rainfall seen over western Europe and increased rainfall seen over western Siberia in the global distribution of the JJA precipitation anomalies (Plate 4). The CO2-CO1 response also shows a positive geopotential
eddy anomaly over northern India, as seen in the sea level pressure response in Figure 3, corresponding to a weakened monsoon trough associated with a weaker Indian monsoon.

The RO2-CO2 response has an anticyclonic anomaly over western Siberia, which acts to reduce the cyclonic response seen in CO2-CO1. Over India there is a negative cyclonic anomaly, corresponding to the monsoon being stronger in RO2 than in CO2. To the northwest of this Indian anomaly a pair of alternating sign anomalies can be discerned heading toward Europe. A similar feature is also present in the LO2-RO2 response and may be a Rossby wave response to changes in the Indian monsoon circulation. Such waves have been identified by Rodwell and Hoskins [1996]. They are generated by the diabatic heating in the Asian monsoon region, propagate west-

Figure 9. Annual cycle of the anomalies over northern India (15°N-30°N/68°E-92°E) for canopy conductance (mm/s), total soil moisture (mm), total precipitation (mm/d), total runoff (mm/d), total evaporation (mm/d), and precipitation minus evaporation (mm/d); months are displayed from January to December on the abscissa.
Figure 10. Vertical gradient of moist static energy (kJ/kg) between 850 and 200 hPa. JJA anomalies over the Eurasian continent.
ward, and interact with air on the southern flank of the mid-latitude westerlies causing it to descend over the Mediterranean region. If the patterns found in the eddy component of the geopotential anomalies actually correspond to a Rossby wave response, some of the RO2-CO2 and LO2-RO2 dynamical changes over Europe may be partly caused by the nonlocal response to changes in the Indian monsoon.

5. Conclusions

Despite the increasing efforts of the scientific community, there is little consensus about the regional details of greenhouse gas-induced hydrometeorological change. Even small perturbations in regional hydrology and water resources may, however, have significant implications for agriculture, water storage and distribution, and for the generation of hydroelectric power. The large spread in the numerical projections is partly related to uncertainties in the representation of climate processes, especially feedbacks associated with oceans, sea ice, and clouds. Besides these major feedbacks the terrestrial vegetation plays also a significant role which must be better understood in order to become more confident in the regional patterns of climate change. Vegetation is a full component of the carbon cycle and the climate system and is therefore likely to respond both to an increase in the atmospheric CO2 concentration and to the induced climate change. The present study is a preliminary work to address this issue in a simple way, with the ARPEGE-Climat atmospheric model coupled to the ISBA land surface scheme. Three doubled-CO2 time-slice experiments have been compared, using the same SST and sea ice anomalies. Besides a control doubled-CO2 experiment (CO2) with no modification of the vegetation properties, two other experiments have been performed in order to investigate the impact of changes in the physiology (RO2) and structure (LO2) of plants.

On the global scale, the vegetation feedbacks only slightly modify the atmospheric response to CO2 doubling, which is in many aspects close to the projections of previous similar studies performed at Météo-France and elsewhere; namely, greater surface warming over land than over oceans, maximum surface warming in high latitudes in winter, little surface warming over the Arctic in boreal summer, an enhanced global mean hydrological cycle, and increased precipitation and soil moisture in high latitudes in winter. On the regional scale, the vegetation feedbacks are much more significant and may exert an influence as strong as the radiative CO2 impact on the surface hydrology. However, the physiological feedback (increase in $R_{\text{min}}$) is partly counteracted by the structural feedback (increase in LAI) since they have opposite effects on the surface evaporation. This cancellation is clearly seen in the seasonal anomaly patterns of surface temperature, sea level pressure, and precipitation. The predicted climate change due to the radiative CO2 impact (experiment CO2) is not greatly modified, even on the regional scale, when both vegetation feedbacks are considered (experiment LO2). This result does not mean that the vegetation feedbacks may be ignored but that their impact is still uncertain due to the poor knowledge of the vegetation response in a doubled-CO2 world. Experiments RO2 and LO2 represent extreme scenarios in which the LAI response is either ignored or probably overestimated due to the significant fertilization effect (±25%) considered in experiment LO2 and to the linear dependence of the canopy conductance on LAI in the ISBA land surface scheme (no saturation for high values of the leaf area index). Note also that the physiological and structural feedbacks will probably take place on different timescales. While the physiological response should be fast, the structural changes will probably occur on longer timescales depending on how fast the plants are able to adapt to the new environmental conditions. On even longer
term, other vegetation responses could appear, such as changes in the vegetation height and in the geographic distribution of many species, which have not been considered in the present study.

Special attention has been paid to the European and Asian summer climates, whose response to CO₂ doubling is significantly modified by vegetation feedbacks. This result shows the high sensitivity of these regional climates, which was also pointed out by Mitchell and Johns [1997] in the Hadley Centre coupled ocean-atmosphere model. Experiment CO₂, looking only at the radiative CO₂ impact, predicts a summer drying over western Europe and a weakening of the Indian monsoon; the increase in Rs, min considered in experiment RO2 is large enough to suppress these significant impacts on the regional hydrological cycle. As found on the global scale, the Rs, min feedback is partly offset by the LAI feedback in experiment LO2, which is therefore intermediate between CO2 and RO2. This result helps confirm that the atmospheric response to the prescribed perturbations of the vegetation properties is statistically robust. The Rs, min feedback slightly dominates the LAI feedback in our experiments. However, we do not claim that the physiological response of the vegetation is more important than its structural response, or that both effects will actually cancel one another. We just conclude that each individual feedback is liable to play a significant role, which is still uncertain due to the poor knowledge of the vegetation response in a doubled-CO₂ world.

There are obviously several limitations in the present study. Firstly, changes in only two vegetation parameters, Rs, min and LAI, have been considered, so possible changes in roughness length or rooting depth have been ignored. It is, however, very difficult to predict such changes on the global scale since there is little observational evidence of such responses to increased CO₂ level. One solution would be to use a fully coupled climate-vegetation model simulating the spatial distribution of the vegetation, so there would be no more need for prescribing the vegetation map in the GCM. Recent studies have suggested that changes in the forest distribution are able to induce significant feedbacks on long-term climate evolution and could have, for example, contributed to the initiation of the last glacial [De Noblet et al., 1996; Kabatzi and Claussen, 1998]. However, if the use of potential vegetation distribution is meaningful for paleoclimate simulations, this modeling strategy might be less adapted to the timescale of CO₂ doubling (less than a century) and is also questionable in many areas where the vegetation is controlled by human activities.

A second limitation in the present study is related to a deficiency of the ISBA land surface scheme. The surface albedo is computed as a linear combination of bare ground, snow, and vegetation albedos, but the weight of the vegetation albedo does not depend on the leaf area index (only on the vegetation cover), so the increase in LAI in experiment LO2 does not lead to a stronger shading effect of the vegetation over snow-covered areas. This radiative impact of the middle-latitudes forests has been emphasized in a recent study [Levis et al., 1999], using a fully coupled climate-vegetation model. Once again, the northward shift of the forests which mainly contributed to the albedo change in this coupled simulation might be overestimated due to the hypothesis of a potential vegetation distribution. Though it is possible that small climate changes may be enough to induce large vegetation responses (since many species grow at or near the limits of tolerated climate), the spatial distribution of the ecosystems does not only depend on climate. The migration of species is still very roughly represented in climate models, so the changes in vegetation cover, simulated by fully coupled climate-vegetation models, remain very uncertain.

More observational studies are needed for a better understanding of the behavior of individual species and continental-scale ecosystems under CO₂-enriched conditions [Bazzaz, 1990; Field et al., 1995; Melillo et al., 1995]. Besides the mean climate the interannual variability and the extreme climatic events (storms, floods, or droughts) may also have dramatic consequences on the vegetation. The terrestrial ecosystems are affected by an array of environmental factors that vary in space and time, and further works are necessary to understand the simultaneous effects of multiple factors on vegetation. While simple experiments may show the effects of a single factor, it is the timing and intensity of interactions between multiple factors that determine how a plant responds to environmental change. Therefore fully interactive soil-vegetation-atmosphere models are needed to address this question properly. The present study is a preliminary attempt to assess the role of the vegetation feedbacks in climate change experiments, given a range of possible vegetation responses. The results demonstrate that vegetation feedbacks should not be ignored in climate change experiments and that GCMs could benefit from having a more interactive representation of the biosphere [Foley et al., 1998; 1999]. Such a scheme is under development at Météo-France [Calvet et al., 1998] and will be included in the ARPEGE climate model in the near future.

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