Climate effects of global land cover change

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[1] When changing from grass and croplands to forest, there are two competing effects of land cover change on climate: an albedo effect which leads to warming and an evapotranspiration effect which tends to produce cooling. It is not clear which effect would dominate. We have performed simulations of global land cover change using the NCAR CAM3 atmospheric general circulation model coupled to a slab ocean model. We find that global replacement of current vegetation by trees would lead to a global mean warming of 1.3°C, nearly 60% of the warming produced under a doubled CO_2 concentration, while replacement by grasslands would result in a cooling of 0.4°C. It has been previously shown that boreal forestation can lead to warming; our simulations indicate that mid-latitude forestation also could lead to warming. These results suggest that more research is necessary before forest carbon storage should be deployed as a mitigation strategy for global warming. **Citation:** Gibbard, S., K. Caldeira, G. Bala, T. J. Phillips, and M. Wickett (2005), Climate effects of global land cover change, *Geophys. Res. Lett.*, *32*, L23705, doi:10.1029/2005GL024550.

1. Introduction

[2] Previous studies of the effects of land cover change [*Betts*, 2000; *Bonan*, 2001; *Govindasamy et al.*, 2001; *Hansen et al.*, 1997; *Brovkin et al.*, 1999; *Bonan*, 1997; *Oleson et al.*, 2004] have indicated that direct historical mid-latitude land cover change has increased surface albedo, leading to cooling. These studies suggested that human-induced land cover change from forest to croplands could lead to a cooling of 0.25°C on a global basis [*Govindasamy et al.*, 2001], which may have contributed to the millennial cooling before the 20th century, and that northern mid-latitude agricultural regions are about 1–2°C cooler in the winter and spring compared to the pre-industrial state due to replacement of forest by croplands [*Betts and Falloon*, 2005].

[3] Studies of tropical deforestation are inconclusive on its effects on local or global climate. Deforestation has been found to warm the Amazon basin [*Costa and Foley*, 2000; *Osborne et al.*, 2004], and in central Argentina the effect of vegetation cover is to lower the surface temperature due to increased evapotranspiration [*Nosetto et al.*, 2005]. However several authors [*Chen et al.*, 2001; *Snyder et al.*, 2004; *Avissar et al.*, 2004] have found that tropical deforestation is likely to induce changes in atmospheric circulation, and that these changes may have consequences on precipitation and temperature patterns on a global scale.

[4] Studies investigating the question of global effects of extreme land cover change (from a "desert" planet to a maximally forested one [*Kleidon et al.*, 2000; *Fraedrich et al.*, 1999]) found an overall cooling effect due to increased evapotranspiration in the forested scenario. However, these models used prescribed sea surface temperatures (SSTs), which constrained the effects of land cover change on global temperature.

[5] We have simulated the effects of extreme land cover changes using the Community Land Model (CLM) and the Community Atmosphere Model (CAM), coupled to a slab ocean model in order to investigate their potential to affect the global climate system without the constraint of specified SSTs. Our goal here is not to reproduce the observed pattern of land cover change, nor to realistically simulate possible future scenarios, but rather to bracket the magnitude of temperature change that is possible in the climate system due to changes in land cover.

2. Offline Land Model Simulations

[6] To investigate the effects of land cover change, we performed simulations with and without atmospheric feedbacks. For simulations without atmospheric feedbacks, we used version 3 of the Community Land Model (CLM3 [*Vertenstein et al.*, 2004]) in its offline mode (with prescribed atmospheric climatologies). CLM3 distinguishes 15 types of vegetation, as well as bare ground, lake, and glacier. Up to four vegetation types are allowed per grid cell. Each vegetation type has its own leaf and stem area, root distribution, optical properties, and canopy top and bottom heights [*Bonan et al.*, 2002].

[7] To investigate the effects of changes in vegetation in this model, we replaced the standard vegetation type map used by CLM3 with maps containing only a single vegetation type in 100% of the occupied grid cells. This was done without regard to whether specific vegetation types could realistically grow in a given grid cell. The percentages of lake and glacier in each gridcell were not changed from the nominal value. We ran the model in offline mode, repeatedly using the monthly atmospheric climatologies data corresponding to the year 1998 from the NCEP/NCAR reanalysis dataset (provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA, from their Web site at http://www.cdc.noaa.gov/) for 20 years. We then discarded the first five years of the simulation, and averaged the last 15 years. We also performed a 20-year control run using the standard vegetation type map supplied with the CLM3 distribution.

[8] We find that the range of responses (2-meter air temperature predicted for all the different vegetation types minus the prediction for bare ground) clearly varies

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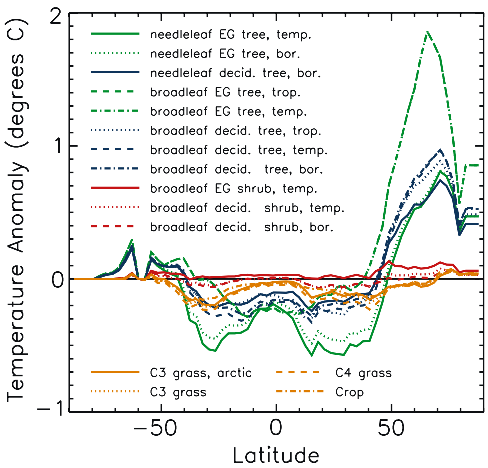

Figure 1. The zonally averaged 2-m air temperature difference between different vegetation types and bare ground in the offline simulation.

depending on latitude, with the strongest responses occurring in the northern region currently occupied by boreal forest (Figure 1). In general, any type of vegetation causes cooling (warming) in low (high) latitudes in comparison to bare ground. Surface albedo change provides the dominant influence in middle and high latitudes [*Betts*, 2000; *Govindasamy et al.*, 2001], with vegetation producing net warming; snow-covered vegetation has much lower albedo than snow-covered bare ground. In the tropics the main effect is via evapotranspiration with vegetation causing net cooling [*Costa and Foley*, 2000]. Evapotranspiration removes heat more efficiently from the surface at low latitudes, partly because of the exponential relationship between temperature and saturation water vapor pressure. Therefore, climate effects of changes in evapotranspiration dominate over the albedo effect at low latitudes. Figure 1 shows that the 15 vegetation types can be roughly divided into two groups: open canopy (grass, shrubs), and closed canopy (trees). Note that Figure 1 does not imply an actual temperature change; instead the temperature differences are to be interpreted as the temperature tendency produced by a localized land cover change on a scale small enough to have no effects on the atmosphere. (The surface temperature response to vegetation change is constrained by the temperature forcing at the reference height). Figure 1 suggests that to estimate the potential range of response to land cover change in a coupled model, it would suffice to consider only two types of vegetation: forest vs. grass and shrublands.

3. Coupled Simulations

[9] To consider effects of atmospheric feedbacks, we used Version 3 of the Community Atmosphere Model (CAM3) [*Collins et al.*, 2004]. The spatial resolution is 2.0° in latitude and 2.5° in longitude, and the model has 26 levels in the vertical. An important aspect of CAM3 is that it has very little systematic bias in the top-of-atmosphere and surface energy budgets. We coupled the CAM3 atmosphere to the CLM land model and to a slab ocean and thermodynamic sea ice model, which allows for interaction with ocean and sea ice components. The slab ocean model employs a spatially and temporally prescribed horizontal ocean heat transport and mixed layer-depth, which ensures realistic SSTs and ice distributions for the present climate.

[10] To approach equilibrium between land, the slab-ocean, and atmosphere, we spun up the coupled models for 20 years, and then ran the model an additional 30 years to obtain climate statistics. We made control runs with both the current vegetation and with no vegetation (bare soil). Based on the results from our offline runs, we ran two vegetated experiments, one with trees and the other with grass/shrubs. For the tree run, the vegetation type was specified in latitude bands, with each band assigned the most common type of tree at that latitude (based on current vegetation). A similar procedure was followed for the grass/shrub simulation. When no trees were present at a particular latitude (near the poles), current vegetation was assumed.

[11] For the tree simulation (Figure 2), compared to bare soil, the overall effect is a warming of 1.6°C globally (2.3°C for the land only). Results from the grass/shrub simulation indicate that the overall influence of grass vs. bare soil is cooling. The amount of cooling is small, 0.03°C globally (0.1°C for the land only). The current vegetation simulation is 0.35°C warmer on a global basis than the bare ground simulation, mainly due to the presence of boreal trees. Actual vegetation is more similar, on average, to open canopy ecosystems than to closed canopy ecosystems. Comparison of the tree and grass runs with the current vegetation run (Table 1) shows that the replacement of current vegetation by trees (grassland) at all latitudes would produce a global mean warming (cooling) of 1.3°C (0.38°C). A pronounced warming of 3.77°C is simulated in the Northern Hemisphere middle and high latitude land surface in the tree simulation.

[12] The temperature difference between the tree simulation and the bare soil simulation (Figure 3a) shows that the

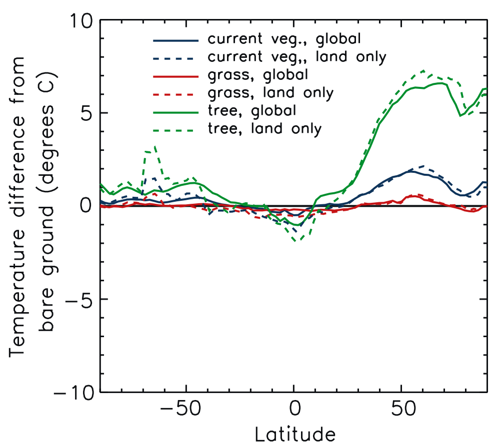

Figure 2. The zonally averaged 2-m air temperature difference for different scenarios (experiment minus bare ground) in the coupled simulations. Solid lines indicate global means; dashed lines are averages over land only. Average temperature anomalies from bare ground are: current vegetation, global: +0.35, land only: +0.39. Grass, global: −0.03; land only: −0.10. Tree, global: +1.6; land only: +2.3.

Table 1. Changes in Annual-Means of Climate Variables in the Tree Simulation With Respect to the Current Vegetation Simulation[a]

	Surface Air Temperature, K	Land Surface Air Temperature, K	Precipitation, %	Albedo, %	Sea Ice Volume, %
Global	1.3 (−0.38)	1.94 (−0.64)	3.2 (0.9)	−8 (2)	−45 (26)
NH (20°N to 90°N)	3.14 (−0.92)	3.77 (−1.23)	5.2 (−0.3)	−12 (2.6)	−54 (32)
TR (20°S to 20°N)	−0.02 (−0.04)	−0.05 (−0.06)	5.4 (2.3)	−4 (1.6)	–
SH (20°S to 90°S)	0.49 (−0.18)	0.64 (−0.12)	0.1 (0.2)	−3 (−0.1)	−18 (10)

[a]Values in parenthesis show the changes in the grassland simulation.

heating effect of forest, which is confined to latitudes poleward of 50° in the offline runs, extends poleward from about 20° in the coupled climate runs (cf. Figures 1 and 2). The albedo-change effect dominates over the evapotranspiration effect from the poles to the tropics in this coupled case. It is also clear from Figure 3a that there are "downwind" heating and cooling effects from the land to the oceans.

[13] Based on the global forest scenario it is not clear whether the heating effect of trees is primarily or exclusively due to the presence of the boreal forest, as opposed to mid-latitude trees. In order to investigate the effects of forestation at mid-latitudes, we ran a simulation in which the vegetation from 30–50° N was replaced by forest, with the rest of the vegetation in its current (control) configuration. The results (Figure 3b) indicate that the direct effect of mid-latitude trees is warming—by 0.68°C compared to bare ground, or 0.27°C compared to current vegetation.

4. Discussion

[14] After 50 years, when the model simulations have approached equilibrium, the difference between the net shortwave flux for the tree and bare ground simulations is 3.17 W/m^2. We have estimated the climate sensitivity of the model from a 50-year doubled CO_2 scenario as 2.2°C, with a radiative forcing of 3.5 W/m^2. This would imply an equilibrium temperature change of 2°C for a radiative forcing of 3.17 W/m^2. This is in close agreement with the model simulated warming between the tree and bare ground scenarios of 1.6°C.

[15] The warming effect due to the presence of trees clearly originates from the effect of trees on the surface albedo (Figure 3c). In the bare ground simulation, the average land albedo is 0.23 for both the offline and the coupled model simulations. It decreases to 0.17 and 0.15 in the offline and coupled cases respectively; snow-covered vegetation has much lower albedo than snow-covered bare ground. The warming leads to a decrease in snow cover—the annual average snowfall decreases 20% at 45–90°N in the tree simulation compared to the bare ground simulation. The decrease in snow leads to a decrease in surface albedo, which leads to an increased absorption of solar radiation, which increases the surface temperature, melting more snow. This snow-albedo feedback effect is responsible for about 25% of the change in surface albedo between the bare ground and tree runs in the coupled model (Figure 3c). It is clear that the maximum surface albedo change occurs in northern areas where snowfall has decreased due to the increase in temperature. In the case of a warmer global climate, however, this albedo effect is likely to be less pronounced than in present-day climate.

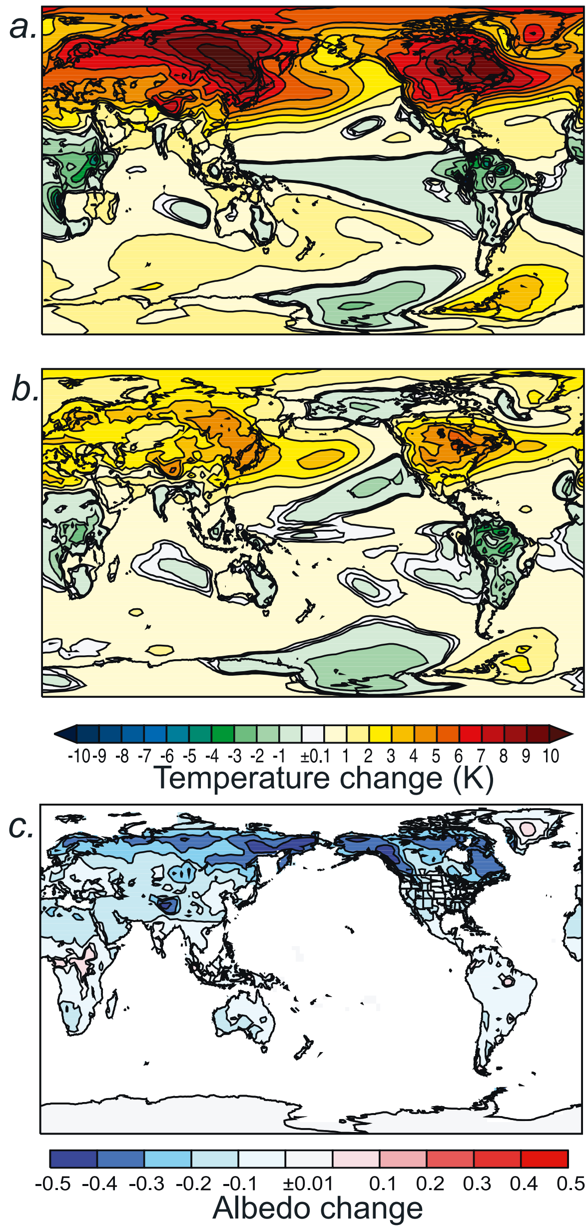

Figure 3. Annual mean 2-m air temperature (°C) for the tree simulation minus the bare ground simulation (top panel), for the tree simulation that has trees over 30–50°N and current vegetation elsewhere minus the bare ground simulation (middle panel), and annual mean surface albedo for the tree simulation minus the bare ground simulation (bottom panel). Note the "downwind" heating and cooling effects from the land to the ocean.

[16] *Kleidon et al.* [2000] have investigated the effects of a change from a desert world to a green planet in a general circulation model with fixed SSTs. In contrast to our results, this model produced a net global land temperature change of $-1.2°C$, and a global change (including oceans) of $-0.3°C$. It is likely that the differences are related to differing treatments of various physical and hydrological processes across the models. Our model, in contrast to *Kleidon et al.* [2000], allows the SSTs to change, and this allows a more realistic representation of the feedbacks between the land and oceans. A dynamic ocean model would likely further amplify the cooling due to deforestation; *Renssen et al.* [2003] reported a southward shift of the main deep convection site in the Atlantic Ocean, leading to enhanced cooling due to a southward expansion of Arctic sea ice.

[17] We can estimate the potential for cooling by carbon sequestration by reforestation if we assume that 10 kg/m^2 of carbon can be stored by planting forests [*Betts*, 2000]. This implies a global land carbon sequestration of 1500 PgC ($\sim$700 ppmv). The cooling potential of storing 1500 PgC is $\sim$ 3.5°C for our model with a climate sensitivity of 2.2°C and a radiative forcing of 3.5 W/m^2 per doubling of CO_2. Since the albedo-induced warming is 1.3°C, the potential cooling due to sequestration of 3.5°C is offset $\sim$40% by the warming due to albedo change. While the albedo change is permanent, the perturbation to atmospheric CO_2 content would be damped by equilibration with the ocean and ultimately with the rock cycles. Using the exponential representation of the decay of a perturbation to atmospheric CO_2 presented by *Maier-Reimer and Hasselmann* [1987], it can be estimated that after $\sim$80 years, less than 40% of the initial perturbation to atmospheric CO_2 would remain in the atmosphere, so that after this time the global forestation would produce net warming, considering both albedo and carbon-storage effects.

[18] Our simulations indicate that the magnitude of warming due to global forestation is of the same order of magnitude as the cooling due to carbon-storage effects. This has important policy implications, since incentives for tree plantations in mid- and high-latitudes may, on long time-scales, produce the opposite effect to that desired. Whereas cooling due to carbon cycle effects may dominate on the decadal time scale, warming associated with albedo effects may dominate on the century time scale. Previous studies [*Betts*, 2000; *Claussen et al.*, 2001] have found that the albedo-change-induced warming due to boreal reforestation could be comparable and opposite to the carbon storage effects of cooling. *Claussen et al.* [2001] have shown, using a coupled climate-carbon cycle model, that boreal forestation leads to warming as the albedo effect is stronger than the effect of CO_2 reduction, while the net effect of tropical forestation is a cooling. Further study is needed to assess whether forestation in mid- latitudes could play a role in the mitigation of climate change.

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References

Avissar, R., R. R. da Silva, and D. Werth (2004), Implications of tropical deforestation for regional and global hydroclimate, *Ecosyst. Land Use Change*, *153*, 73–83.

Betts, R. A. (2000), Offset of the potential carbon sink from boreal forestation by decreases in surface albedo, *Nature*, *408*, 187–200.

Betts, R. A., and P. Falloon (2005), Biogeophysical effects of land use on climate: Model simulations of radiative forcing and large-scale temperature change, *Agric. For. Meteorol.*, in press.

Bonan, G. B. (1997), Effects of land use on the climate of the United States, *Clim. Change*, *37*, 449–486.

Bonan, G. B. (2001), Observational evidence for reduction of daily maximum temperature by croplands in the Midwest United States, *J. Clim.*, *14*, 2430–2442.

Bonan, G. B., K. W. Oleson, M. Vertenstein, S. Levis, X. Zeng, Y. Dai, R. E. Dickinson, and Z.-L. Yang (2002), The land surface climatology of the Community Land Model coupled to the NCAR Community Climate Model, *J. Clim.*, *15*, 3123–3149.

Brovkin, V., A. Ganapolski, M. Claussen, C. Kubatzki, and V. Petoukhov (1999), Modeling response to historical land cover change, *Global Ecol. Biogeol.*, *8*, 509–517.

Chen, T. C., J. H. Yoon, K. J. St. Croix, and E. S. Takle (2001), Suppressing impacts of the Amazonian deforestation by the global circulation change, *Bull. Am. Meteorol. Soc.*, *82*, 2209–2216.

Claussen, M., V. Brovkin, and A. Ganopolski (2001), Biophysical versus biogeochemical feedbacks of large-scale land cover change, *Geophys. Res. Lett.*, *28*, 1011–1014.

Collins, W. D., et al. (2004), Description of the NCAR Community Atmosphere Model (CAM 3.0), *NCAR Tech. Note NCAR/TN-464+STR*, Natl. Cent. for Atmos. Res., Boulder, Colo.

Costa, M. H., and J. Foley (2000), Combined effects of deforestation and doubled atmosphere CO2 concentration on the climate of Amazonia, *J. Clim.*, *13*, 18–24.

Fraedrich, K., A. Kleidon, and F. Lunkeit (1999), A green planet versus a desert world: Estimating the effect of vegetation extremes on the atmosphere, *J. Clim.*, *12*, 3156–3163.

Govindasamy, B., P. B. Duffy, and K. Caldeira (2001), Land use changes and Northern Hemisphere cooling, *Geophys. Res. Lett.*, *28*(2), 291–294.

Hansen, J., M. Sato, A. Lacis, and R. Ruedy (1997), The missing climate forcing, *Philos. Trans. R. Soc. London, Ser. B*, *352*, 231–240.

Kleidon, A., K. Fraedrich, and M. Heimann (2000), A green planet versus a desert world: Estimating the maximum effect of vegetation on the land surface climate, *Clim. Change*, *44*, 471–493.

Maier-Reimer, E., and K. Hasselmann (1987), Transport and storage of CO_2 in the ocean—An inorganic ocean-circulation carbon cycle model, *Clim. Dyn.*, *2*, 63–90.

Nosetto, M. D., E. G. Jobbagy, and J. M. Paruelo (2005), Land-use change and water losses: the case of grassland afforestation across a soil textural gradient in central Argentina, *Global Change Biol.*, *11*, 1101–1117.

Oleson, K. W., G. B. Bonan, S. Levis, and M. Vertenstein (2004), Effects of land use change on North American climate: Impact of surface datasets and model biogeophysics, *Clim. Dyn.*, *23*, 117–132.

Osborne, T. M., D. M. Lawrence, J. M. Slingo, A. J. Challinor, and T. R. Wheeler (2004), Influence of vegetation on the local climate and hydrology in the tropics: Sensitivity to soil parameters, *Clim. Dyn.*, *23*, 45–61.

Renssen, H., H. Goosse, and T. Fichefet (2003), On the non-linear response of the ocean thermohaline circulation to global deforestation, *Geophys. Res. Lett.*, *30*(2), 1061, doi:10.1029/2002GL016155.

Snyder, P. K., J. A. Foley, M. H. Hitchman, and C. Delire (2004), Analyzing the effects of complete tropical forest removal on the regional climate using a detailed three-dimensional energy budget: An application to Africa, *J. Geophys. Res.*, *109*, D21102, doi:10.1029/2003JD004462.

Vertenstein, M., K. Oleson, S. Levis, and F. Hoffman (2004), Community Land Model Version 3.0 (CLM3.0) user's guide. (Available at http://www.cgd.ucar.edu/tss/clm/distribution/clm3.0/index.html)

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