Changing feedbacks in the climate– biosphere system

F Stuart Chapin III[1*], James T Randerson[2], A David McGuire[3], Jonathan A Foley[4], and Christopher B Field[5]

Ecosystems influence climate through multiple pathways, primarily by changing the energy, water, and greenhouse-gas balance of the atmosphere. Consequently, efforts to mitigate climate change through modification of one pathway, as with carbon in the Kyoto Protocol, only partially address the issue of ecosystem–climate interactions. For example, the cooling of climate that results from carbon sequestration by plants may be partially offset by reduced land albedo, which increases solar energy absorption and warms the climate. The relative importance of these effects varies with spatial scale and latitude. We suggest that consideration of multiple interactions and feedbacks could lead to novel, potentially useful climate-mitigation strategies, including greenhouse-gas reductions primarily in industrialized nations, reduced desertification in arid zones, and reduced deforestation in the tropics. Each of these strategies has additional ecological and societal benefits. Assessing the effectiveness of these strategies requires a more quantitative understanding of the interactions among feedback processes, their consequences at local and global scales, and the teleconnections that link changes occurring in different regions.

Front Ecol Environ 2008; 6(6): 313–320, doi:10.1890/080005

The factors that determine properties of both ecosystems and the climate system have changed more rapidly in the past 50 years than during the previous 10 000 years (Steffen *et al.* 2004; Figure 1). Our children will probably see even more profound changes during their lifetimes. Some of these, such as changes in climate and atmospheric composition, alter the dynamic interactions between land, ocean, and atmosphere and, therefore, future transformations in climate and the ecosystems on which society depends (IPCC 2007a). Development of policies that reduce rates of climate change, while sustaining the services provided by ecosystems, require a clear understanding of these dynamics.

In a nutshell:

- Ecosystems affect climate through multiple pathways, each requiring policy considerations
- Regions differ in management actions that will most effectively mitigate climate-change effects
- Reductions in desertification rates reduce the likelihood of drought in arid zones; reductions in greenhouse-gas emissions will be particularly effective in reducing climate forcing in industrialized nations; and reductions in deforestation rates will be especially effective in the tropics
- Equity issues are a central challenge in developing global strategies to mitigate climate change

[1]*Institute of Arctic Biology, University of Alaska Fairbanks, Fairbanks, AK* [*]*(fffsc@uaf.edu);* [2]*Department of Earth System Science, University of California, Irvine, CA;* [3]*US Geological Survey, Alaska Cooperative Fish and Wildlife Research Unit, University of Alaska Fairbanks, Fairbanks, AK;* [4]*Center for Sustainability and the Global Environment, University of Wisconsin, Madison, WI;* [5]*Department of Global Ecology, Carnegie Institution of Washington, Stanford, CA*

Changes in ecosystems influence the climate system through several processes (Figure 2), including (1) emission of greenhouse gases, which cause an imbalance in the Earth's energy budget at the top of the atmosphere; (2) altered albedo (the proportion of solar radiation that the Earth's surface reflects back to space), which influences the amount of heat transferred from ecosystems to the atmosphere; (3) altered evapotranspiration (evaporation from the Earth's surface plus that from leaves), which cools the surface and provides moisture to form clouds and fuel atmospheric mixing; (4) altered longwave radiation, which depends on surface temperature and cloudiness; (5) changes in production of aerosols (small particles that scatter and absorb light); and (6) changes in surface roughness, which determines the strength of coupling between the atmosphere and the surface and, therefore, the efficiency of water and energy exchange. For trace gases and aerosols, the impact of an individual constituent on climate depends on the magnitude of the instantaneous forcing and the turnover time of each constituent in the atmosphere (ie the total quantity divided by the average rate of input and loss; Table 1). In general, energy, water, and highly reactive compounds from fossil-fuel emissions (eg nitric oxide, sulfur dioxide) have such short atmospheric lifespans that they have strong local or regional effects, as well as global consequences. In contrast, the effects of greenhouse gases, such as carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O), are globally dispersed, so their impacts are averaged over the entire planet.

Discussions about, and efforts to reduce human impacts on, the climate system have generally focused on green-

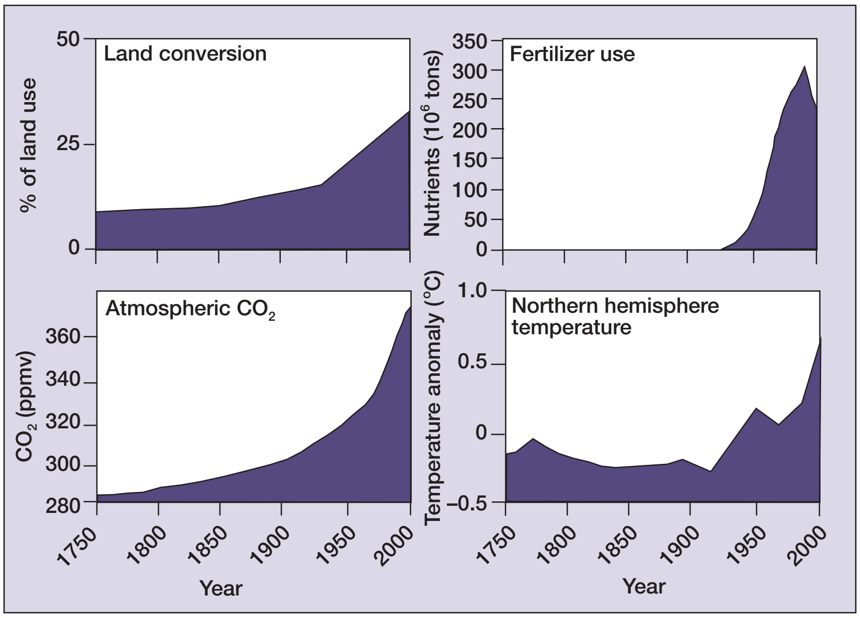

Figure 1. *Selected changes in ecosystems (land conversion and agricultural fertilizer use) and the climate system (atmospheric* CO_2 *and surface air temperature) since 1750 (Steffen* et al. *2004).*

house-gas balance only, largely ignoring the multiple pathways by which ecosystems interact with the climate system (Field *et al.* 2007). The Kyoto Protocol, for example, addresses only CO_2 emissions and ignores the effects of altered ecosystem carbon storage on other pathways of ecosystem–atmosphere interaction. Here, we discuss four major ways in which human activities have altered multiple feedbacks to the climate system: (1) CO_2 emissions, (2) climate warming, (3) desertification, and (4) changing forest cover. We show that consideration of multiple feedbacks between ecosystems and the atmosphere raises important new scientific questions and suggests new policy directions that might reduce the overall human footprint on the climate system.

Table 1. Turnover time and spatial scale of linkage between atmospheric constituents and the climate system

Constituent	*Turnover time[1]*	*Spatial scale*
CO_2	3 years[2]	Global
CH_4	8.4 years	Global
N_2O	120 years	Global
H_2O	10 days	Sub-continental
Aerosols	Days to weeks	Regional to continental
NO_x	< 1 day	Regional
SO_2	< 1 day	Regional

Notes: This represents the average time that each constituent remains in the atmosphere. The atmosphere mixes globally in about 1 year. [1]Turnover time = atmospheric pool divided by the annual flux to the atmosphere. [2]This is the mean residence time of CO_2 in the atmosphere with respect to exchange with terrestrial ecosystems (gross primary production) and air–sea fluxes. Note that injections of CO_2 into the atmosphere from fossil-fuel burning have a lifetime of hundreds to thousands of years, due to their effects on ocean chemistry. Ocean chemistry was in steady state with respect to atmospheric CO_2 levels prior to the Industrial Revolution; now, fossil-fuel CO_2 is consuming carbonate ions and lowering pH in the oceans, which in turn reduces their uptake capacity.

Ecosystem effects of CO_2

Fossil-fuel emissions of CO_2 are thought to be the largest direct human cause of recent climate warming (IPCC 2007a). However, terrestrial ecosystems fix through photosynthesis and release through respiration/combustion about ten times more CO_2 than is released from combustion of fossil fuels and altered land use annually. Consequently, ecosystems have been viewed as one possible avenue for "solving the CO_2 problem" without reducing anthropogenic emissions. Over the past several decades, the land and oceans have indeed acted as a large negative feedback to climate warming (ie a feedback that *slows* climate warming) by absorbing and sequestering about 55–60% of the CO_2 released to the atmosphere by fossil-fuel emissions and land-use change (Canadell *et al.* 2007; Figure 3). The magnitude of this feedback depends on several processes in terrestrial ecosystems, including (1) the capacity of land plants to increase photosynthesis and carbon storage in response to rising atmospheric CO_2, and (2) the sensitivity of net primary production and heterotrophic respiration to increasing temperatures and shifting patterns of drought (Friedlingstein *et al.* 2006). Initial model simulations suggest that terrestrial ecosystem processes may play an important role in regulating these feedbacks over the 21st century. Over timescales of several centuries, ocean processes will probably become increasingly important, as terrestrial sinks saturate or become sources. Over this period, however, ocean sinks may also decrease in the quantity of CO_2 absorbed, as surface heating increases stratification and slows overturning.

The capacity of land and ocean sinks to remove anthropogenic carbon derived from fossil-fuel combustion and land-use change from the atmosphere has declined from about 60% to 55% of human emissions between 1960 and 2007 (Canadell *et al.* 2007). This proportion may continue to decline as the capacity of terrestrial ecosystems to sequester carbon saturates and forest cover decreases (Gitz and Ciais 2003). It is therefore unlikely that terrestrial ecosystems and oceans, as presently constituted, will "solve the CO_2 problem", without substantial reductions in fossil-fuel emissions.

At the scale of a single leaf or plant, increasing atmospheric CO_2 above current ambient concentrations typically causes an increase in photosynthesis and/or a decrease in transpiration (Drake *et al.* 1997) – effects that also contribute to patterns observed at landscape-to-global scales (Field *et al.* 2007; Lobell and Field 2008). Although some of the global-scale trends in tree growth can be explained without invoking a CO_2 response (Caspersen *et al.* 2000), elevated CO_2 accounts for at least some of the observed growth enhancement, and probably some increment of carbon storage (Norby *et al.* 2005; Long *et al.* 2006).

Carbon sequestration by terrestrial ecosystems may also alter other feedbacks to the climate system. The stimulation of photosynthesis by rising CO_2, for example, may increase leaf area and forest cover, slightly darkening the Earth's surface and increasing absorption of radiation. This could have a warming effect that partly offsets the negative feedback resulting from carbon sequestration (Matthews *et al.* 2007; Figure 3). In addition, the decrease in transpiration by individual leaves in response to elevated levels of CO_2 (Field *et al.* 1995) may warm the surface (Sellers *et al.* 1996), although other adjustments in ecosystem structure and species composition – including increases in leaf area and vegetation cover – may cancel this effect at regional scales. The net effect of ecosystems on carbon sequestration is least pronounced in old forests, especially those whose growth is strongly limited by nutrient availability (Körner *et al.* 2005).

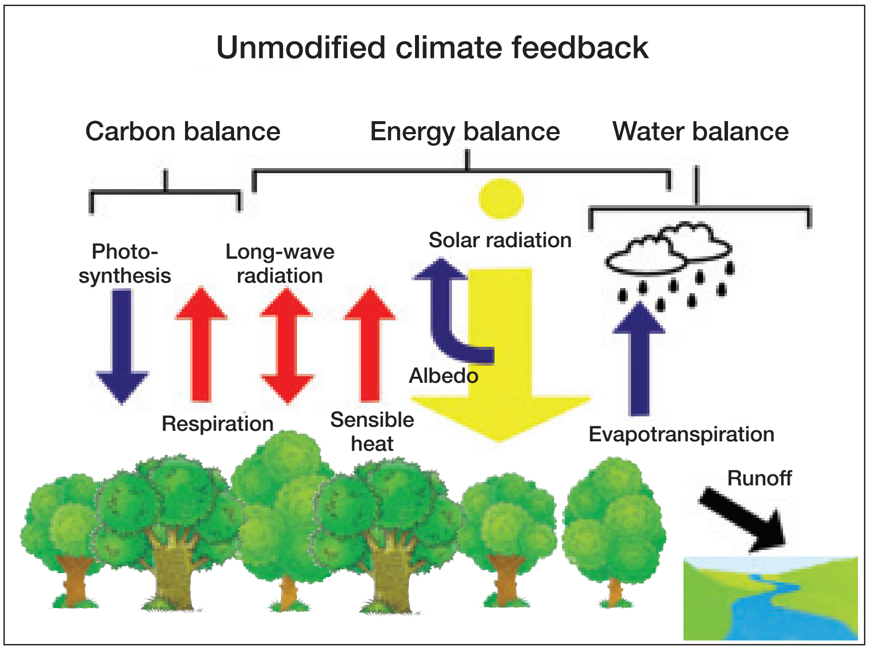

Figure 2. *Three major categories of climate feedbacks (each shown by the arrows beneath the brackets) between ecosystems and the climate system. Carbon balance is the difference between CO_2 uptake by ecosystems (photosynthesis) and CO_2 loss to the atmosphere by respiration. Energy balance is the balance between incoming solar radiation, the proportion of this incoming solar radiation that is reflected (albedo), and the transfer of the absorbed radiation to the atmosphere as sensible heat (warming the near-surface air), evapotranspiration (cooling the surface), and long-wave radiation. Water balance between the ecosystem and atmosphere is the difference between precipitation inputs and water return in evapotranspiration; the remaining water leaves the ecosystem as runoff. Each of these ecosystem–atmosphere exchanges influences climate, with arrows showing the direction of mass or energy transfer. Cooling effects on climate are shown by blue arrows; warming effects by red arrows. Other climate feedbacks that are influenced by ecosystems (but not shown in this diagram) include effects of particulates, CH_4, N_2O, ozone, and reflectance by clouds.*

In summary, ecosystems respond to increased atmospheric concentrations of CO_2 primarily by removing a fraction of it from the atmosphere (ie carbon sequestration). The resulting cooling effect on climate may be offset to a modest degree by changes in canopy cover that increase energy absorption. Furthermore, the net cooling effect on climate of carbon sequestration in land and ocean ecosystems is likely to decline as elevated CO_2 levels reduce the sensitivity of photosynthesis to additional changes, as tropical forest cover declines, and as ocean stratification and acidification intensifies. Thus, ecosystems are unlikely to solve the problem of rising atmospheric CO_2. Reduction in CO_2 emissions in those sectors and nations that produce the greatest quantity of emissions is likely to be the most effective long-term policy for reducing atmospheric CO_2 concentration.

■ Climate warming

Recent climate warming, resulting mainly from accumulation of greenhouse gases in the atmosphere, alters multiple feedbacks to the climate system, most of which amplify the rate of warming. As expected, these effects are most pronounced in cold climates, where warming has been most pronounced and physical and biological processes are particularly sensitive to temperature.

In the Arctic, warming has caused an increase in both photosynthesis and respiration, but the net effect on carbon balance has been variable regionally, with increased plant growth and net carbon accumulation in moist areas and net carbon loss in dry areas; the net effect has probably been a small amount of carbon sequestration (McGuire *et al.* 2000; Callaghan *et al.* 2004; Sitch *et al.* 2007), slowing climate warming.

However, changes in the surface energy budget in the Arctic have had a much larger and more consistent feedback to climate. In the northern oceans, the summer extent of sea ice has decreased; the resulting decline in albedo causes more absorption of radiation, which warms the air and water. Since the impacts of changes in energy balance are most pronounced at local to regional scales, this heating causes more melting of sea ice and, thus, further warming of the ocean and atmosphere. A similar process occurs on land, as warmer climate causes snow to melt earlier (by 1–2 days per decade), which reduces albedo, increases energy absorption, and amplifies the rate of local to regional warming (Chapin *et al.* 2005; Euskirchen *et al.* 2007; Figure 3). Over the longer term, increases in shrub and tree cover will probably have an even stronger positive feedback to Arctic warming, by reducing both summer and

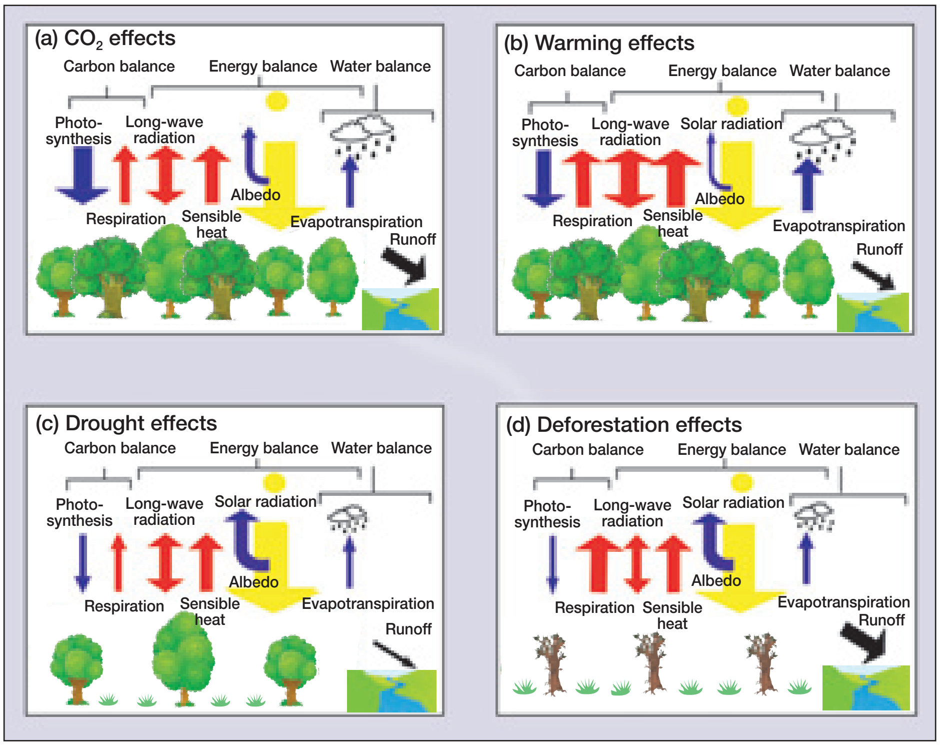

Figure 3. *Human modifications of climate feedbacks. The width of the arrows shows the magnitude of changes in each flux, relative to the standard case shown in Figure 2 (in which arrows are normalized to a standard width), resulting from (a) elevated CO_2, (b) climate warming, (c) drought, and (d) deforestation, as described in Panel 1.*

winter albedo (Chapin *et al.* 2005).

In contrast to a 30-year record of greening in the Arctic – presumably in response to climate warming, as documented by satellite records of vegetation greenness – boreal forest greenness increased until the 1990s and has since declined, as a result of some combination of increasing drought, insect outbreaks, and wildfire (Goetz *et al.* 2005). This suggests a decline in the capacity of boreal forests to sequester carbon. As in the Arctic, the feedbacks from a changing energy budget appear to predominate. Earlier snowmelt acts as a positive feedback to climate warming, just as in the Arctic.

Wildfire disturbances release carbon (a positive feedback to warming), but also replace an absorptive tree canopy with a reflective snow surface (increased albedo) during fall, winter, and spring. This reduces energy absorption and acts as a negative feedback to warming (Figure 4). The net effect of fire is probably a small negative feedback to warming at a global scale, but with much of the albedo-driven cooling concentrated over boreal land areas (Randerson *et al.* 2006). Future increases in boreal fires, however, may also increase the deposition of black carbon on sea ice and the Greenland ice sheet (Flanner *et al.* 2007; McConnell *et al.* 2007), with uncertain long-term consequences for ice-sheet dynamics and feedbacks to Arctic climate.

Carbon stored in permafrost constitutes another potentially large positive feedback to warming. There is at least as much carbon stored in permafrost as in the Earth's entire atmosphere (Zimov *et al.* 2006), and thawing permafrost releases substantial CH_4 (a greenhouse gas with an effect that is 25 times more potent per molecule than CO_2) from enlarging thaw lakes (Walter *et al.* 2006). However, it is difficult to predict how quickly CO_2 and CH_4 might be released from thawing permafrost to the atmosphere.

In summary, observed temperature trends demonstrate that the net effect of climate warming at high latitudes is a positive regional feedback to climate warming. This occurs primarily because the reductions in sea-ice extent and seasonal snow cover cause a decline in albedo and, therefore, an increase in energy absorption and heat transfer to the atmosphere. Expected future changes in Arctic vegetation and thawing of permafrost will probably magnify high-latitude warming (a positive feedback), whereas changes in boreal forest cover associated with increases in fire and insect outbreaks may have the opposite effect (a negative feedback). Although the net effect of these high-latitude changes in plant functional types is uncertain, they are unlikely to negate the strong positive feedback from declining snow-ice albedo, resulting in continued high-latitude warming. The human activity that most strongly contributes to this warming is emission of greenhouse gases (IPCC 2007a), suggesting that reductions in these emissions would be the most effective way to reduce climate warming.

Changing arid lands

Just as ecosystem feedbacks have amplified rates of climate warming at high latitudes, ecosystem changes may have amplified the magnitude and duration of regional droughts in dry areas. For example, in the Sahel, a dry region south of the Sahara Desert, a drought that extended through the final three decades of the 20th century appeared to be initiated either by changing sea-surface temperatures in the adjacent Atlantic Ocean, or by land degradation in the Sahel region due to overgrazing (Foley *et al.* 2003a). However, although patterns of ocean circulation or overgrazing appear to have *triggered* the drought, they cannot readily explain its 30-year duration (Foley *et al.* 2003a), which greatly exceeds the normal length of droughts in other parts of the world.

Millions of people died from drought-associated famine and disease over the 30-year duration of the drought – an example of the tragic societal consequences of changes in climate–ecosystem feedbacks. It is thought that the initial drought caused a substantial reduction in plant cover, extending beyond the lands already overgrazed by local herders. As Sahel vegetation declined, albedo increased and evapotranspiration decreased; the associated decline in energy and moisture transfer to the atmosphere further reduced convective uplift and associated monsoon rains in the region (Zeng *et al.* 1999; Wang and Eltahir 2000; Foley *et al.* 2003a; Figure 3). Although the drought in the Sahel was probably triggered by changes in ocean circulation and/or land-use practices, ecosystem feedbacks apparently contributed to its magnitude and extended duration.

Even localized changes in land cover and albedo can trigger hydrologic changes. In Western Australia, for example, extensive areas of heathland were cleared for wheat. The native heathlands had a lower albedo and therefore greater convective uplift, causing movement of moist air from the wheatlands toward the heathlands. This generated a local pattern of air circulation similar to that described for the Sahel, with a 10% increase in precipitation over the heathlands and a 30% decrease in precipitation over the wheatlands (Chambers 1998; Figure 5). A more heterogeneous pattern of heath and wheat would probably have prevented this local change in precipitation.

In summary, local and regional droughts may be amplified by ecosystem feedbacks that decrease vegetation cover, energy absorption, and convective uplift. This is an important issue: desertification is now widespread in arid regions of the world, due to both climatic and social changes (Reynolds and Stafford Smith 2002; Safriel *et al.* 2005). To what extent might these human influences contribute to drought? Alternatively, to what extent might afforestation (the deliberate planting of new forest tracts) or other efforts to increase vegetation cover reduce the likelihood of continued drought? The apparent importance of ecosystem feedbacks suggests that appropriate management could potentially reduce the likelihood of extended droughts.

S Chambers

Figure 4. *Recently burned Alaskan black spruce forest in spring, showing high albedo of the snow-covered surface.*

Changing forest cover

Complex social and economic forces contribute to large-scale changes in forest cover throughout the world (Shvidenko *et al.* 2005). Forests are being cleared in the tropics and in boreal Canada and Siberia, but are regrowing on abandoned agricultural lands in Europe and temperate North America, altering feedbacks to the climate sys-

Panel 1. Effects of human activities on ecosystem–climate feedbacks (see Figure 3)

Human activities alter several pathways of ecosystem–climate feedbacks (Figure 3). Elevated CO_2 affects ecosystem feedbacks to the climate system through increased photosynthesis, which removes CO_2 from the atmosphere; reduced albedo, which increases energy transfer to the ecosystem; and reduced stomatal conductance, which reduces the surface cooling effect of transpiration.

High-latitude climate warming affects ecosystem feedbacks to the climate system through melting of sea ice, earlier snowmelt, and shrub and forest expansion, all of which reduce albedo and increase energy transfer to the ecosystem (and, subsequently, to the atmosphere). Warming also increases decomposition, risk of wildfire and insect outbreaks, and permafrost thaw, all of which release CO_2 to the atmosphere. These effects are partially offset by photosynthetic CO_2 uptake, associated with greater plant growth in the Arctic (but not the boreal forest).

Drought affects ecosystem feedbacks to the climate system through reduced strength of the monsoon. The drought-induced decline in vegetation increases albedo, which reduces convective uplift, marine moisture advection, and strength of the monsoon. Together, these factors tend to maintain drought. Overgrazing can aggravate these effects by reducing vegetation, which strengthens the albedo-induced decline in the monsoon.

Deforestation alters ecosystem feedbacks to the climate system through increased albedo, which reduces energy transfer to the ecosystem (and, subsequently, to the atmosphere); reduced transpiration, which reduces moisture transport to the atmosphere; and net CO_2 release, which increases the heat-trapping capacity of the atmosphere. The balance among these contrasting climate feedbacks is poorly known, although the albedo (cooling) effects are strongest at high latitudes, and the moisture and carbon-balance feedbacks (warming) are strongest in the tropics. Preventing deforestation or expanding forest cover would have the opposite effects: warming through reduced albedo (strongest at high latitudes) and cooling through increased transpiration and carbon sequestration (strongest in the tropics).

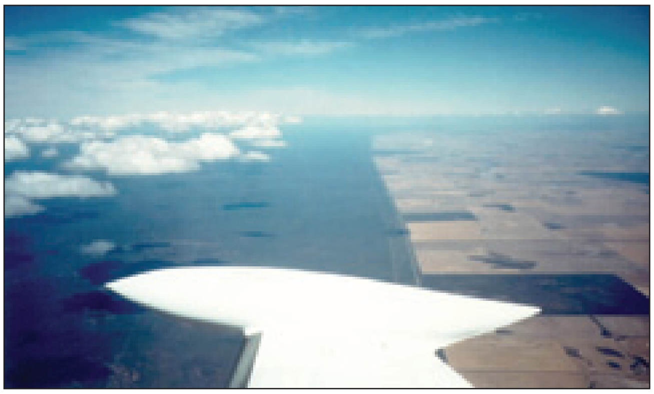

Figure 5. *Boundary between heathland and wheat croplands in southwestern Australia (Chambers 1998). The heathland absorbs more radiation (low albedo) and transmits a larger proportion of this energy to the atmosphere as sensible heat than do adjacent croplands. This causes air to rise over the heathland and draws in moist air laterally from the irrigated cropland, which causes subsidence of air over the cropland. Rising moist air forms clouds and increases precipitation by 10% over heathland, whereas subsiding dry air reduces precipitation by 30% over the cropland.*

tem at local to global scales. In recognition of these climate feedbacks, the Kyoto Protocol allows nations to use afforestation or reforestation (regrowth of previously existing forests) to meet some of their commitments to reduce the net release of CO_2 to the atmosphere (a cooling effect on climate). A consideration of the interactions of multiple climate feedbacks suggests that more forest area or larger trees do not always result in a cooler climate.

Logging in the boreal forest reduces the quantity of carbon remaining in the forest. However, logging effects on carbon budgets depend on the rate at which forest products decompose (whether they are used to produce paper or structural timber, for example), the rate of forest regrowth, and the release of CO_2 and CH_4 from thawing permafrost or newly formed thaw lakes and wetlands (McGuire *et al.* 2006; Walter *et al.* 2006; Zimov *et al.* 2006). Reforestation to offset these carbon losses (ie to slow global climate warming through carbon storage) also has a regional warming effect on climate, through reduced albedo (replacement of a reflective snow surface by an energy-absorbing forest canopy), just as in succession after boreal fire (Figure 3). The net effect of these climate feedbacks has not been quantified, except through modeling studies, but increases in forest cover will certainly provide less climate mitigation (and perhaps even warming; Bala *et al.* 2007) than would be predicted based solely on their effect on atmospheric CO_2.

The balance between the warming and cooling effects of changes in forest cover varies with latitude and ecosystem type. The potential cooling effects of deforestation through increased albedo are most pronounced in the boreal forest, due to the striking contrast in albedo between forests and an unforested snow surface (Euskirchen *et al.* 2007). Changes in water balance predominate in the tropics, where a combination of high solar input to evaporate water, high soil moisture availability, and a long photosynthetic season combine to drive large water fluxes from ecosystems to the atmosphere. In the Amazon, for example, 20–30% of annual precipitation derives from water that is recycled by evapotranspiration (Costa and Foley 1999). Model simulations suggest that extensive deforestation could lead to a drier, more savanna-like climate that would be less suitable for reforestation (Shukla *et al.* 1990; Foley *et al.* 2003b). Warm temperatures in the tropics probably also cause more rapid loss of soil carbon after logging than at high latitudes, augmenting the warming impacts of tropical deforestation.

In summary, the net warming feedback to climate from deforestation is strongest in the tropics (Bala *et al.* 2007). Conversely, planting of new forests is progressively more effective in reducing the potential for climate warming as we move from cold to warmer, wetter climates (Figure 6). Thus, reducing rates of deforestation and fostering reforestation in the tropics may be particularly effective in mitigating climate warming (Gullison *et al.* 2007).

Socially mediated feedbacks between ecosystems and climate

We have shown that ecosystem feedbacks play an important role in modulating the effects of ecosystem or land-use change on the climate system. However, ecosystem changes in the coming decades will probably involve an additional layer of interaction, as society responds to climate change by altering land use or land cover in ways that feed back to further influence climate. For example, as discussed earlier, droughts in semi-arid zones can interact with social changes to trigger desertification, which increases albedo, weakens moisture-bearing monsoons, and predisposes semi-arid climate systems to drought. Other important classes of socially mediated climate feedbacks include the sensitivity of tropical deforestation to changes in the hydrologic cycle, shifts in temperate and boreal agriculture in response to drought and warming, and changes in trade and fossil-fuel consumption caused by climate-induced changes in infectious disease, human migration, and civil strife.

The lengthening of the growing season in the southern boreal forest may be a stimulus for widespread, human-driven conversion of forest to grassland and cropland in southeastern Siberia. This loss of forest cover might counteract some warming, with large changes in albedo during spring and summer serving as a negative feedback to regional warming.

More generally, many classes of land use are sensitive to climate and are likely to have important consequences for the evolution of atmospheric CO_2, air temperature, and water cycling over the next several centuries. We are only now beginning to understand how the ecosystem/climate feedbacks discussed here may interact with climate feedback processes (Mölders and Kramm 2007). For example, we cannot yet identify thresholds at which these interactions change dramatically, and we cannot calculate the probability that a regionally focused climate change in one place (for example, from albedo effects associated with deforestation) might have teleconnections to changes in temperature or precipitation in other places. Still, climate–ecosystem feedbacks are understood better than climate–land-use–ecosystem feedbacks. The latter are complicated not only by the expanded set of decisions they involve, but also by the prospect that the climate signal initiating some of these feedbacks will weaken or strengthen the motivation for action to mitigate climate change. These climate–land-use–ecosystem feedbacks are completely absent from state-of-the-art climate models and are represented only crudely or not at all in Earth-system and integrated-assessment models.

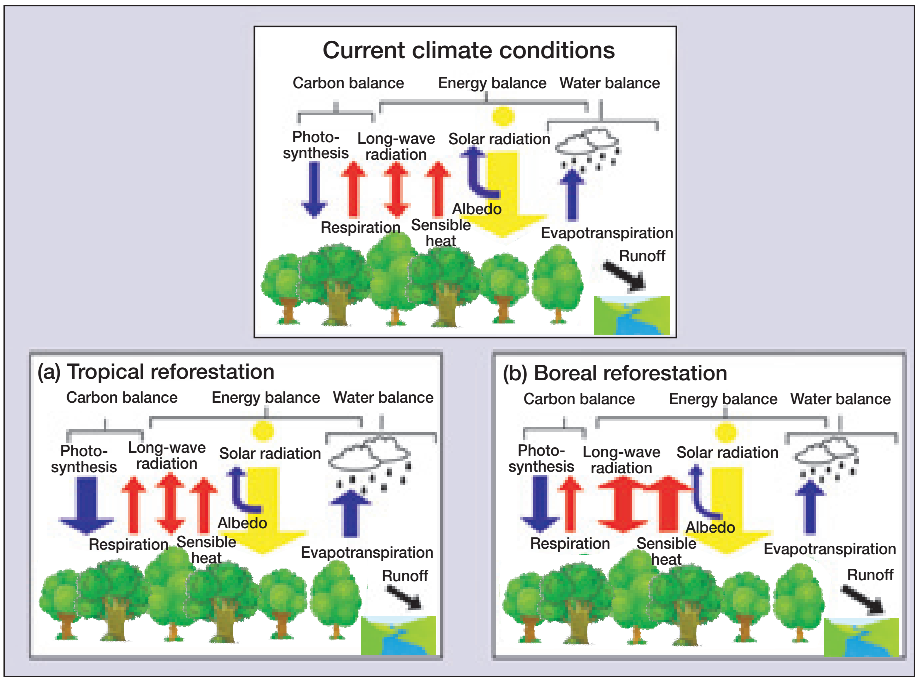

Figure 6. *Effects of (a) tropical and (b) boreal reforestation on climate feedbacks through changes in carbon, energy, and water budgets. (See Figure 2 for explanation of arrows.)*

Representing socially mediated feedbacks in climate system models is challenging, in part because many of the elements are only weakly coupled. For example, other economic, social, and political factors are likely to be more important than climate in shaping future trajectories of land cover (Ellis and Ramankutty 2008). Nevertheless, the carbon losses, albedo changes, and aerosol impacts caused by the sensitivity of land-use change to climate could exceed the magnitude of all the direct climate–ecosystem linkages described above.

Conclusions

Five major messages emerge from this paper:

(1) Ecosystems influence climate through multiple pathways, so efforts to mitigate climate through consideration of only one of these pathways, as with carbon in the Kyoto Protocol, are incomplete. Consideration of multiple feedbacks could lead to climate-mitigation strategies that provide more consistent mitigation of climate change, but will require careful quantification of these feedbacks and their interactions.

(2) There is no "one-size-fits-all" solution for mitigating warming through managing ecosystems. Practices that effectively mitigate climate warming in the tropics may prove less effective or counterproductive in other regions.

(3) Industrial greenhouse-gas emissions that contribute to climate warming originate primarily from industrialized nations in the temperate zone (Raupach *et al.* 2007), but some of the most promising strategies for ecosystem-based mitigation involve reducing deforestation in tropical developing nations or reducing desertification of arid zones. Although there are many opportunities for reducing these emissions, some of the lowest cost, highest benefit options for mitigation of climate change (eg sustaining biodiversity) come from reducing tropical deforestation (IPCC 2007b). The challenge is to find an equitable approach for distributing the attendant social and economic costs among nations.

(4) Climate change is a much broader issue than temperature. Associated changes in the hydrologic cycle often have even greater societal impacts through changes in precipitation, evapotranspiration, runoff, and water available for human use.

(5) Although our understanding of the multiple feedbacks between climate and ecosystems is far from complete and requires careful quantification, it is clear that ecosystems cannot "solve the climate problem" by removing all the CO_2 produced by fossil-fuel emissions. The ultimate solution to this problem may require major reductions in emissions from fossil-fuel combustion. Given the highly non-linear nature of feedbacks between ecosystems and climate, the benefits of reduced emissions will probably be much greater if implemented soon rather than in the distant future.

Acknowledgements

We thank D Baldocchi, E Euskirchen and J Walsh for their constructive comments on the manuscript. Research leading to this synthesis was supported in part by Bonanza Creek LTER (Long Term Ecological Research) program (funded jointly by NSF grant DEB-0620579 and USDA Forest Service, Pacific Northwest Research Station grant PNW06-JV-11261952-431).

References

Bala G, Caldeira K, Wickett M, *et al.* 2007. Combined climate and carbon-cycle effects of large-scale deforestation. *P Natl Acad Sci USA* **104**: 6550–55.

Callaghan T, Bjorn LO, Chernov Y, *et al.* 2004. Effects on the function of Arctic ecosystems in the short- and long-term perspectives. *Ambio* **33**: 448–58.

Canadell JG, Le Quéré C, Raupach MR, *et al.* 2007. Contributions to accelerating atmospheric CO_2 growth from economic activity, carbon intensity, and efficiency of natural sinks. *P Natl Acad Sci USA* **104**: 10288–93.

Caspersen JP, Pacala SW, Jenkins JC, *et al.* 2000. Contributions of land-use history to carbon accumulation in US forests. *Science* **290**: 1148–51.

Chambers S. 1998. Short- and long-term effects of clearing native vegetation for agricultural purposes. Adelaide, Australia: Flinders University of South Australia.

Chapin III FS, Sturm M, Serreze MC, *et al.* 2005. Role of land-surface changes in Arctic summer warming. *Science* **310**: 657–60.

Costa MH and Foley JA. 1999. Trends in the hydrological cycle of the Amazon basin. *J Geophys Res* **104**: 14189–98.

Drake BG, Gonzalez-Meler MA, and Long SP. 1997. More efficient plants: a consequence of rising atmospheric CO_2? *Annu Rev Plant Phys* **48**: 607–40.

Ellis EC and Ramankutty N. 2008. Putting people on the map: anthropogenic biomes of the world. *Front Ecol Environ* **6**. doi: 10.1890/070062.

Euskirchen SE, McGuire AD, and Chapin III FS. 2007. Energy feedbacks to the climate system due to reduced high latitude snow cover during 20th century warming. *Glob Change Biol* **13**: 2425–38.

Field CB, Jackson RB, and Mooney HA. 1995. Stomatal responses to increased CO_2: implications from the plant to the global scale. *Plant Cell Environ* **18**: 1214–25.

Field CB, Lobell DB, Peters HA, *et al.* 2007. Feedbacks of terrestrial ecosystems to climate change. *Annu Rev Environ Resour* **32**: 1–29.

Flanner MG, Zender CS, Randerson JT, *et al.* 2007. Present-day climate forcing and response from black carbon in snow. *J Geophys Res-Atmos* **112**: D11202. doi:10:1029/2006JD008003.

Foley JA, Coe MT, Scheffer M, *et al.* 2003a. Regime shifts in the Sahara and Sahel: interactions between ecological and climatic systems in northern Africa. *Ecosystems* **6**: 524–39.

Foley JA, Costa MH, Delire C, *et al.* 2003b. Green surprise? How terrestrial ecosystems could affect Earth's climate. *Front Ecol Environ* **1**: 38–44.

Friedlingstein P, Cox P, Betts R, *et al.* 2006. Climate–carbon cycle feedback analysis: results from the C^4MIP model intercomparison. *J Climate* **19**: 3337–53.

Gitz V and Ciais P. 2003. Amplifying effects of land-use change on future atmospheric CO_2 levels. *Global Biogeochem Cy* **17**: 1024. doi:10.29/2002GB001963.

Goetz SJ, Bunn AG, Fiske GA, *et al.* 2005. Satellite-observed photosynthetic trends across boreal North America associated with climate and fire disturbance. *P Natl Acad Sci USA* **102**: 13521–25.

Gullison RE, Frumhoff PC, Canadell JG, *et al.* 2007. Tropical forests and climate policy. *Science* **316**: 985–86.

IPCC (Intergovernmental Panel on Climate Change). 2007a. Climate change 2007: the physical science basis. Contribution of Working Group I to the fourth assessment report of the Intergovernmental Panel on Climate Change. Cambridge, UK: Cambridge University Press.

IPCC (Intergovernmental Panel on Climate Change). 2007b. Summary for policymakers. In: Metz B, Davidson OR, Bosch PR, *et al.* (Eds). Climate change 2007: mitigation. Contribution of Working Group III to the fourth assessment report of the Intergovernmental Panel on Climate Change. Cambridge, UK: Cambridge University Press.

Körner C, Asshoff R, Bignucolo O, *et al.* 2005. Carbon flux and growth in mature deciduous forest trees exposed to elevated CO_2. *Science* **309**: 1360–62.

Lobell DB and Field CB. 2008. Estimation of the carbon dioxide (CO_2) fertilization effect using growth rate anomalies of CO_2 and crop yields since 1961. *Glob Change Biol* **14**: 39–45.

Long SP, Ainsworth EA, Leakey ADB, *et al.* 2006. Food for thought: lower-than-expected crop yield stimulation with rising CO_2 concentrations. *Science* **312**: 1918–21.

Matthews HD, Eby M, Ewen T, *et al.* 2007. What determines the magnitude of carbon cycle–climate feedbacks? *Global Biogeochem Cy* **21**: GB2012. doi:10.1029/2006GB002733.

McConnell JR, Edwards R, Kok GL, *et al.* 2007. 20th-century industrial black carbon emissions altered Arctic climate forcing. *Science* **317**: 1381–84.

McGuire AD, Chapin III FS, Walsh JE, *et al.* 2006. Integrated regional changes in Arctic climate feedbacks: implications for the global climate system. *Annu Rev Environ Resour* **31**: 61–91.

McGuire AD, Clein JS, Melillo JM, *et al.* 2000. Modeling carbon responses of tundra ecosystems to historical and projected climate. The sensitivity of pan-Arctic carbon storage to temporal and spatial variation in climate. *Glob Change Biol* **6**: 141–59.

Mölders N and Kramm G. 2007. Influence of wildfire induced land-cover changes on clouds and precipitation in interior Alaska – a case study. *Atmos Res* **84**: 142–68.

Norby RJ, DeLucia EH, Gielen B, *et al.* 2005. Forest response to elevated CO_2 is conserved across a broad range of productivity. *P Natl Acad Sci USA* **102**: 18052–56.

Randerson JT, Liu H, Flanner M, *et al.* 2006. The impact of boreal forest fire on climate warming. *Science* **314**: 1130–32.

Raupach MR, Marland G, Ciais P, *et al.* 2007. Global and regional drivers of accelerating CO_2 emissions. *P Natl Acad Sci USA* **104**: 10288–93.

Reynolds JF and Stafford Smith DM (Eds). 2002. Global desertification: do humans cause deserts? Berlin, Germany: Dahlem University Press.

Safriel U, Adeel Z, Niemeijer D, *et al.* 2005. Dryland systems. In: Hassan R, Scholes R, and Ash N (Eds). Millennium Ecosystem Assessment: ecosystems and human well-being. Washington, DC: Island Press.

Sellers PJ, Bounoua L, Collatz GJ, *et al.* 1996. Comparison of radiative and physiological effects of doubled atmospheric CO_2 on climate. *Science* **271**: 1402–06.

Shukla J, Nobre C, and Sellers P. 1990. Amazon deforestation and climate change. *Science* **247**: 1322–25.

Shvidenko A, Barber DV, Persson R, *et al.* 2005. Forest and woodland systems. In: Hassan R, Scholes R, and Ash N (Eds). Millennium Ecosystem Assessment: ecosystems and human well-being – current state and trends. Cambridge, UK: Cambridge University Press.

Sitch S, McGuire AD, Kimball J, *et al.* 2007. Assessing the carbon balance of circumpolar Arctic tundra using remote sensing and process modeling. *Ecol Appl* **17**: 213–34.

Steffen WL, Sanderson A, Tyson PD, *et al.* 2004. Global change and the Earth system: a planet under pressure. New York, NY: Springer-Verlag.

Walter MK, Zimov SA, Chanton JP, *et al.* 2006. Methane bubbling from Siberian thaw lakes as a positive feedback to climate warming. *Nature* **443**: 71–75.

Wang G and Eltahir EAB. 2000. Ecosystem dynamics and the Sahel drought. *Geophys Res Lett* **27**: 95–98.

Zeng N, Neelin JD, Lau KM, *et al.* 1999. Enhancement of interdecadal climate variability in the Sahel by vegetation interaction. *Science* **286**: 1537–40.

Zimov SA, Schuur EAG, and Chapin III FS. 2006. Permafrost and the global carbon budget. *Science* **312**: 1612–13.

