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Biogeophysical versus biogeochemical feedbacks of large-scale land cover change

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Abstract. Large-scale changes in land cover affect near-surface energy, moisture and momentum fluxes owing to changes in surface structure (referred to as biogeophysical effects) and the atmospheric CO_2 concentration owing to changes in biomass (biogeochemical effects). Here we quantify the relative magnitude of these processes as well as their synergisms by using a coupled atmosphere-biosphere-ocean model of intermediate complexity. Our sensitivity studies show that tropical deforestation tends to warm the planet because the increase in atmospheric CO_2 and hence, atmospheric radiation, outweighs the biogeophysical effects. In mid and high northern latitudes, however, biogeophysical processes, mainly the snow-vegetation-albedo feedback through its synergism with the sea-ice-albedo feedback, win over biogeochemical processes, thereby eventually leading to a global cooling in the case of deforestation and to a global warming, in the case of afforestation.

Introduction

A significant amount of evidence, theoretical and empirical, has gradually been acquired which indicates that changes in land cover matter in the global climate system. Examples arise from the interpretation of paleoclimatic and paleobotanic records (e.g., [*Kutzbach et al.*, 1996], [*Claussen and Gayler*, 1997]) or historical land use (e.g., [*Bonan*, 1999]). Changes in land cover affect global climate by feedbacks between vegetation and atmosphere which directly modify near-surface energy, moisture, and momentum fluxes via changes in albedo, roughness, leaf area, for example, and by changes in atmospheric CO_2 concentration owing to changes in biomass. In the following we refer to these processes as to biogeophysical and biogeochemical effects, respectively. So far most studies focussed on either biogeophysical (e.g., [*Betts*, 1999],[*Brovkin et al.*, 1999], [*Chase*, 2000],[*Kleidon et al.*, 2000],[*Pitman and Zhao*, 2000]) or biogeochemical (e.g., [*DeFries et al.*, 1999]) effects of land cover changes; hence little is known about the relative magnitude nor the interaction between these processes on the global scale. Therefore, we performed a sensitivity study and a factor-separation analysis using scenarios of large-scale land cover change in different regions of the world. The scenarios are not meant to resemble any realistic scenarios of historic or potential future land cover change. Instead, they are part of a "thought experiment" in which we study the principles of vegetation-climate interaction and the role of boreal and tropical vegetation in the coupled system.

Methodology

To analyze biospheric feedbacks operating in the climate system we used a climate-system model of intermediate complexity CLIMBER-2 (version 2.3) which encompasses all relevant components of the climate system under consideration including terrestrial and oceanic carbon cycles. The model CLIMBER-2.3 has a coarse resolution of 10 degrees in latitude and 51 degrees in longitude [*Petoukhov et al.*, 2000]. The model encompasses a 2.5-dimensional dynamical-statistical atmosphere model, a multibasin, zonally averaged ocean model including sea ice and an oceanic carbon cycle, and a dynamic model of terrestrial biosphere. Results of CLIMBER-2.3 compare favorably with data of present-day climate [*Petoukhov et al.*, 2000], with paleoclimatic reconstructions [*Claussen et al.*, 1999] and with results from comprehensive models when applied to a broad spectrum of sensitivity analyses [*Ganopolski et al.*, 2000]. Concerning vegetation dynamics and terrestrial carbon fluxes, our model results agree with others [*Cramer et al.*, 2000].

To define a state of reference as base line for our sensitivity study we performed a control simulation (CNTL) in which the fully coupled atmosphere-ocean-vegetation model was run to an equilibrium under pre-industrial CO_2 concentration of 280 ppm. Earth orbital parameters and location of inland ice masses were kept constant at present-day values.

The sensitivity of the coupled atmosphere-ocean-vegetation system to large-scale changes in land cover was tested by complete deforestation and afforestation from the potential equilibrium state within zonal belts of 10 degrees width. At the time of deforestation, all forest carbon pools in the belt of deforestation were instantaneously emptied and are kept empty. Then, grassland was allowed to enter the deforested region and carbon pools of grassland were allowed to accumulate. Vegetation outside the area of deforestation was also free to expand or shrink. The simulation was continued for another 1000 years such that atmosphere, ocean, and vegetation including oceanic and terrestrial carbon pools could reach a new equilibrium. In the case of afforestation, the tree fraction was set to unity, and the forest carbon pools were allowed to adjust freely in the belt of afforestation.

Global Sensitivities

In Figure 1a, the change in atmospheric CO_2 concentrations in relation to changes in tree fraction (in units of CO_2 ppm per 10^6 km^2 of forest area) are shown for deforestation and afforestation, respectively. As expected, the sensitivity $\Delta CO_2/\Delta F_t$ is negative in all cases, because deforestation ($\Delta F_t < 0$) leads to a increase of CO_2 in the atmosphere,

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and afforestation ($\Delta F_t > 0$), to a decrease. The sensitivity is larger in the tropics than in high northern latitudes, roughly by a factor of 2. Because of higher biomass stock in the tropics, the net release of carbon from tropical deforestation (for example of some 14.3 kg C m^{-2} in the case of deforestation in 0°S-10°S) is large in comparison with the net release from boreal deforestation (some 5.5 kg C m^{-2} in the belt of 50°N-60°N). The net release encompasses the release of carbon from trees biomass and soil under the forest and the uptake of carbon by grassland which is allowed to enter the deforested area as well as changes in regions not directly affected by deforestation. The later changes are small. For example, tropical deforestation in the belt of 0°S-10°S leads to an overall net change of some -136 Pg C and an increase in biomass and soil carbon of some +15 Pg C outside the region of deforestation. For boreal deforestation in the belt of 50°N-60°N, we find net values some -90 Pg C and some -9 Pg C, respectively. Qualitatively, but with opposite signs, the same is true in the cases of afforestation. The change in global mean, near-surface annual temperature relative to the change in tree fraction is depicted in Figure 1b. $\Delta T_{globe}/\Delta F_t$ appears to be negative in low latitudes, but positive in high northern latitudes.

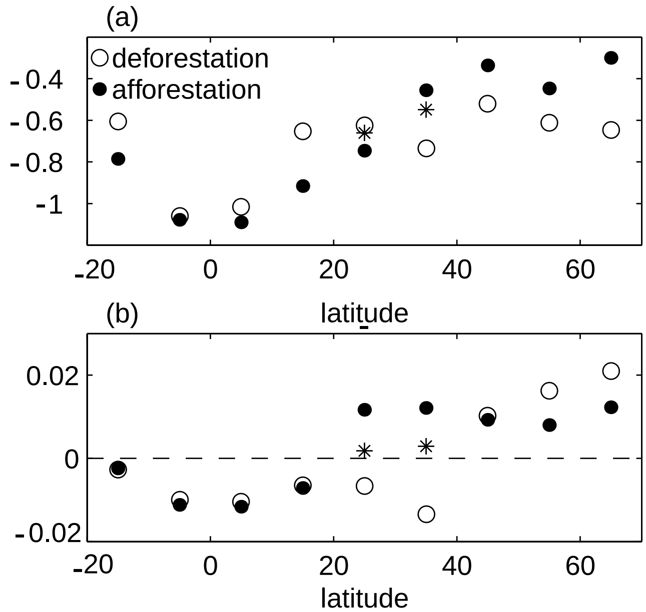

Figure 1. Ratio of changes of global atmospheric CO_2 concentration (a) and global mean near-surface temperature (b) to changes in forest area owing to deforestation and afforestation in latitudinal belts of 10 degrees width. Units given in ppmv CO_2 per 10^6 km^2 and K per 10^6 km^2 of tree area directly affected by deforestation (open circles) or afforestation (full circles), respectively. An asterix indicates a simulation in which deserts, i.e. grid cells with marginal or no vegetation cover, are left unchanged in the case of afforestation.

Contribution Factors

To explain the change in temperature sensitivity, we analyzed biogeochemical and biogeophysical feedbacks separately. For two cases of deforestation scenarios, in high northern latitudes (50°N-60°N) and in the tropics (0°-10°S), we performed a factor-separation analysis [*Stein and Alpert*, 1993]. In addition to the control run (labelled CNTL) and the full simulation (DPC), we ran a simulation (DP) in which near-surface energy, moisture and momentum fluxes were allowed to respond to the change in land cover, but carbon storages were kept constant at their respective values obtained by the CNTL simulation. In a fourth simulation (DC), only changes in carbon pools were allowed, while near-surface energy, moisture, and momentum fluxes are not directly affected by deforestation. The difference between simulations DP and CNTL is interpreted as pure biogeophysical contribution to the effects of deforestation (labelled XP in Table 1), and the difference between DC and CNTL, as pure biogeochemical contribution (XC). By comparing all simulation DPC, CNTL, DP, and DC we can evaluate the synergism between biogeophysical and biogeochemical contributions (XPC = DPC + CNTL -DP -DC). The contribution factors resulting from the factor-separation analysis are listed in Table 1.

Table 1. Contribution of biogeophysical and biogeochemical processes to changes in near-surface global mean (T_{globe}) and regional (T_{region}) temperatures owing to complete deforestation in zonal belts of 50°N - 60°N and 0° - 10°S, respectively. DPC refers to results of the fully coupled atmosphere-ocean-vegetation model and CNTL, to the equilibrium simulation without any external land cover change. XP is the pure biogeophysical contribution, XC the pure biogeochemical contribution, and XPC the synergism between these contributions (see text).

	50°N - 60°N		0° - 10°S	
	T_{globe}	T_{region}	T_{globe}	T_{region}
DPC-CNTL	-0.11	-0.67	+0.16	+0.29
XP	-0.23	-0.82	-0.04	+0.13
XC	+0.09	+0.12	+0.19	+0.15
XPC	+0.03	+0.03	+0.02	+0.01

The biogeophysical contribution to changes in global and regional temperatures are negative, i.e., biogeophysical processes tend to cool the near-surface atmosphere - except for the tropics, where temperatures in the region of deforestation increase (see subsets DP-CNTL in Figure 2). The cooling in high northern latitudes can be attributed to the snow-vegetation albedo feedback, or more precisely to the synergism between the biogeophysical feedback and the sea-ice albedo feedback. The snow-vegetation albedo feedback, or sometimes called taiga-tundra feedback, arises because of the difference in the albedo of snow-covered forest and snow-covered flat vegetation or polar desert. A decrease in forest cover enhances the albedo of the deforested region mainly in spring and early summer, thereby leading to a cooling. Cooling at high northern latitudes favors expansion of Arctic sea ice which in turn, increases albedo, thereby exacerbating the cooling. These results are in line with previous investigations (e.g., [*Bonan et al.*, 1992],[*Brovkin et al.*, 1999]). Deforestation not only yields an increase in surface albedo, but also a reduction in evaporation. The latter effect tends to increase the sensible heat flux which leads to a warming of the near-surface atmosphere in summer (not shown); however, the change in albedo obviously outweighs the change in evaporation in high northern latitudes on annual average. In the tropics, the hydrological effect wins which leads to a strong decrease in evapotranspiration and,

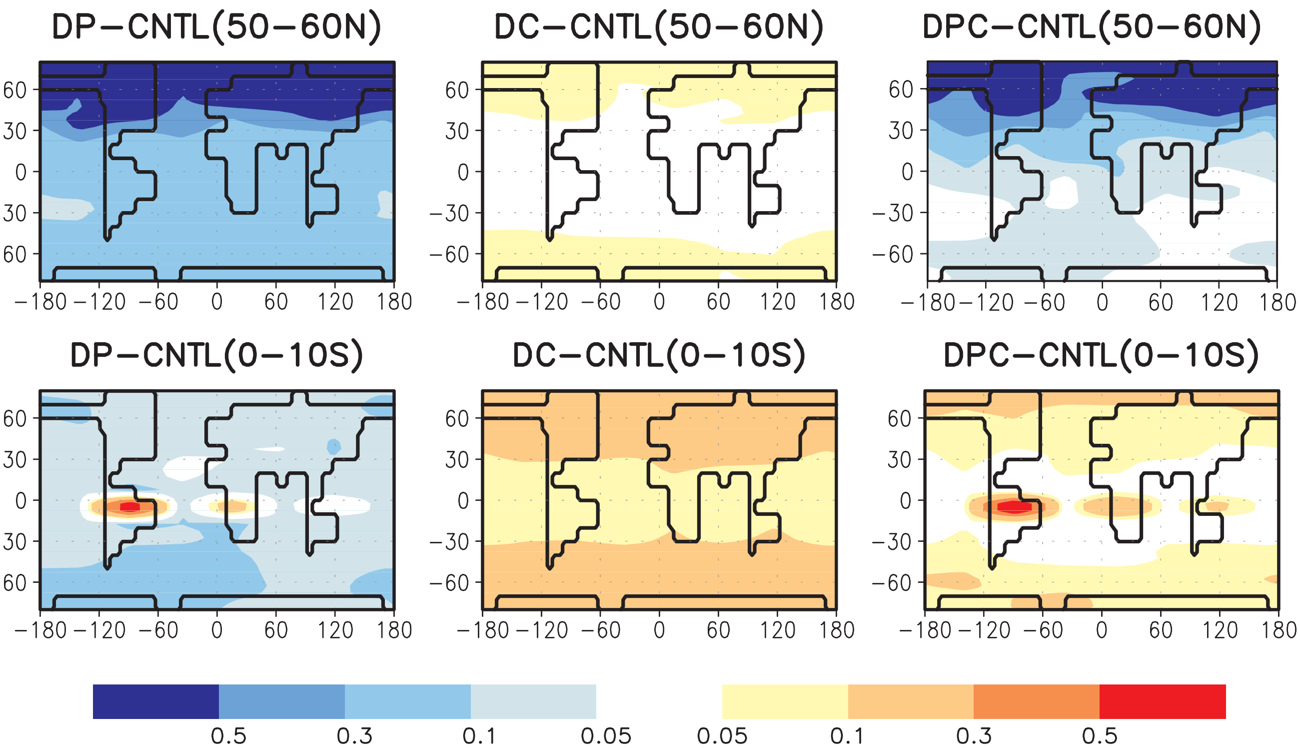

Figure 2. Global pattern of temperature differences between three deforestation simulations (DP, DC, DPC) and the control climate (CNTL). In all deforestation simulations, deforestation is applied to boreal forests the zonal belt between 50°N and 60°N (labelled 50-60N) and to tropical forests in the belt 0° and 10°S (0-10S). The simulation DPC depicts the response of the fully coupled system. In DP, the carbon storages are fixed to value found in the control climate CNTL, i.e., DP reflects the effects of biogeophysical feedbacks only. In DC, near-surface energy, moisture and momentum fluxes are not directly affected by deforestation, but the carbon fluxes are allowed to change, i.e., DC shows the pure biogeochemical effect of deforestation.

subsequently, precipitation, thereby warming the deforested region [*Henderson-Sellers et al.*, 1993], [*Polcher and Laval*, 1993]. Global cooling owing to biogeophysical effects of tropical deforestation arises because of the reduction in evaporation and subsequent reduction in atmospheric water vapor and, thus, in atmospheric radiation. Diminished atmospheric radiation, in turn, cools the ocean surface leading to further reduction in evaporation and cooling which has been discussed in more detail recently [*Ganopolski et al.*, 2000].

Biogeochemical contribution factors are always positive (see subsets DC-CNTL in Figure 2). In the case of tropical deforestation, they are larger than the contribution factor associated with biogeophysical processes, in the case of boreal deforestation, they are smaller. Interestingly, the release of CO_2 owing to tropical deforestation compensates extra-tropical cooling in high northern latitudes (see subsets DPC-CNTL in Figure 2). Synergisms are much smaller than pure contributions; hence the full signal can be estimated fairly accurately by just adding pure biogeophysical and biogeochemical contributions.

The situation in the northern subtropics is a bit more complicated. Deforestation in the northern subtropics basically means deforestation of subtropical forests in Asia and North America. Afforestation, however, leads to an (artificial) growth of trees in subtropical deserts. A more realistic scenario would be to leave subtropical deserts as they are, i.e., to keep vegetation cover unchanged in grid cells with marginal or none vegetation in the control run. In an additional experiment, we have done this for the deforestation in the belts of 20°N to 30°N and 30°N to 40°N. In either case, we find that the sign of sensitivities for subtropical afforestation is similar to that of afforestation in mid- and high northern latitudes.

Conclusions

With respect to the current discussion on afforestation as a means to curb global warming owing to anthropogenic greenhouse gas emissions, one may conclude from our study that afforestation of boreal zones is counterproductive, because in the long run, this measure will even exacerbate global warming. However, a word of caution has to said. Our study focuses on the equilibrium state achieved after some transient phase.

It has to be emphasized that we did not intended to aim at a realistic transient scenario of anthropogenic land cover change. (The rather large land-cover changes we prescribed in the scenarios come closer, perhaps, to differences in land cover between glacials and interglacials.) Instead, we have explored the role of biogeophysical and biogeochemical effects in the global climate system from a more general point of view. We conclude that tropical and boreal ecosystemsseem to play different roles. Tropical deforestation warms the planet (on annual average), and, in combination with enhanced atmospheric CO_2, it increases terrestrial carbon pools in regions where growth is limited by temperature. Boreal deforestation cools the planet, and it leads to a decrease of terrestrial carbon in regions not directly affected by land cover change.

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