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The influence of land-use change and landscape dynamics on the climate system: relevance to climate-change policy beyond the radiative effect of greenhouse gases

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Our paper documents that land-use change impacts regional and global climate through the surface-energy budget, as well as through the carbon cycle. The surface-energy budget effects may be more important than the carbon-cycle effects. However, land-use impacts on climate cannot be adequately quantified with the usual metric of 'global warming potential'.

A new metric is needed to quantify the human disturbance of the Earth's surface-energy budget. This 'regional climate change potential' could offer a new metric for developing a more inclusive climate protocol. This concept would also implicitly provide a mechanism to monitor potential local-scale environmental changes that could influence biodiversity.

Keywords: global climate change; regional climate change; landscape change; landscape dynamics; climate-system dynamics

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1. Introduction

Policy-related quantification of human influences on climate has focused largely on changes in atmospheric composition. However, a large body of work has demonstrated that land-cover change provides an additional major forcing of climate, through changes in the physical properties of the land surface. The global radiative forcing by surface albedo change may be comparable with that due to anthropogenic aerosols, solar variation and several of the greenhouse gases. Moreover, in regions of intensive human-caused land-use change such as North America, Europe and southeast Asia, the local radiative-forcing change caused by surface albedo may actually be greater than that due to all the well-mixed anthropogenic greenhouse gases together (IPCC 2001).

Surface albedo change can be compared with greenhouse-gas emissions through the concept of radiative forcing (Betts 2000), but changes in vegetation cover can also modify the surface heat fluxes directly. This cannot be quantified in terms of radiative forcing, so a full quantification of land-use impacts on climate requires a new approach. Furthermore, as well as influencing local long-term weather conditions, regional-scale land-cover change can impact on the global climate system through teleconnections (Avissar 1995; Pielke 2001*a*; Claussen 2002). Remote changes in different locations may be of opposing sign, so spatial averaging may under represent the true global significance of the land-use effects.

These aspects of human influence on climate are not currently accounted for under the Kyoto Protocol. One reason for this may be the difficulty in objectively comparing the effects of different local land-surface changes with each other and with the effects of changing atmospheric composition. However, the neglect of land-use effects will lead to inaccurate quantification of contributions to climate change, with the danger that some actions may give unintended and counterproductive results. It is therefore important that possible metrics for land-use effects are explored. Here we discuss some approaches to this problem.

2. Historical land-use change

A documentation of global patterns of land-use change from 1700 to 2000 is presented in Klein Goldewijk (2001). Klein Goldewijk reports on worldwide changes of land to crops of 136, 412 and 658 Mha in the periods 1700–1799, 1800–1899 and 1900–1990, respectively. Conversion to pasture was 418, 1013 and 1496 Mha in these three time periods. Figure 1 illustrates these changes, including an acceleration of tropical deforestation during the 20th century. O'Brien (2000) also documents land-use change for recent years (table 1).

Apart from their role as reservoirs, sinks, and sources of carbon, tropical forests provide numerous additional 'ecosystem services'. Many of these ecosystem services directly or indirectly influence climate. The climate-related ecosystem services that tropical forests provide include the maintenance of elevated soil moisture and surface air humidity, reduced sunlight penetration, weaker near-surface winds and the inhibition of anaerobic soil conditions. Such an environment maintains the productivity of tropical ecosystems (Betts 1999) and has helped sustain the rich biodiversity of tropical forests.

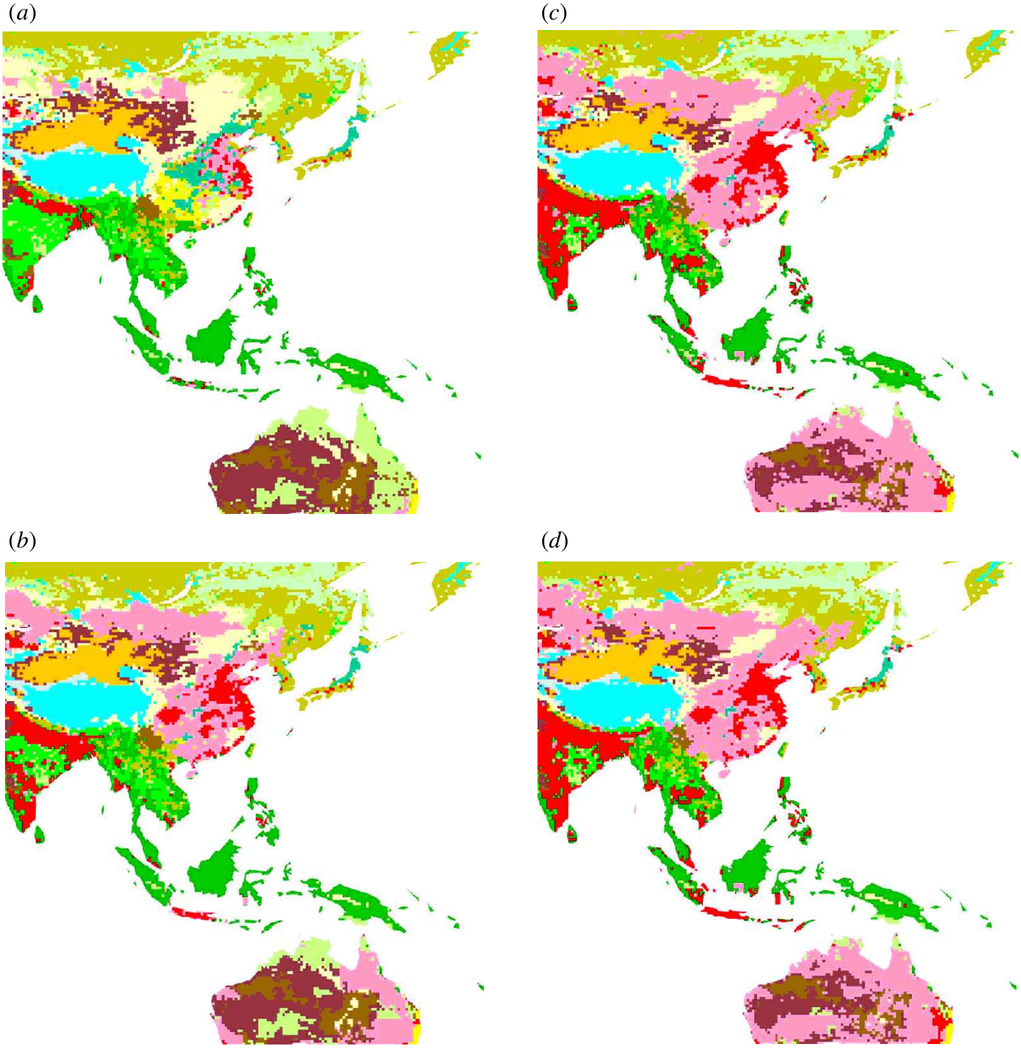

Figure 1. Examples of land-use change from (*a*) 1700, (*b*) 1900, (*c*) 1970 and (*d*) 1990. The human-disturbed landscape includes intensive cropland (red) and marginal cropland used for grazing (pink). Other landscape includes, for example, tropical evergreen and deciduous forest (dark green), savannah (light green), grassland and steppe (yellow), open shrubland (maroon), temperate deciduous forest (blue), temperate needleleaf evergreen forest (light yellow) and hot desert (orange). Of particular importance in this paper is the expansion of the cropland and grazed land between 1700 and 1900. (Reproduced with permission from Klein Goldewijk 2001).

3. Impacts of land-cover change on climate

The significant role of the land within the climate system should not be surprising. As discussed by Wu & Newell (1998) for El Niño events, warming of a relatively small area in the tropical eastern and central Pacific Ocean has global climate consequences. This occurs because tropical cumulonimbus clouds occur in this region during an El Niño event, and not during average ocean conditions. These deep cumulus

Table 1. *Tropical forest extent and loss* (*rainforest and moist deciduous forest ecosystems*) (Source: World Resources Institute (1994), adapted from O'Brien (2001).)

country	rainforest extent in 1990 (kha)	decrease 1981–1990 (%)	moist deciduous forest extent in 1990 (kha)	decrease 1981–1990 (%)
Brazil	291 597	6	197 082	16
Indonesia	93 827	20	3 366	20
Dem. Rep. Congo (Zaire)	60 437	12	45 209	12
Columbia	47 455	6	4 101	38
Peru	40 358	6	12 299	6
Papua New Guinea	29 323	6	705	6
Venezuela	19 602	14	15 465	36
Malaysia	16 339	36	0	0
Myanmar	12 094	24	10 427	28
Guyana	11 671	0	5 078	6
Suriname	9 042	0	5 726	4
India	8 246	12	7 042	10
Cameroon	8 021	8	9 892	12
French Guiana	7 993	0	3	0
Congo	7 667	4	12 198	4
Ecuador	7 150	34	1 669	34
Lao People's Dem. Rep.	3 960	18	4 542	18
Philippines	3 728	62	1 413	54
Thailand	3 082	66	5 232	54
Vietnam	2 894	28	3 382	28
Guatemala	2 542	32	731	0
Mexico	2 441	20	11 110	30
Belize	1 741	0	238	0
Cambodia	1 689	20	3 610	20
Gabon	1 155	12	17 080	12
Central African Republic	616	14	28 357	8
Cuba	114	18	1 247	18
Bolivia	0	0	35 582	22

clouds permit the export of heat, moisture and kinetic energy to higher latitudes that do not normally receive such tropical export. Wu & Newell concluded that the long persistence, the spatial coherence of the ocean warming and its large magnitude are the reasons for this major role of El Niño events within the Earth's climate system.

Tropical land-use change has been shown in Chase *et al.* (1996, 2000), and summarized in Claussen (2002), to have an effect on the climate system similar to that from an El Niño event. Since thunderstorms preferentially form over land (Lyons 1999), the role of the tropical land surface should be expected to have a greater

effect on global climate than implied by its per cent areal coverage of the Earth's surface alone. General circulation model (GCM) simulations by Chase *et al*. (2000) indicated that regional landscape change can result in alterations to surface fluxes elsewhere in the world through nonlinear feedbacks within the atmosphere's global circulation (figure 2).

The alteration of tropical landscapes, primarily the conversion of forests to agriculture or pasture, changes the partitioning of solar insolation into its sensible and latent turbulent heat forms. Less transpiration associated with the agricultural and pasture regions results in less thunderstorm activity over this landscape. Lawton *et al*. (2001), for example, illustrates, for Costa Rica, the significant regional effects that tropical deforestation has on the ecological environment of adjacent mountains. The longitudinal distribution of thunderstorms in the tropics is also changed.

Unlike an El Niño, however, where the system changes back to 'average' and La Niña conditions (over time, we have learned to identify the global impacts of the different situations), land-use change is often permanent, so its global effects are not as obvious. Further, the atmospheric feedback due to similar land-use changes are variable, depending on the geographical domain or the existing land use (Niyogi 2000; Niyogi *et al*. 2002*a*).

Deliberate land-use change (afforestation or reforestation) has been accepted as a mechanism to remove CO_2 from the atmosphere and sequester carbon in trees and soils. However, as discussed by Betts (2000) and Pielke (2001*b*), this activity may have other effects in terms of the radiative forcing in the atmosphere. For example, in regions subject to significant snow cover, afforestation would result in a lower surface albedo and hence a positive radiative forcing, resulting in a net warming effect despite the removal of CO_2 from the atmosphere (figure 3). Similarly, increases in the surface fluxes of water vapour could result in positive radiative forcing.

The biogeochemical effect of enhanced CO_2 and trace-gas concentration, and of aerosol deposition (such as nitrogen), on landscape dynamics has also not been adequately considered. For example, Jenkinson *et al*. (1991) demonstrated a significant positive feedback where soils released carbon to the atmosphere under warming conditions. More recently, Lenton (2000), using a simple box model, and Cox *et al*. (2000), using GCM-sensitivity experiments, showed that biogeochemical feedbacks in conjunction with an increased CO_2 radiative warming produced an amplified regional and global-warming response. Eastman *et al*. (2001*a*, *b*) used a regional climate model in a sensitivity study and suggested a cooler daytime and warmer nighttime in the central Great Plains in response to greater plant growth in a doubled-CO_2 atmosphere.

Niyogi *et al*. (2002*b*) used a coupled process-based model to show that the carbon-assimilation potential for each of the GCM land-use categories (comprising both C3 and C4 photosynthesis pathways) is sensitive to the soil-moisture availability. The presence of drought and hydrological feedbacks associated with land-use change locally or through teleconnections, therefore, has a direct impact on the source/sink capabilities of the terrestrial ecosystem.

These studies illustrate the significant role that biogeochemistry has within the climate system. This feedback, along with other climate forcings and feedbacks (Pielke 2001*b*), makes climate prediction on time-scales of years and longer a particularly difficult problem.

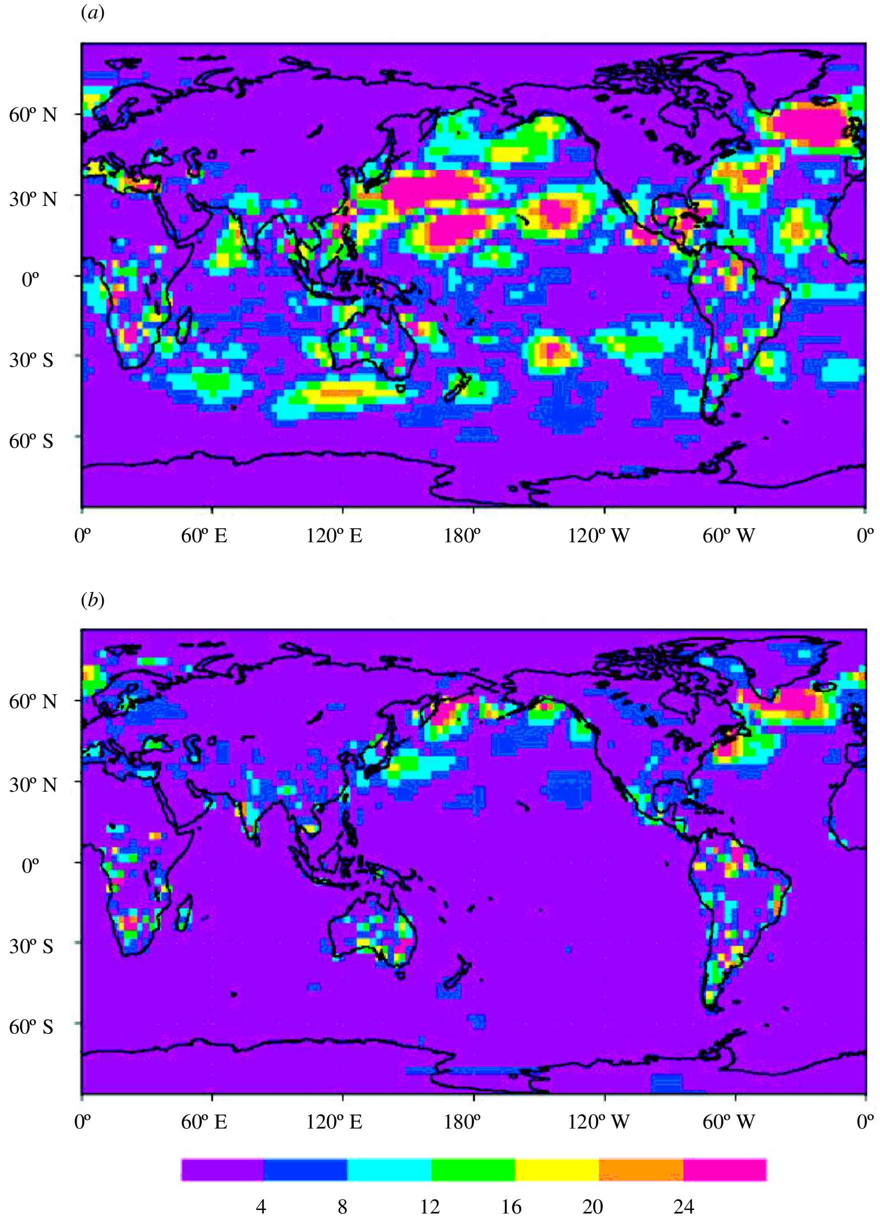

Figure 2. The 10-year average absolute-value change in surface latent turbulent heat flux in W m^{-2} worldwide as a result of the land-use changes: (*a*) January, (*b*) July. (Adapted from Chase *et al*. (2000).)

4. Quantifying land-use forcing of climate

As is evident in this issue, carbon has become the currency used to assess the human intervention in the Earth's climate system. Impacts on climate are compared in terms of radiative forcing, which can be considered as perturbations to the Earth's radiation budget prior to feedbacks from the rest of the climate system. The concept

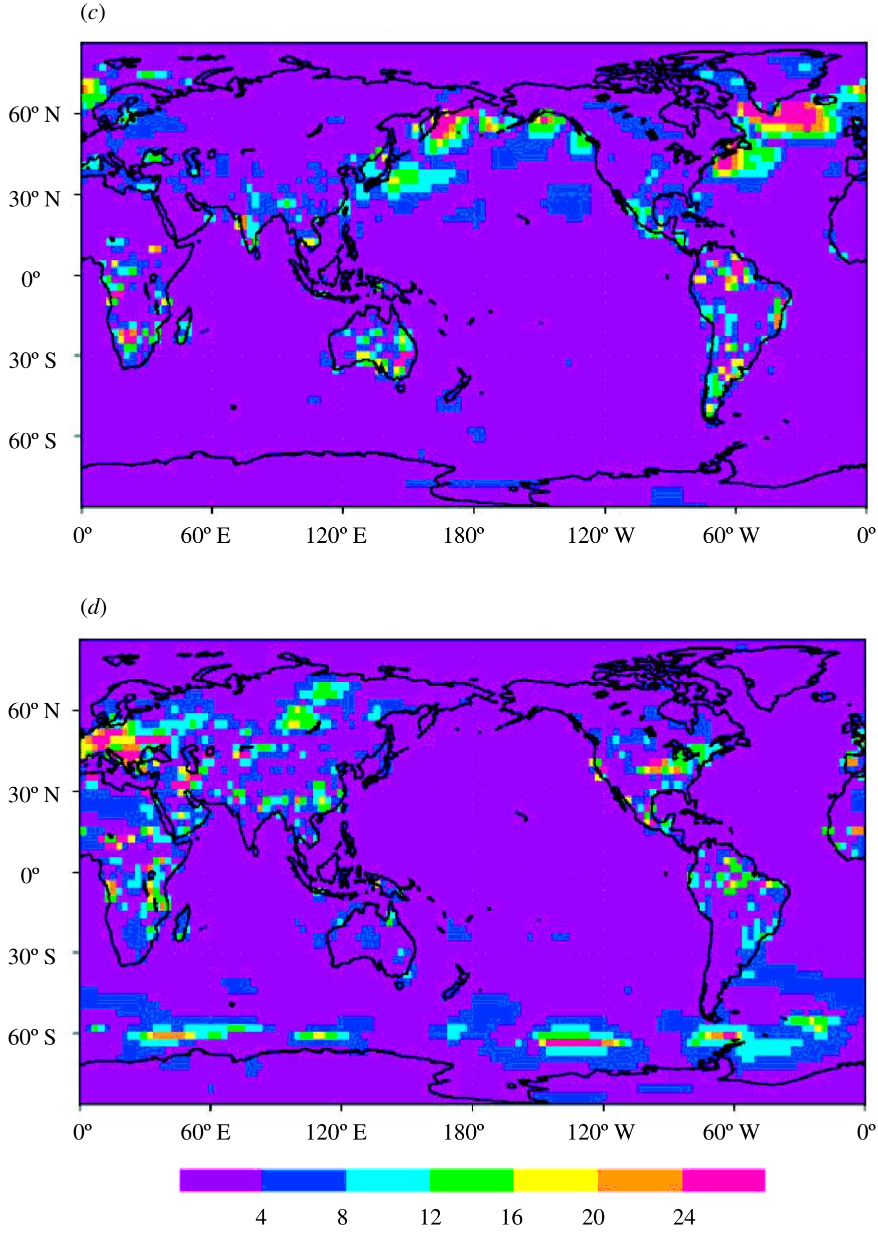

Figure 2. (*Cont.*) The 10-year average absolute-value change in sensible turbulent heat flux in W m^{-2} worldwide as a result of land-use changes: (*c*) January, (*d*) July. (Adapted from Chase *et al.* (2000).)

of global-warming potential (GWP), where

$$\mathrm{GWP} = \frac{\int_0^{\mathrm{TH}} a_x x(t)\,\mathrm{d}t}{\int_0^{\mathrm{TH}} a_r r(t)\,\mathrm{d}t}, \tag{4.1}$$

has been adopted to convert other atmospheric constituents into their equivalent in terms of CO_2 atmospheric radiative forcing (IPCC 2001). Here, 'TH' is the time period over which the calculation is considered, a_x is the radiative efficiency due to a unit change in atmospheric abundance of the substance x (i.e. $\mathrm{W\,m^{-2}\,kg^{-1}}$) and $x(t)$ is the time-dependent decay of the abundance from an instantaneous release of the substance. The denominator is the same expression, but for the reference substance r, defined to be CO_2.

The effects of land-surface albedo change can be quantified in terms of radiative forcing (Hansen *et al*. 1997; Betts 2001), and this has been used in attempts to compare the global significance of historical land-use change with that of other drivers of climate change (IPCC 2001). Betts (2000) suggested that radiative-forcing calculations could be used to translate albedo changes into equivalent carbon emissions (figure 3); this could be useful for quantifying land-use changes in regions where the main impact is on surface albedo, such as areas subject to significant snow cover. However, in other regions, changes in other land-surface properties may not exert a radiative forcing, but still significantly influence climate. For example, the partitioning of available energy into latent and sensible heat fluxes exerts a direct impact on near-surface air temperature, so a change in this partitioning should be considered a climate forcing. Radiative forcing, and hence GWP, are therefore not able to represent the full impact of land-cover change in all regions. Some new means of quantifying land-use forcing is therefore required.

Separation of the components of the surface-energy budget could provide a possible starting point, with the surface-heat energy being separated into

$$Q_{\mathrm{N}} + Q_{\mathrm{H}} + Q_{\mathrm{LE}} + Q_{\mathrm{G}} = 0, \tag{4.2}$$

where

$$Q_{\mathrm{N}} = Q_{\mathrm{S}}(1 - A) + Q^{\downarrow}_{\mathrm{LW}} - Q^{\uparrow}_{\mathrm{LW}}. \tag{4.3}$$

Here, Q_{N} is the net radiative flux, Q_{H} is the turbulent sensible heat flux, Q_{LE} is the turbulent latent heat flux (evaporation/transpiration), Q_{G} is the heat flux into the Earth's surface, Q_{S} is the solar irradiance, A is the surface albedo, $Q^{\downarrow}_{\mathrm{LW}}$ is the downward atmospheric irradiance and $Q^{\uparrow}_{\mathrm{LW}}$ is the upward surface irradiance. The magnitude of these fluxes can be expressed in units of watts per square metre or joules per unit of time (for example, a globally averaged value of $1\ \mathrm{W\,m^{-2}}$ is equal to 1.61×10^{23} J per decade, with $1\ \mathrm{W} = 1\ \mathrm{J\,s^{-1}}$). One measure of land-cover change forcing of climate could be the perturbation to one of the components of the surface-energy balance equation (4.2) prior to feedbacks from the rest of the climate system.

In the past, climate-change metrics have been concerned with globally averaged responses, as exemplified in the GWP. However, as discussed previously in this paper, global-scale climate changes can also occur due to land-use change, where Q_{H} and Q_{LE} are changed regionally but there is no change in the global average value. This occurred in the situation described by Chase *et al*. (1996, 2000), where there was no significant global-averaged change in these quantities (figure 2). Figure 4 illustrates the anthropogenically caused change in leaf area index (see Nemani *et al*. 1996), which were used in the Chase *et al*. studies. It was the spatial redistribution of the Q_{H} and Q_{LE} pattern that resulted in the global climate change. Globally averaged climate change may therefore bear no well-defined relation to the real changes experienced in any region and these regional changes, which can be of any sign, are what

impact on people and will stimulate mitigation strategies to be applied. Therefore, we identify a need for a surface 'regional climate-change potential' (RCCP), which addresses this deficiency. The RCCP would be defined to quantify where a direct human-caused change (either positive or negative) alters any of the individual terms in (4.2). These changes could be scaled by the surface area of the Earth to place them in a global context.

To provide a land-use-forcing term free of feedbacks, as in the case of radiative forcing, a land-surface scheme could be used to calculate perturbations to the surface-energy budget excluding feedbacks from the atmosphere. However, assessing the true global significance of these feedbacks is not trivial, since atmospheric circulation changes could give rise to remote climate changes that do not relate linearly to the perturbation in the region of land-use change (e.g. Gedney & Valdes 2000). To illustrate this point, the model results from Chase *et al.* (2000) presented in figure 2 show the 10-year average absolute-value differences in the surface-energy flux for January and July (i.e. Q_{LE} and Q_{H} from (4.2)) between the climate with the current landscape and the climate with potential landscape that should exist under current atmospheric conditions without human intervention. Figure 5 presents the 10-year average value difference in the surface-energy flux between the two experiments for just the locations where the human-caused landscape change was prescribed in the model.

The 10-year average changes for the locations where only the land cover was altered were 5.3 and 4.7 $\mathrm{W\,m^{-2}}$ in January and 8.8 and 6.7 $\mathrm{W\,m^{-2}}$ in July for surface sensible heat flux and latent heat flux, respectively. The globally weighted changes in the sum of the absolute values of the surface sensible and latent heat flux changes for figure 5 were 0.7 $\mathrm{W\,m^{-2}}$ in January and 1.08 $\mathrm{W\,m^{-2}}$ in July. With teleconnections changes also included (figure 2), however, the globally averaged changes in the surface fluxes were significantly larger, with values of 9.47 $\mathrm{W\,m^{-2}}$ in January and 8.90 $\mathrm{W\,m^{-2}}$ in July. These results clearly demonstrate that regional landscape change can result in alterations to surface fluxes elsewhere in the world through nonlinear feedbacks within the atmosphere's global circulation. If the other terms in the surface-heat budget (4.2) were included, the magnitude of the differences (i.e. the RCCP) would presumably be even larger.

A potential based purely on surface-flux perturbations would not be analogous to GWP, because the latter is an index requiring an expression in terms of the abundance of atmospheric constituents. Direct comparison of land-cover change effects with greenhouse-gas emissions therefore remains a challenge. One possibility might be to dispense with quantification in terms of forcings and compare anthropogenic influences by modelling whole-system responses to individual causes of climate change. For example, Claussen (2001) compared biogeophysical and biogeochemical effects of land-cover change in terms of the overall climate change (as opposed to the forcing). A number of simulations were performed with an intermediate-complexity Earth system model, in each run altering a small part of the land surface and simulating its effects on global mean temperature (through changes in surface properties and carbon exchange). However, as stated above, global mean temperature changes may conceal important local and regional changes, so applications of this type of modelling approach again require a global quantification of absolute regional changes.

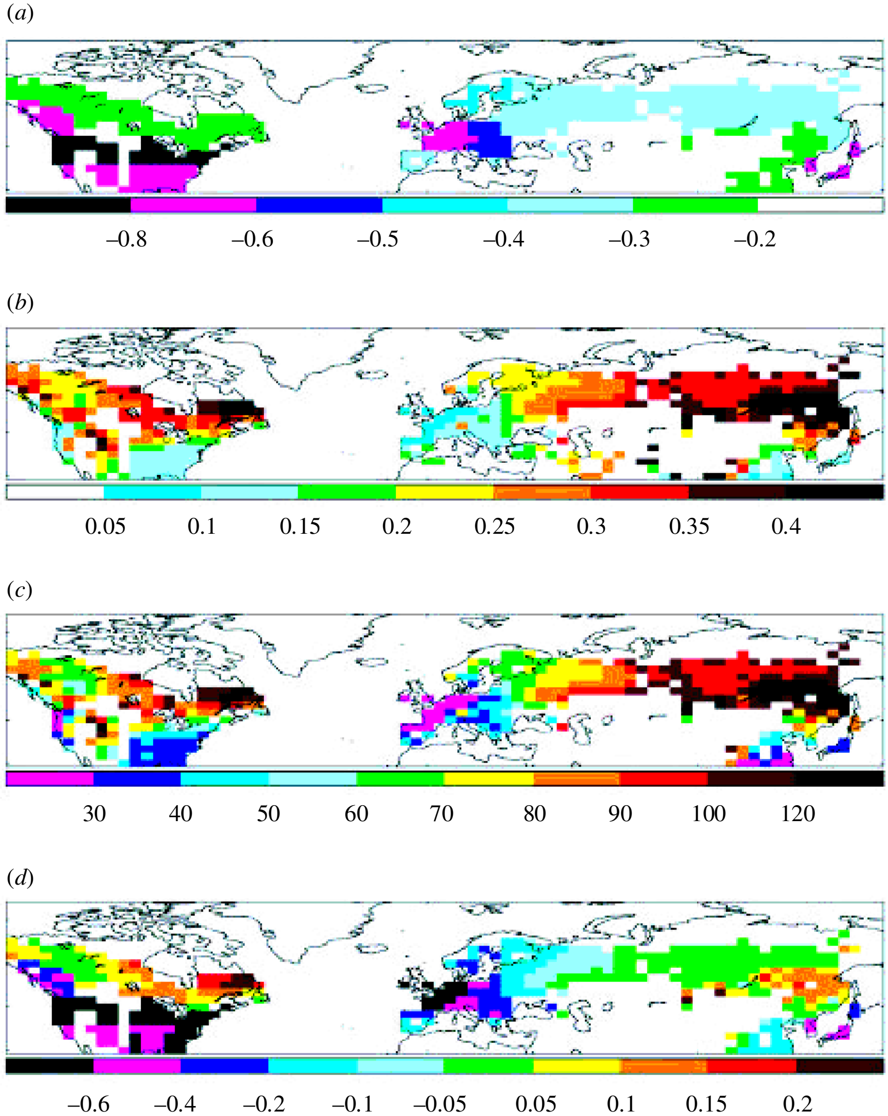

Figure 3. Radiative forcing of climate by afforestation, considering illustrative 1 ha plantations in the temperate and boreal forest zones. Calculations apply to the time at the end of one forestry rotation period, relative to the start of the rotation period with plantation areas unforested. (*a*) Global mean longwave radiative forcing due to CO_2 removal through sequestration (nW m^{-2} ha^{-1}). (*b*) Global mean shortwave radiative forcing due to albedo reduction (nW m^{-2} ha^{-1}). (*c*) Carbon emissions that would give the same magnitude of radiative forcing as the albedo reduction (tC ha^{-1}). (*d*) Net radiative forcing due to afforestation, found by summing (*a*) and (*b*) (nW m^{-2} ha^{-1}). Positive forcing implies a warming influence; where (*d*) shows positive values, afforestation would warm climate rather than cooling it, as would be expected by considering carbon sequestration alone (Betts 2000).

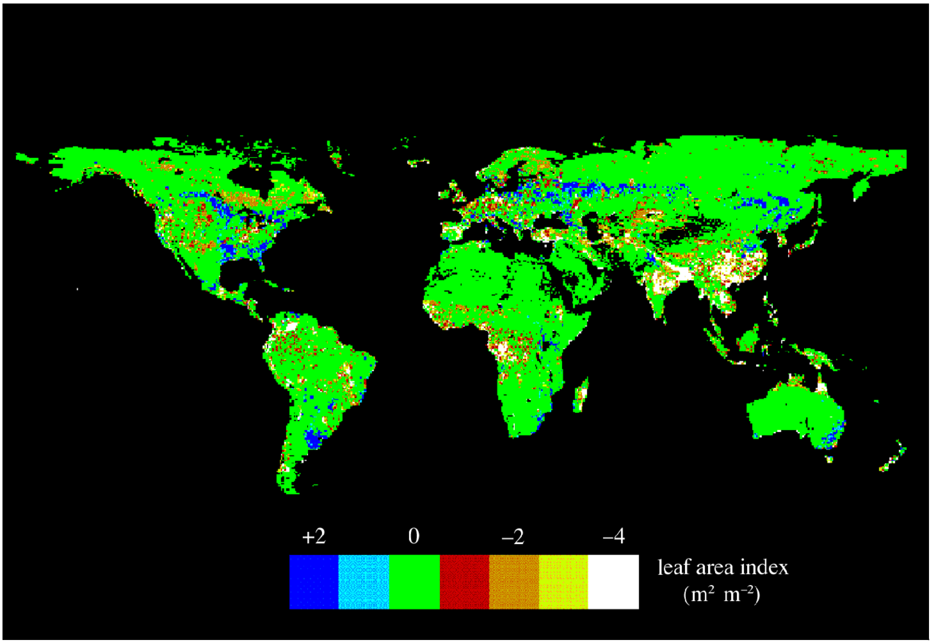

Figure 4. Effect of land-use changes on plant-canopy density (potential/actual). Scale 0.5 latitude $\times$ 0.5 longitude.

5. Conclusions

Atmospheric and ocean circulation patterns and their subsequent involvement within the planet's climate are dynamic, variable and difficult to predict. This limits the ability to predict the impact of land-use change and landscape dynamics on climate patterns. As a result, manipulating land-surface conditions for the purpose of carbon sequestration under the Kyoto Protocol could have a variety of unanticipated impacts on global and regional climate. The Kyoto Protocol uses only the GWPs of the regulated greenhouse-gas molecules listed in its Annex A as its mitigation currency. A more complete indication of human contributions to climate change will require the climatic influences of land-surface conditions and other processes to be factored into climate-change-mitigation strategies. Many of these processes will have strong regional effects that are not represented in a globally averaged metric. The currency of global and regional human-caused changes in terms of a regional climate change potential could offer a new metric useful for developing a more inclusive protocol. This concept would also implicitly provide a way to monitor potential local-scale environmental changes that could influence biodiversity.

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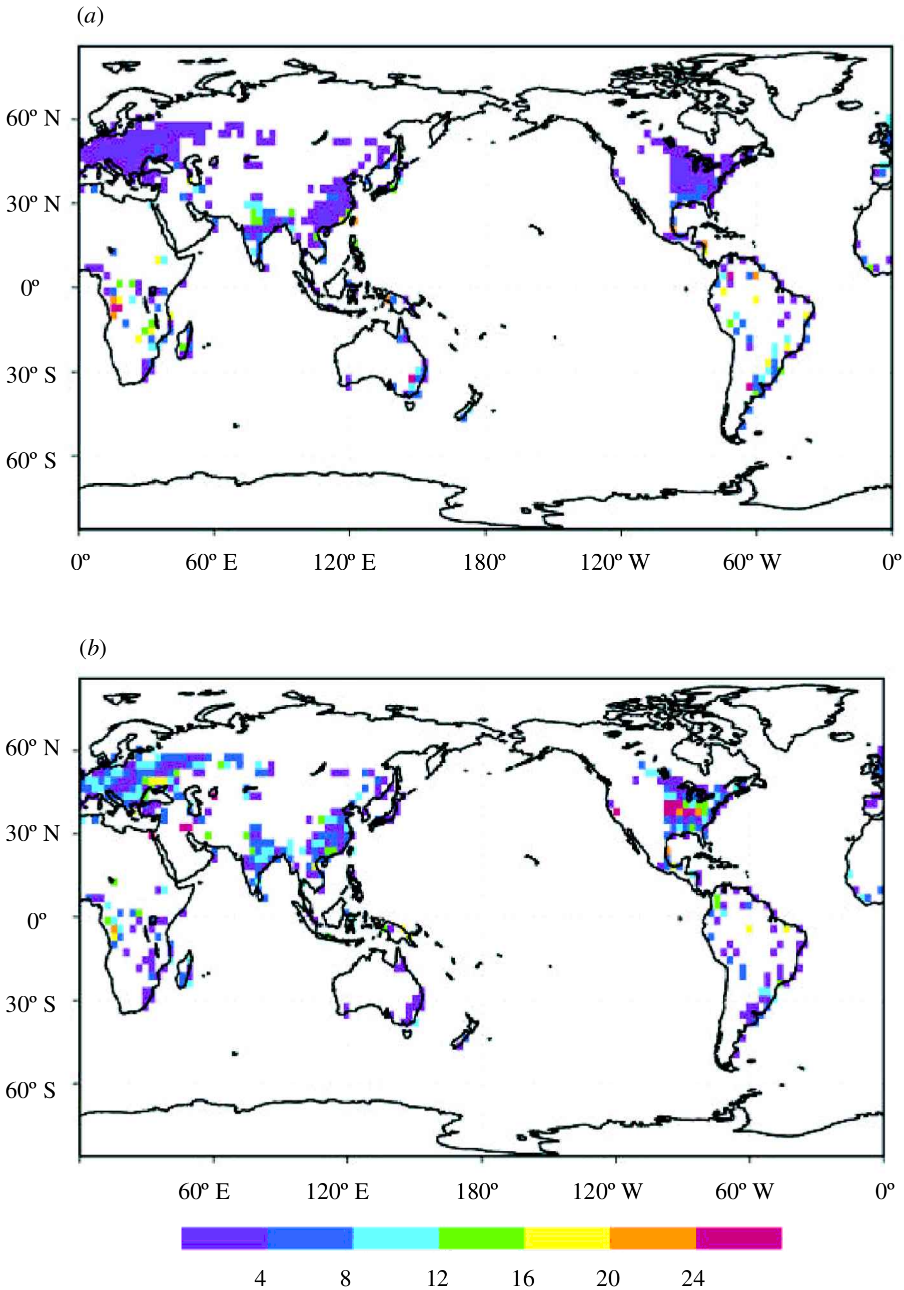

Figure 5. The 10-year average absolute-value change in surface latent turbulent heat flux in W m^{-2} at the locations where land-use change occurred: (*a*) January. (*b*) July. (Adapted from Chase *et al*. (2000).)

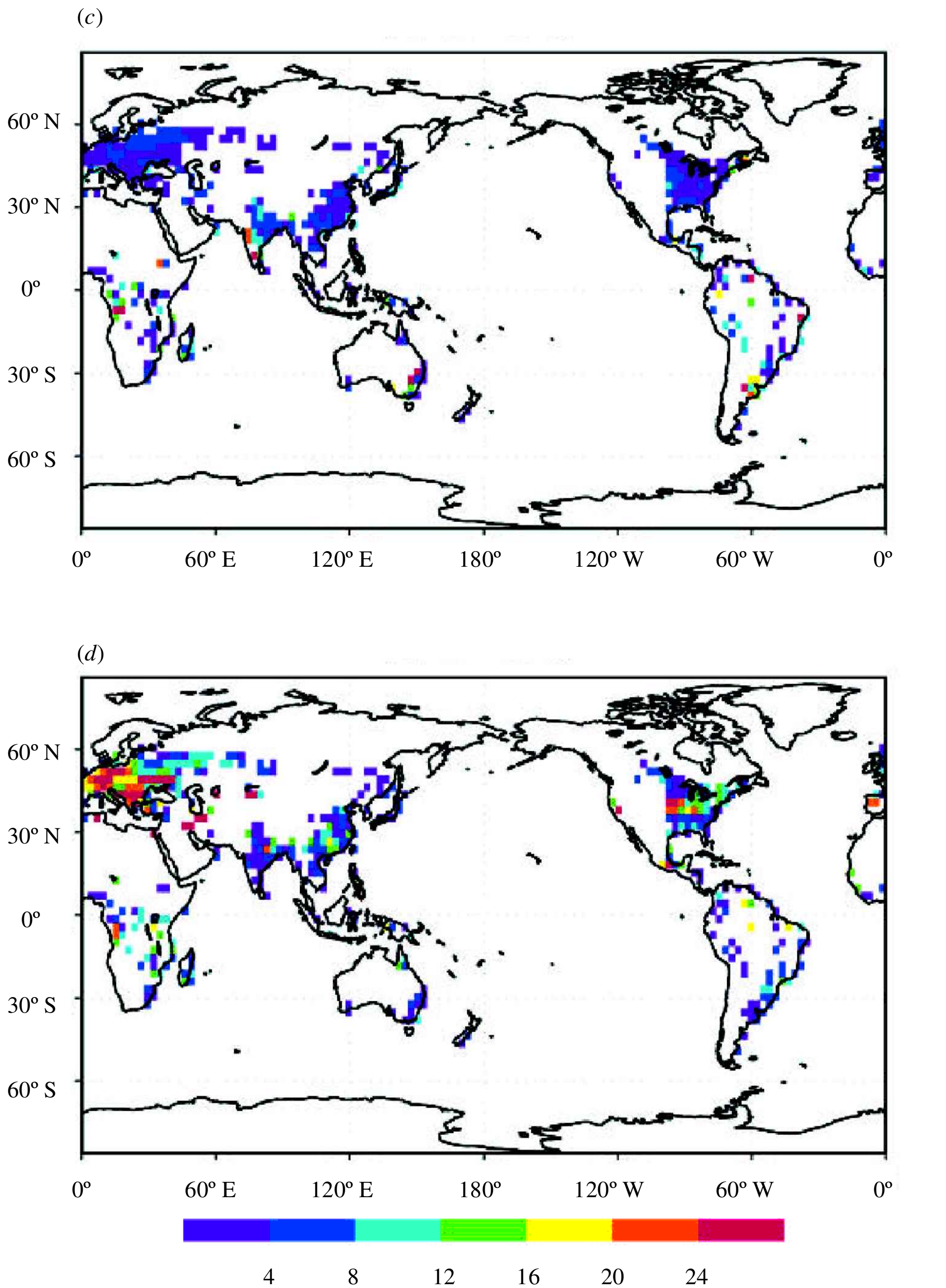

Figure 5. (*Cont.*) The 10-year average absolute-value change in surface sensible heat flux in W m^{-2} at the locations where land-use change occurred: (*c*) January; (*d*) July. (Adapted from Chase *et al.* (2000).)

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