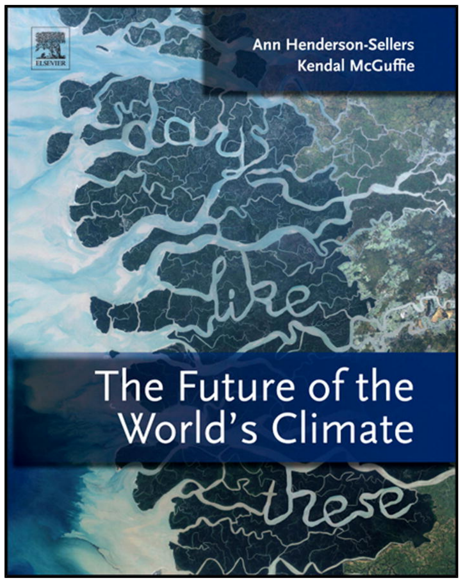

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Chapter 4

Human Effects on Climate Through Land-Use-Induced Land-Cover Change

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Chapter Outline

4.1. INTRODUCTION: LAND CHANGE AND CLIMATE

The modification of the landscape by humans and their ancestors is very ancient. The history of deforestation from prehistoric times through to the present was explored by Williams (2003). In a remarkable account, he highlighted the fundamental connectivity between human development and large-scale land modification. Large-scale landscape modification affects climate. There is a direct effect on atmospheric carbon dioxide (CO_2) concentrations, on the energy availability at the surface, on the partitioning of energy and water at the surface, and on the emissions of aerosols and various volatile organic components. However, while it is indisputable that large-scale landscape modification affects climate (according to theory and local-scale observations), there is a necessity to identify the footprints of land-cover changes on climate, to define 'large-scale', what is meant by 'modification', and what scale the 'effect on climate' is and how it is realized. In terms of scale, 'large-scale' is used here to mean of the order of a hundred kilometres – the scale of a climate model grid element, a large catchment, or a major city (cf. Cleugh and Grimmond, 2012, this volume). In terms of climate modification, 'climate' means a longer-term statistic (e.g., decades), but does not necessarily imply averages. 'Modification' means a measurable impact on climate statistics that can be formally attributed to the land-cover change and is detectable above the background noise induced by natural variability.

There are two ways that large-scale landscape modification can occur. Humans can clear forests with fire, a chain-staw, tractor or buldozer. They also grow crops or build cities. This is a direct modification that is very obvious to all of us. Emissions of CO_2 from the burning of fossil fuels can also indirectly affect the large-scale landscape via the CO_2 fertilization effect (Collatz et al., 1991; Field et al., 1995; Sellers et al., 1996) that might lead to a change in leaf size, vegetation height, and stomatal function, via competition species types and via evolution of the nature of the vegetation itself. This indirect effect, among others, is subtler than deforestation but globally integrated may be more important (Dickinson, 2012, this volume).

This chapter provides an overview of how humans affect climate through land surface modification. We focus only on the physical and biogeophysical impacts because Dickinson (2012, this volume) highlights the biogeochemical effects of land-cover change while Harvey (2012, this volume) reviews fast and slow feedbacks on climate. To simplify the text, the term LULCC (land-use-induced

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land-cover change) will be used throughout; also simply termed anthropogenic land-cover change. Firstly, the scale of human modification will be highlighted to place this perturbation in context. Secondly, the theory that explains why landscape modification should affect climate will be explored. Thirdly, evidence of how landscape modification affects climate will be presented with the implications for the future climate discussed.

4.2. THE SCALE OF HUMAN MODIFICATION

Humans have undertaken intensive LULCC at a scale commonly underestimated by scientists working in other fields of climate science. Several groups have undertaken reconstructions of the scale of LULCC − work that underpins current efforts to determine the impact of this change on climate (e.g., Defries et al., 1995; Ramankutty and Foley, 1999; Klein Goldewijk, 2001; Hurtt et al., 2006; Pongratz et al., 2008). Figure 4.1a shows that by 1500 large areas of Western Europe had been partially cleared for agriculture and for timber. LULCC intensified, particularly in Western Europe, through to 1800. Indeed, by 1750, 7.9−9.2 million km^2 (6%−7% of the global land surface; note that all these percentages are calculated over land excluding Greenland and Antarctica) were in cultivation (Forster et al., 2007) although only in Western Europe and perhaps parts of northern China had the intensity of LULCC led to more than ~60% agricultural cover for a given region (Figures 4.1b, 4.1c, 4.1d). By 1990, 45.7−51.3 million km^2 (35%−39% of global land surface) was in cultivation, forest cover had decreased by about 11 million km^2, and intensive LULCC had impacted parts of the USA, much of Western Europe, India, northern China, and elsewhere. Large areas of the Southern Hemisphere underwent LULCC throughout the nineteenth century. By 2000 (Figure 4.1f) the fingerprint of human activity through LULCC has only *not* affected a few desert regions, the central Amazon and Congo Basins, and the Arctic and Antarctic (not shown). Figure 4.1 needs to be interpreted qualitatively. A careful examination points to some changes, such as over Australia between 1800 and 1900, which seem unlikely, but the overall pattern of changes are probably reliable at large scales. Williams (2003) provides a detailed account of these global and regional changes as well as their underlying causation.

The easiest and so far most traditional way to quantify the impact of these changes is by means of the surface albedo (the fraction of incoming solar radiation reflected by a surface), which can be translated into a change in radiative forcing. Forests are dark to the wavelengths of visible light and absorb very efficiently (albedos range from around 0.05−0.2). Croplands are commonly much more reflective (around 0.2−0.25 when they are snow-free, but much higher in the presence of snow: Bonan (2002)). Thus, the global net seasonal and annual impact of LULCC is an increase in albedo. While there is considerable uncertainty in the precise impact of LULCC on radiative forcing, Forster et al. (2007) suggest a reduction of 0.2 W m^{-2} ± 0.2 W m^{-2} on the global average. While this appears to be a large range (perhaps as large as 0.4 W m^{-2} or as small as 0.0 W m^{-2}), available evidence points to it being very small compared to well-mixed greenhouse gases (GHGs) (about +2.5 W m^{-2}).

Reasoning of this kind has led to the role of LULCC being ignored in most climate projections. The reports by the Intergovernmental Panel on Climate Change (IPCC, e.g., the Fourth Assessment Report, Solomon et al. (2007)) highlight LULCC as an area of uncertainty but the model simulations assessed in the Fourth Assessment Report do not include LULCC in terms of how it modifies the surface (only direct emissions of CO_2 are accounted for). Pitman et al. (2009) discuss the implications of omitting LULCC from the IPCC assessments and conclude that at the global-scale there is no evidence that this matters to global metrics, although a more recent study by Davin and de Noblet-Ducoudré (2010) demonstrated that oceanic feedbacks have the potential to enhance the LULCC impacts. Overall, IPCC AR4 statements of the likely amount of global warming associated with a given emission scenario are not flawed globally due to the lack of inclusion of LULCC. However, while GHGs are globally well-mixed, LULCC is highly concentrated at present in Western Europe, the eastern US, China, and India (Figure 4.1f). Thus, while the global impact may be negligible, and the impact on global means may be small, the global impact is contributed from a fraction of the land surface strongly coincident with human population (cf. Cleugh and Grimmond, 2012, this volume). Furthermore, it has been pointed out (Pielke et al., 2002; Davin et al., 2007) that radiative forcing is not a good way to measure the impact of LULCC on surface climate. A regionally-significant impact on climate can be achieved by LULCC without any change in albedo if the partitioning of energy or rainfall is modified. LULCC changes the seasonality of heat and moisture fluxes, changes the probability of extremes, and may provide a local perturbation that triggers a change in the atmosphere sufficient to cause changes remote from the perturbation. Thus, while the global impact of LULCC on mean radiative forcing and mean global climate sensitivity may well be negligible, this finding is really only relevant in theoretical studies since humans and the Earth's terrestrial flora and fauna live, source their food, and source their water at local and regional-scales.

The question is therefore not whether LULCC affects the global climate (it does via release of, for example, CO_2; see Dickinson, 2012, this volume). Rather, it is whether it affects climate at any space or timescale of significance for living bodies; and clearly 'significant' is a value judgement. A major change in climate over one region may be

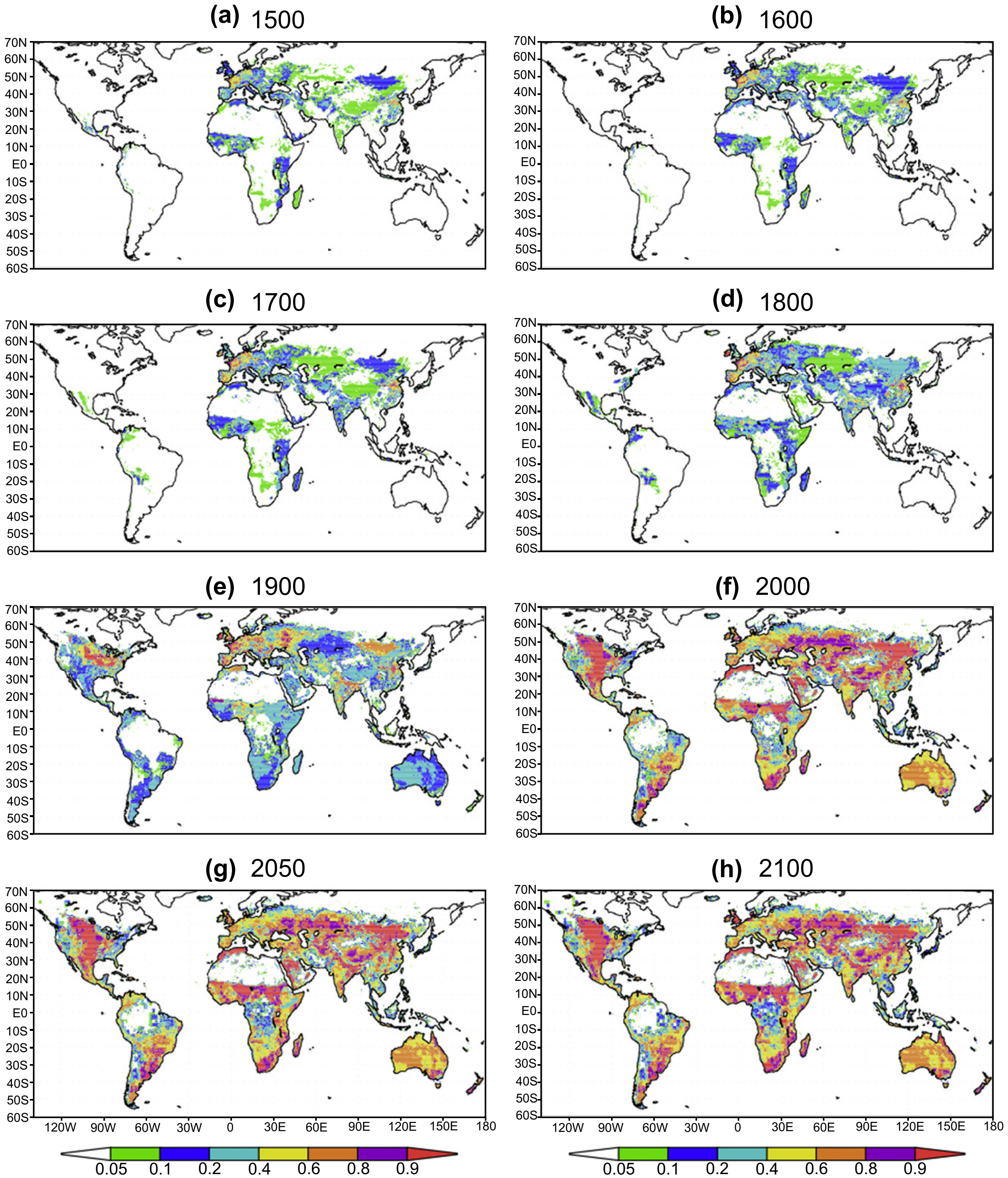

FIGURE 4.1 Reconstructed and projected LULCC for various time periods. The scale is the relative fraction of any grid box containing the sum of pasture or crops. *(Source: These data were downloaded from the Land Use Harmonization web site at* http://luh.unh.edu*).*

insignificant to the global climate scientist, but highly significant to the affected population.

4.3. MECHANISMS/PROCESSES THROUGH WHICH LULCC AFFECTS CLIMATE

4.3.1. The Terrestrial Carbon Balance

The Earth's land surface stores very large amounts of carbon and, where LULCC is represented by deforestation, large amounts of CO_2 are released into the atmosphere. Houghton (2003) estimated that LULCC has contributed net emissions of 156 GtC during the period 1850 to 2000. This is approximately 12–35 ppmv of the total CO_2 rise between 1850 and 2000 of ~90 ppmv (Matthews et al., 2004). The latest IPCC assessment (Denman et al., 2007) suggests that the net land-use change contribution to total CO_2 emissions is about 1.6 GtC per year comprising large tropical sources offset slightly by non-tropical sinks. Emissions of this scale are globally significant, add to the radiative forcing induced via the burning of fossil fuels, and the sum greatly exceeds the global significance of LULCC on radiative forcing via surface changes in albedo. However, to avoid overstating the importance of LULCC, while CO_2 emissions from this source exceeded those from fossil fuels up to 1900, fossil fuel emissions have risen rapidly since ~1950 and now comprise ~85% of total annual emissions (Raupach and Canadell, 2010), while emissions from LULCC have increased slowly over the twentieth century.

Based solely on radiative forcing concerns, the IPCC's non-explicit representation of LULCC seems reasonable. However, LULCC also affects climate via other physical and biogeophysical mechanisms, but the specific mechanisms depend on the nature of the perturbation, the climate of the region undergoing LULCC, and how the region's land cover may recover after being abandoned. For LULCC to affect climate requires a change to one of the fluxes that links the land to the atmosphere. This can occur due to a change in the albedo and thereby the net radiation, via mechanisms that affect how net radiation is partitioned between sensible and latent heat fluxes, or how rainfall is partitioned into evaporation, runoff, drainage, and soil moisture, and thereby in latent heat flux. To formalize these relationships and potential impacts of LULCC the surface energy balance and the surface water balance equations provide an ideal framework. The following provides common ways to represent these processes, particularly in climate models, but should not be interpreted as implying these are the only methods available.

4.3.2. The Surface Energy Balance

The shortwave radiation emitted by the Sun is reflected, absorbed, or transmitted by the atmosphere. An amount of energy ($S\downarrow$) reaches the Earth's surface and some is reflected (depending on the albedo, α). Of 100 units of energy entering the global climate system, 46–47 are absorbed by the surface (Rosen, 1999; Trenberth et al., 2009). Infrared radiation is also received ($L\downarrow$) and emitted ($L\uparrow$) by the Earth's surface (depending on the temperature and emissivity of the land and atmosphere). The net balance of the incoming and reflected shortwave radiation, and the incoming and emitted longwave radiation at the Earth's surface is the net radiation (R_n):

$$R_n = S\downarrow(1-\alpha) + L\downarrow - L\uparrow \tag{4.1}$$

Of the 100 units of solar energy entering the global climate system, at the top of the atmosphere over land, about 30–31 (according to the latest estimates by Trenberth et al., 2009) are exchanged as sensible and latent heat fluxes – the turbulent energy fluxes. The land surface significantly influences how these 30–31 units of energy are further partitioned between sensible (H: thermal convection) and latent heat (λE: moist convection) fluxes. A perfectly dry desert typically releases all 31 units of energy as H, while a water surface releases most of those 31 units as λE. The land surface also stores energy on diurnal, seasonal, and longer timescales (thousands of years in the case of heat stored in permafrost for example). R_n must be balanced by H, λE, the soil heat flux (G), and the chemical energy stored during photosynthesis and released by respiration (F):

$$R_n = H + \lambda E + G + F \tag{4.2}$$

LULCC directly changes the albedo, α, that perturbs the amount of energy available to drive the surface energy balance. From Equation (4.1), it is clear that a change in α must affect R_n and, therefore, affect all the terms in Equation (4.2). It is this perturbation that is highlighted in Forster et al. (2007) as a small reduction in radiative forcing because, on average, past LULCC has increased α.

In terms of the atmosphere, it is important to partition R_n between H and λE as well as possible. λE cools the surface very efficiently, moving 2.5×10^6 J for each kilogram of water evaporated away from the surface. Thus, a reduction in λE leads to a warmer surface, but also means less water vapour is contributed to the atmosphere, which, in turn, tends to decrease cloudiness and precipitation (Seneviratne et al., 2010). Decreases in H tend to cool the planetary boundary layer and reduce convection (Betts et al., 1996) and also reduce boundary layer depth. Complex feedbacks exist whereby changes in clouds or precipitation feed back to modify the initial perturbation to albedo. Precisely how LULCC affects the partitioning of R_n is, therefore, critical to understanding the role of LULCC on the dynamics of the atmosphere and on the coupled atmosphere–ocean system.

The two key equations that describe H and λE can be written in many ways. The link between surface characteristics and the turbulence that drives the exchange of H and λE and CO_2 is the aerodynamic resistance for heat (r_{ah}) and water (r_{aw}). Following Bonan (2002):

$$r_{ah} = \frac{1}{k^2 u_a}\left[\ln\left(\frac{z-d}{z_{0m}}\right) - \varphi_{m(\zeta)}\right]\left[\ln\left(\frac{z-d}{z_{0h}}\right) - \varphi_{h(\zeta)}\right] \tag{4.3}$$

$$r_{aw} = \frac{1}{k^2 u_a}\left[\ln\left(\frac{z-d}{z_{0m}}\right) - \varphi_{m(\zeta)}\right]\left[\ln\left(\frac{z-d}{z_{0w}}\right) - \varphi_{w(\zeta)}\right] \tag{4.4}$$

Quantities such as k (von Kármán constant) are independent of the nature of the surface. Other quantities include u_a (wind speed at height z) and the roughness length for momentum (z_{0m}). The functions $\varphi_m(\zeta)$, $\varphi_h(\zeta)$, and $\varphi_w(\zeta)$ account for the influence of atmospheric stability on the turbulent fluxes. This leaves three terms: z (height in the surface layer), d (the displacement height), and z_{0h}, z_{0w} (the roughness lengths for heat and moisture). The variables $z_{0h} + d$ and $z_{0w} + d$ are the effective heights at which heat and moisture are exchanged with the atmosphere. The key fact linking Equations (4.3) and (4.4) to LULCC is that changing a landscape affects $z_{0h} + d$ and $z_{0w} + d$. A typical value of $z_{0h} + d$ and $z_{0w} + d$ for a grassland or crop is a few tens of centimetres depending on the type of grass or crop, although this is highly dependent on the type of crop. For a forest, $z_{0h} + d$ and $z_{0w} + d$ is around 8–10 metres. Thus, changing the nature of the vegetation changes (commonly by two orders of magnitude) $z_{0h} + d$ and $z_{0w} + d$, which, through Equations (4.3) and (4.4), directly affect r_{ah} and r_{aw}, which, in turn, affect H and λE. The H can be represented as a quasi-diffusive process:

$$H = \frac{T_s - T_r}{r_{ah}} \rho c_p \tag{4.5}$$

where T_s is the surface temperature (itself a function of the surface energy balance), T_r is a reference temperature above the surface, ρ is the air density, and c_p is the specific heat of air. Thus, a change in r_{ah} directly affects H — something that must also affect λE through Equation (4.2). However, λE fluxes are underpinned by a more complex process than H as it involves most of the challenges of simulating H, plus all those processes that provide the water to evaporate (canopy interception, root water uptake, soil moisture diffusion, etc.). While there are several ways to represent λE, the aerodynamic approach is commonly used (Sellers, 1992):

$$\lambda E = \left(\frac{e_s - e_r}{r_{surf} + r_{aw}}\right)\frac{\rho c_p}{\gamma} \tag{4.6}$$

where e_s is the saturated vapour pressure at the surface from which evaporation is occurring, e_r is the vapour pressure at a reference height, and γ is the psychrometric constant. Thus, as with H, the change in $z_{0e} + d$ and thereby the change in r_{aw} must directly impact λE. However, Equation (4.6) contains an additional term, r_{surf}, the surface resistance to the transfer of water from the surface to the air, which is highly variable depending on the land-cover type. In general, from terrestrial surfaces, λE can be thought of as the sum of three fluxes: a flux from the ground (λE_g), a flux from the leaves (λE_v), and a flux from the air within the canopy to the air above the canopy (λE_a):

$$\lambda E_g = \left(\frac{e_g - e_{ac}}{r_{ac}}\right)\frac{\rho c_p}{\gamma} \tag{4.7}$$

$$\lambda E_v = \left(\frac{e_v - e_{ac}}{r_c}\right)\frac{\rho c_p}{\gamma} \tag{4.8}$$

$$\lambda E_a = \left(\frac{e_{ac} - e_a}{r_a}\right)\frac{\rho c_p}{\gamma} \tag{4.9}$$

where the subscripts for λE and e refers to the ground (g), canopy surface (v), canopy airspace (ac), and the air above the canopy (a). Different vegetation types have different levels of ground cover. A dense forest may cover nearly 100% of the surface, making λE_g negligible. Removing this forest completely makes λE_v irrelevant and λE_g dominant. These fluxes differ because the resistance terms used in the formulations of Equations (4.7), (4.8), and (4.9) represent different pathways for the water molecules. Both r_{ac} and r_a are physically based resistances: r_c, the canopy resistance, includes biological processes that are active in managing losses of water via the stomates to balance gains of CO_2. Since each type of vegetation has a specific optimization to balance water loss with carbon gain, each type of vegetation has evolved strategies that lead to different behaviour of r_c. Most simply (Dickinson et al. 1991):

$$r_c = \frac{< r_s >}{L} \tag{4.10}$$

where the angled brackets denote an inverse average over the range of the canopy leaf area index, L such that:

$$< () > = \frac{L}{\int^L dL/()} \tag{4.11}$$

and the stomatal resistance, r_s, is:

$$r_s = r_{s\,\min} f_1(T) f_2(D) f_3(PAR) f_4(\theta) f_5(\Psi)\ldots \tag{4.12}$$

where each of the functions f range from zero to one and represents a dependence of r_s on temperature (T), vapour pressure deficit (D), photosynthetically active radiation (PAR), soil moisture availability (θ), and nutrient stress (Ψ). This approach, based on Jarvis (1976), is commonly used in land-surface schemes coupled to climate models, but it is limiting because it does not couple the carbon, water, and energy balances at the level of the leaf. More recently (e.g., Collatz et al., 1991; Bonan, 1995; Arora, 2002), models have included parameterizations based on Ball et al. (1987) that link the response of stomatal conductance (g_s) to the rate of net CO_2 uptake (A_n), the relative humidity (h_s), and the CO_2 partial pressure at the leaf surface:

$$\frac{1}{r_s} = g_s = m\frac{A_n h_s P}{c_s} + b \tag{4.13}$$

where m is a constant and b is a coefficient obtained from observations ($b = 0.01$ and 0.04 for C3 and C4 plants respectively), P is atmospheric pressure, and c_s is the CO_2 partial pressure at the leaf surface. A_n, in turn, is defined as the lesser of w_c (the Rubisco-limited rate of photosynthesis and w_j (the light-limited rate allowed by the rate of regeneration of the Ribulose-1,5-Bisphosphate (RuBP) molecule):

$$A_n = \min(w_c, w_j) - R_d \tag{4.14}$$

and R_d is the dark respiration rate. Both w_c and w_j vary as a function of the vegetation type. Therefore, LULCC affects A_n, which, in turn, affects g_s and r_s, and therefore r_c and therefore λE and therefore H. Furthermore, LULCC affects L, which affects the scaling of r_s to r_c (Equation (4.10)) and λE and therefore H.

There are two remaining major vegetation-related parameters that affect the exchange of energy and water between the land and the atmosphere. The leaf area index is important in intercepting rainfall and the subsequent redistribution of intercepted rainfall to throughfall, stemflow, or re-evaporation. The amount of rainfall interception is based on attributes including canopy characteristics, rainfall intensity and duration (Dickinson et al., 1986), and wind. The importance of rainfall interception is linked to the timescales over which rainfall is returned to the atmosphere. Intercepted water that fails to fall through to the soil re-evaporates rapidly, commonly on timescales of minutes to hours, due to the large aerodynamic roughness of canopies, high ventilation, and large surface area in contact with the atmosphere. In contrast, precipitation that infiltrates into the soil tends to remain there for much longer periods (days to months) and may reach rivers or groundwater to be effectively lost to the atmosphere, potentially for years to centuries. A key rationale for incorporating canopy interception into land-surface models (LSMs) is, therefore, to capture the higher time frequency response driven by canopy interception (Scott et al., 1995), which could trigger changes in the diurnal cycle of clouds and affect radiation, in comparison to the slower timescale response of soil-based processes.

Precipitation interception (I) by canopies varies depending on the vegetation type. Rutter et al. (1972) suggest it is commonly 20%–40% of annual rainfall in conifers and 10%–20% in hardwoods. Many models of canopy interception have been proposed, but many parallel Sellers et al. (1986) where interception of rainfall (P) is assumed to be similar to the treatment of the transmission of the solar beam for spherically distributed leaves, which takes into account the factor of leaf angle distribution:

$$I = P(0.25(1 - e^{L(p_1+p_2)})) \tag{4.15}$$

where p_1 and p_2 are vegetation-type dependent parameters that depend on leaf angle parameters (Sellers et al., 1986). Note the use of L in the exponent; thus, I is dependent on L, which is dependent on vegetation type. Typically, LULCC, in the form of deforestation, reduces L but the indirect impact of human activity via CO_2 fertilization has the potential to increase L.

The final piece of the jigsaw is the response of vegetation to the risk of moisture stress. Some plants respond to moisture stress by dieback. Grasses in particular are shallow-rooted and tend to simply die under severe moisture stress and regrow from seed when moisture is again available. Trees tend to allocate more carbon to grow roots that penetrate more deeply into the soil, potentially to tap groundwater (Jackson et al., 1996; Canadell et al., 1996). As a result, trees can withstand dry periods and maintain transpiration and carbon uptake and thereby maintain a cooler and moister local climate (Bonan, 2002). Following LULCC in the form of deforestation, therefore, an environment of relatively low seasonality in the moisture flux (because the trees can tap deep soil water) is replaced by crops or grasses that may evaporate actively in spring and early summer, but as the soil dries the large evaporative flux is replaced with a large sensible flux with implications for the overlying atmosphere. Kleidon and Heimann (2000) used a relatively coarse resolution climate model to explore the role of deep roots in tropical forests. They suggest that when deep roots are correctly represented, deforestation can have a significant impact on the simulation of the atmosphere and these can lead to teleconnections via changes in moisture transport (cf. McGuffie and Henderson-Sellers, 2004).

Thus, LULCC affects *both* the energy available to drive the system via a change in the albedo and changes in the way net radiation is partitioned between sensible and latent heat fluxes because it affects the capacity of the soil–vegetation system to evaporate. Changes in LAI, roughness, stomatal conductance, photosynthesis, and root–soil water interactions all combine to affect the balance of H and λE. *This is not merely on the annual or decadal timescale.* LULCC also directly impacts the seasonality of these fluxes. Deforestation tends to lead to a drier surface that is warmer, with a high seasonality in λE and a higher probability of severe moisture stress late in the season. Drier surfaces tend to have a larger diurnal temperature cycle with higher maximum temperatures while, if sustained for longer periods, can lead to a higher probability of heatwaves that are suppressed in a moister and cooler forested environment.

Figure 4.2 illustrates the impact of LULCC in the tropics using observational data measured over two sites in Amazonia: a near-undisturbed forest (the Tapajos National Forest: Miller et al. (2009)) and a cleared forest now covered with pasture (Para, Brazil: Fitzjarrald and Sakai (2010)). These sites are ~80 km apart and therefore not precisely comparable, but are exposed to similar large-scale meteorology. Figure 4.2a shows the impact of LULCC on temperature measured at 2 m above the surface for the average of three days in January. The pasture site is warmer throughout the day, by ~1°C at night and by more than 2°C during the day. The two sites are contrasting in terms of the surface energy balance. The pasture site experiences relatively low latent heat (Figure 4.2b) and high sensible heat fluxes (Figure 4.2c) while the forested site experiences high latent heat and low sensible heat. This is not an isolated occurrence; Figure 4.3a shows that the forest site is cooler throughout the year (2002 is shown here) by 1.5°C–2°C. The capacity of the forests to maintain a high evaporative flux through the year, and through any period of lower rainfall, is shown in Figure 4.3b. The pasture experiences high evaporation early in the year (as high or higher than the forest) but, from July, while the forest maintains a flux of ~100 W m^{-2}, the evaporation from grasses drops to ~20 W m^{-2} and the sensible heat increases, highlighting a moisture-limited surface (Figure 4.3c). Sensible heat remains low through the year in the forest (Figure 4.3c).

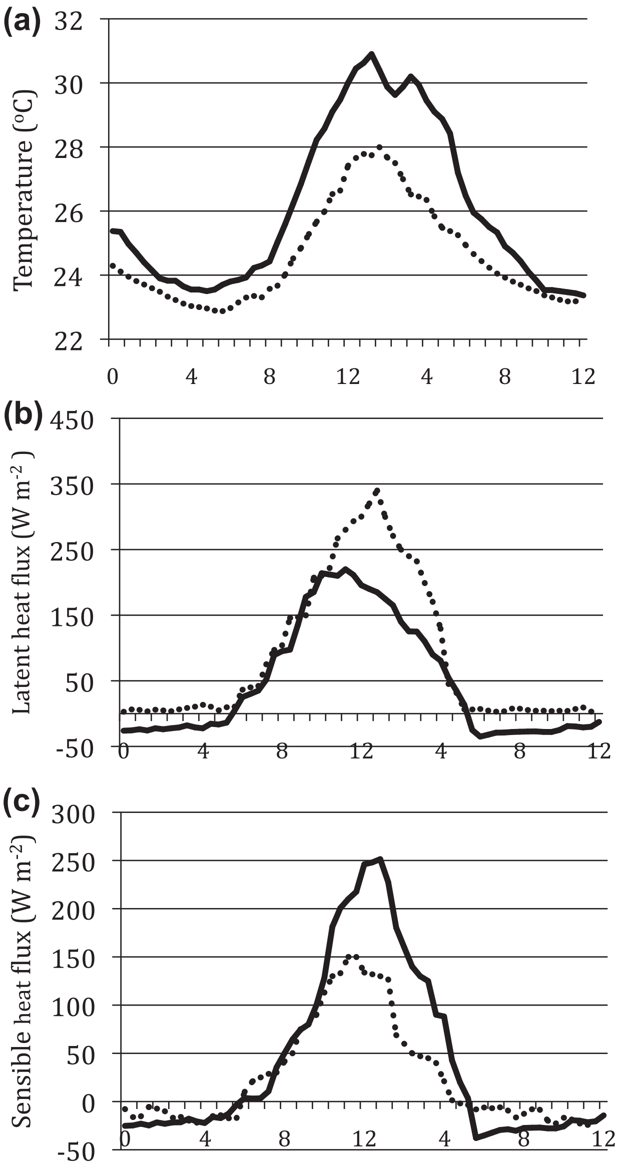

FIGURE 4.2 Diurnal cycles of temperature, latent heat flux, and sensible heat flux for a near-undisturbed forest (the Tapajos National Forest: Miller et al. (2009) – dotted line) and a cleared forest now covered with pasture (Para, Brazil: Fitzjarrald and Sakai (2010) – solid line). Data are averaged over three January days.

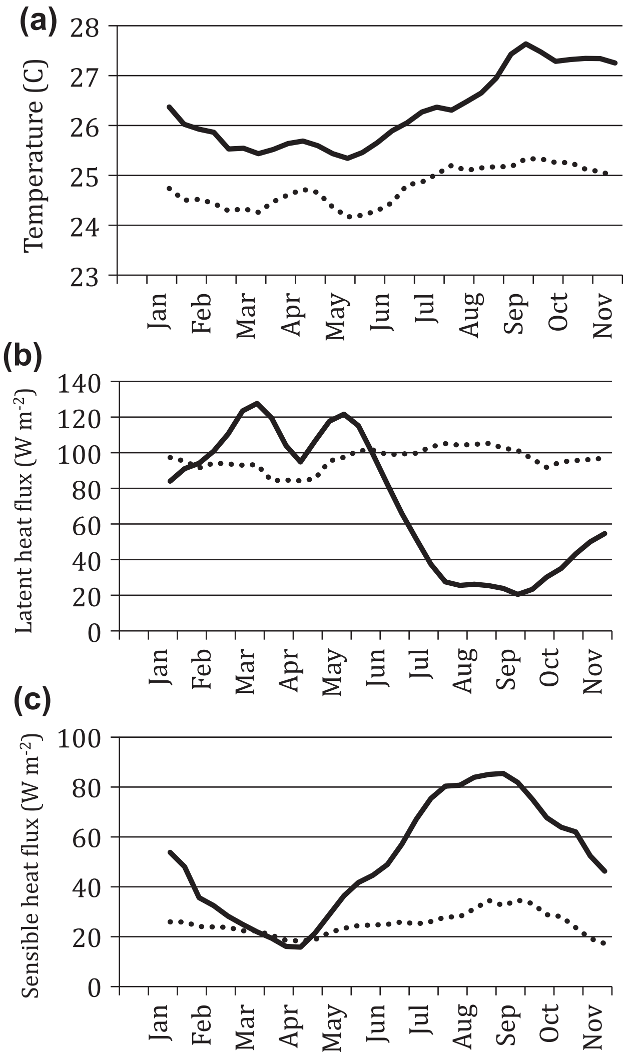

FIGURE 4.3 As Figure 4.2, but averaged over a full annual cycle. Lines are smoothed to highlight the differences between the near-undisturbed forest (dotted line) and the cleared forest (solid line).

The scale of this difference, particularly in the second half of the year, is considerable and is sustained to an extent that could likely affect cloud cover, boundary layer depth, and atmospheric moisture content, if spatially significant.

Figures 4.2 and 4.3 represent one measured example of the impact of LULCC over a relatively short period of time. It is important not to extrapolate this impact over the scale of the LULCC (Figure 4.1), since the way fluxes respond to deforestation (for example) depend very much on the background climate (Davin and de Noblet-Ducoudré, 2010) and the characteristics of the native vegetation being removed. Nor should one assume that Figures 4.2 and 4.3 are representative of all days or seasons in the tropics. It is also important not to prejudge the case that these changes would perturb the atmosphere in a climatological sense or assume that, if clouds and atmospheric moisture are affected, this would necessarily have a wider scale impact.

4.3.3. The Surface Water Balance

Rainfall (P) falling on the surface is partitioned between evaporation (E) and runoff (R) with changes in soil moisture (S) and snow mass (N):

$$P = E + R + \Delta S + \Delta N \tag{4.16}$$

However, this can be expanded to account for interception loss I, the difference between evaporation pathways (through the soil, E_s, or the vegetation E_t) and the partitioning of total runoff between fast (R_{surf}) and slow (R_{drain}) components:

$$P - I = E_s + E_t + R_{surf} + R_{drain} + \Delta S + \Delta N \tag{4.17}$$

Land-cover change affects I, the partitioning of E between E_s and E_t via resistances to the flow of water, canopy cover, and the distribution of roots. Land-cover change affects the intensity of rainfall hitting the surface because (in the case of deforestation) of reduced I and likely increased R_{surf}. If more rainfall generates fast runoff then this water tends to flow into rivers and is rapidly removed from a catchment in comparison with water that infiltrates and increases soil moisture. Changes in the rate of infiltration and changes in the production of surface runoff and drainage have been shown to have clear impacts on the amount of soil water available for plants, and therefore on the values of all evaporative fluxes. These potentially affect P via atmospheric feedbacks (changes in atmospheric moisture and convergence), as illustrated by Ducharne et al. (1998) through sensitivity studies changing the formulation of infiltration/runoff/drainage processes. Moreover, different land-cover types also have distinctly different root distributions (Canadell et al., 1996; Jackson et al., 1996) that can lead to very different strategies to extract water from the soil, and therefore to very different amounts of water transferred back to the atmosphere through evapotranspiration (Kleidon and Heimann 2000). These processes, though, have not yet received enough attention in most LSMs (Feddes et al., 2001) and such omission may very well lead to an underestimate of the impacts of LULCC on climate.

4.3.4. The Snow−Climate Feedback

Over 50% of Eurasia and North America can be seasonally covered by snow (Robinson et al., 1993), leading to significant spatial and temporal fluctuations in surface conditions. The properties of snow (e.g., high albedo, low roughness length, and low thermal conductivity) lead to global-scale impacts (Vernekar et al., 1995). Snow is also one of the key feedbacks within the climate system and plays a very strong role as a positive feedback. In the context of LULCC, snow falling on a forest tends to drop, over relatively short time periods, from the canopy on to the soil. The forest, therefore, remains radiatively dark, absorbing solar radiation to stay relatively warm. If the forest is removed and replaced with grasses or crops, then these tend to be senescent in winter and small amounts of snow can transform a region from being relatively dark due to the low reflectivity of most organically rich soils to being highly reflective once snow accumulates enough to mask the vegetation. The snow−albedo feedback, coupled with changes in vegetation, is therefore dramatically different depending on the nature of the vegetation. In a landmark paper, Betts (2000) explored this issue to show that forestation in regions subject to seasonal snow cover led to significant regional warming, marginally due to a lower albedo through summer and, most importantly, due to the reduction in the impact of snow in winter and spring caused by the masking effect of the trees. Conversely, deforestation that typically warms in the tropics can cool in the higher latitudes where snow becomes a stronger force through winter and spring, as discussed in Davin and de Noblet-Ducoudré (2010).

4.3.5. Summary

The coupling of the three major roles of the terrestrial surface − the surface energy, water, and carbon budgets − is intimate. A change in λE due to a change in net radiation, or the mechanisms that influence λE, affects the surface energy balance. Since λE (W m^{-2}) and E (mm, mm per unit time, or kg m^{-2}) are linked via the latent heat of vaporization, changes in λE affects the surface water balance and, therefore, the partitioning of P between E and R. Since λE can occur as transpiration, λE is coupled into the carbon balance (see Dickinson, 2012, this volume). Changes in transpiration can affect carbon uptake, but changes in atmospheric CO_2 can also reduce stomatal conductance, reduce transpiration, reduce total evaporation, and therefore have the potential to increase runoff via the surface water balance (Gedney et al., 2006; Betts et al., 2007).

4.4. LINKS BETWEEN LULCC AND CLIMATE

A change in the surface energy balance *must* affect the overlying atmosphere. A reduction in λE reduces the flow of moisture into the atmosphere, reducing the likelihood of rainfall and reducing cloud cover. An increase in H tends to warm the atmosphere, deepening the boundary layer and likely increasing advection of heat and moisture. A large-scale reduction in vegetation cover also reduces the aerodynamic drag on the atmosphere, increasing wind speeds and increasing the distances from, or to which, moisture and heat can be advected. However, while LULCC must affect the overlying atmosphere, the observational evidence that this is climatologically significant is not extensive. The absence of observational evidence does not mean there is no climatological influence (as reported below), but disentangling the LULCC impact from the influence of, for example, increased GHGs, changes in aerosols concentration, and natural variability remains very challenging.

One region that has been extensively studied is Western Australia. In a body of work covering almost two decades, Tom Lyons has established a regionally significant impact of LULCC on the micrometeorology of specific Australian areas (Lyons et al., 1993; 1996; Lyons, 2002), more specifically on the characteristics of the boundary layer and on convection. Changes in the regional meteorology have been linked to LULCC using a combination of observations and modelling, and the perturbation appears to have affected cloudiness resulting from changes in sensible heat fluxes (Ray et al., 2003), at least on short timescales. There is further evidence of LULCC affecting observed temperatures over the USA (Bonan, 2001) and at local-scales there is strong evidence in many regions. Irrigation, a form of land use that does not necessarily imply changes in land cover, has also been recognized as a significant way to impact the diurnal amplitude of temperature, its daily and even seasonal values, and evapotranspiration (de Ridder and Gallée, 1998; Boucher et al., 2004; Sacks et al., 2009). However, almost every other observationally-based example of LULCC affecting climate is a short-term case study or is focused on perhaps the most intensive form of LULCC — urbanisation (see Arnfield, 2003). Urbanisation does affect climate (Karl et al., 1988; Gallo et al., 1999; Hale et al., 2008; Cleugh and Grimmond, 2012, this volume).

In a recent review (Seneviratne et al., 2010), the potential role of soil moisture is separated into three classes. Soil moisture and LULCC are not directly comparable of course, but both can represent a major and coherent land-based anomaly, and therefore some insight sourced from analyses of soil moisture might be applicable to LULCC. Firstly, Seneviratne et al. (2010) note that higher rainfall usually leads to higher soil moisture. Secondly, they ask whether higher soil moisture leads to higher evapotranspiration. They point out that this relationship would only hold in regions defined as transitional — that is, where a combination of energy limitation and water limitation defines the evapotranspiration. Clearly, over a water surface, the availability of water does not limit evaporation; rather, the availability of energy to drive evaporation provides the constraint. Similarly, over a hot desert, energy is unlikely to be the limit to evaporation; rather, the lack of water represents the limitation. At some points in-between, 'transitional' situations arise that are more complex than these examples and both water and energy can provide limits to evapotranspiration. We can, therefore, ask a similar question to Seneviratne et al. (2010): does LULCC (represented by deforestation) lead to lower evapotranspiration? As for soil moisture, the answer depends on the detail of the controls on evapotranspiration. Depending on the specific environment, LULCC (deforestation) could decrease evapotranspiration, but it could also increase the release of latent heat (i.e., evapotranspiration) if albedo is decreased, for example, via the planting of a particularly dark crop (Ridgwell et al., 2009). LULCC could also increase evapotranspiration if selective logging increases the effective roughness by introducing small cleared areas within a forest (Mei and Wang, 2010). Thirdly, Seneviratne et al. (2010) ask whether higher evapotranspiration leads to higher rainfall. They note that this is not straightforward and, while a large number of modelling studies have identified a possible feedback between higher evapotranspiration and higher rainfall (Beljaars et al., 1996; Koster et al., 2000; Betts, 2004; Pal and Eltahir, 2008), others have pointed to the convective instability and/or cloud formation being stronger over dry surfaces (Ek and Mahrt, 1994; Findell and Eltahir, 2003; Hohenegger et al., 2009). The point here is that while LULCC (deforestation) tends to decrease evapotranspiration that might decrease rainfall, it is not necessarily always the case that it does. Indeed, regional warming, generally associated with decreased evapotranspiration, may lead to increased advection of air masses and potentially of water vapour, thereby favouring rainfall or changes in deep convection that might also trigger increased precipitation (Polcher, 1995).

An additional challenge relates to the 'hotspots' identified originally by Koster et al. (2004) (Figure 4.4). These are regions where soil moisture anomalies were shown to have a direct impact on the overlying atmosphere. Persistence patterns in soil moisture and evapotranspiration (Seneviratne et al., 2006; Teuling et al., 2006) or the correlation of evapotranspiration with temperature

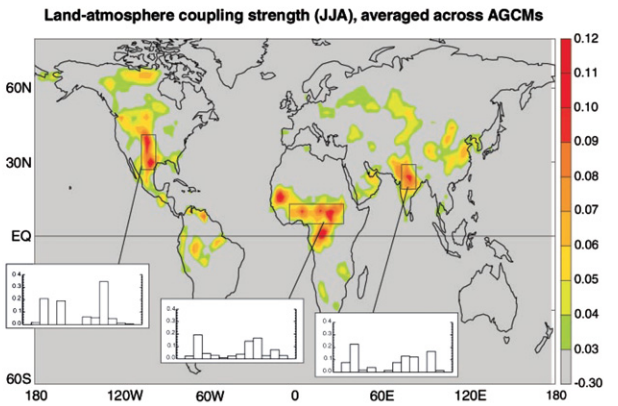

FIGURE 4.4 A diagnostic of the strength of the land-atmosphere coupling for boreal summer averaged across 12 atmosphere-only climate models. Regions coloured orange and red represent 'hot-spots' where the coupling is strong. *(Source: Koster et al., 2004. Reprinted with permission from AAAS.)*

(Seneviratne et al., 2006) can be used as indirect estimates of soil moisture–evapotranspiration coupling. Betts (2004) identified several tight coupling relationships between the soil moisture and boundary layer structure, cloud cover, and radiation using re-analysis in selected river basins. These relate to soil moisture, but similar 'hotspots' probably relate to LULCC. This implies that in some regions, LULCC at a given scale may not perturb climate because it happens to occur in a region of weak land–atmosphere coupling. Conversely, a smaller perturbation may have a significant impact on the atmosphere because it occurs in a tightly coupled region. LULCC might be expected to have the main impact where:

(a) land–atmosphere coupling is strong;
(b) rainfall recycling is strong; and
(c) most rainfall originates via local evaporation.

We neither know where these regions coincide, nor do we know if LULCC changes land–atmosphere coupling strength.

In the absence of strong observationally-based evidence, modelling provides useful insight to guide us on the impacts of LULCC on climate. There is a vast literature using regional and, to a lesser extent, global-scale models to explore the impact of LULCC. Some of the best examples have focused on Amazonian LULCC. They include Henderson-Sellers and Gornitz (1984), Dickinson and Henderson-Sellers (1988), Polcher and Laval (1994), Polcher (1995), McGuffie et al. (1995), Sud et al. (1996), Chase et al. (2000), and Avissar and Werth (2005). The impact of temperate deforestation on climate has been explored by Bonan (1997, 1999) and Oleson et al. (2004), while the impact of global LULCC has been considered by (among others) Chase et al. (2000), Betts (2000), Betts (2001), Govindasamy et al. (2001), Zhao et al. (2001), Bounoua et al. (2002), Zhao and Pitman (2002), Findell et al. (2007, 2009), Lawrence and Chase (2010), Pielke et al. (1998, 2002), Davin and de Noblet-Ducoudré (2010), and Pongratz et al. (2010). Mahmood et al. (2010) provide detailed reviews of much of this work, but focus mainly on the evidence in support of the impact of LULCC rather than any confounding evidence or weaknesses that apply to the reviewed studies. In general, we see a suite of warning signs that are worth watching for when reading the literature on LULCC. These do not mean the results identified by these warning signs are wrong, but it means they need to be considered with appropriate caution.

1. *Simulations conducted over a few days.* These commonly highlight a response by the atmosphere but impacts over a few days may be compensated a few days later to generate a minimal net impact. One certainly cannot extrapolate from a simulation of a few days to conclude a climatologically significant response.
2. *Simulations focused on a specific meteorological event* (e.g., thunderstorms; Shepherd, 2005). While LULCC might trigger an event (Gero and Pitman, 2006), one cannot infer a climatologically significant result from single events. Even well-conducted studies (e.g., Beljaars et al., 1996) may be examples of a special case too rare to be climatologically significant.
3. *Simulations that convert large regions from tropical forest to short grassland, or convert all global forests to grassland* (Davin and de Noblet-Ducoudré, 2010). These are important in examining the potential or theoretical impact of LULCC. Some cases are among the most famous climate model experiments conducted (Henderson-Sellers and Gornitz, 1984) but, as Figure 4.1 suggests, it is not a plausible outcome in the twenty-first century. The focus herein is on perturbations that are plausible in the same way as emission scenarios used in AR4 are plausible, while a climate simulation to 2100 using a CO_2 concentration of 3500 ppmv would be seen as a sensitivity study.
4. *With exceptions, simulations conducted with a single climate model.* Climate models vary greatly (Randall et al., 2007) and project very different responses to a perturbation. Koster et al. (2004) showed that the land is coupled to the atmosphere with very different strengths and this may lead to major differences between models in the impact of LULCC. In the same way as the IPCC would not base conclusions on one climate model, we should not base our conclusions of the significance of LULCC on any single model.
5. *Simulations that have assessed statistical significance using tests that do not take autocorrelation into account* (e.g., a Student's t-test). These tend to overestimate the significance of a change in the climate (see Findell et al. (2006) for an explanation) and generate excessive false positives ('false' in a statistical sense).

6. *Simulations using first-generation LSMs* (Pitman, 2003). These are likely to overestimate the impact of a perturbation (Chen et al., 1997) because they do not represent the coupled energy, water, and carbon cycles (cf. Dickinson, 2012, this volume).
7. *Simulations that have been updated using newer versions of a given climate model, where simulations with fixed sea surface temperatures (SST) were redone using mixed layer models, or where an experiment has been repeated more robustly than earlier examples.* Old experiments with older models are not necessarily wrong but, if newer and more physically realistic versions are available, it is likely wise to consider them.
8. *Simulations conducted at a coarse spatial resolution.* The relationship between model resolution and the scale of impact on the atmosphere is not clear, but early evidence (Hohenegger et al., 2009) points to a strong sensitivity between soil moisture anomalies and impact associated with the intensity of convection. While this does not necessarily imply a strong sensitivity between LULCC, convection, and model resolution, it does suggest that results from coarse resolution models should be treated cautiously. Defining the scale at which a model is 'coarse' in this context is not currently possible.

In the following we have tried to identify some selected publications that are less affected by these weaknesses. There are no individual studies that avoid all these weaknesses, so if these conditions were strictly enforced there would be virtually no results from models to discuss. However, there are some publications that have avoided key weaknesses that make them particularly useful to look at more closely. The following presents our judgement of where the science of LULCC and its impact on the atmosphere and climate currently stands – in the sense that these represent the most recent efforts to rigorously examine the impact of LULCC and provide guidance on how to resolve major methodological weaknesses.

4.4.1. Hasler et al. (2009)

In a well-constructed experiment, Hasler et al. (2009) undertook multimodel simulations to address the single-model weakness of earlier work by Avissar and Werth (2005) and Werth and Avissar (2002, 2005). They used three climate models, each at $4^\circ \times 5^\circ$ resolution and undertook 52-year simulations using fixed SSTs, omitting the first 4 years to avoid spin-up issues. The use of three models makes this an important paper that made a significant step forward in addressing results generated with a single climate model, which is why it is discussed here. The models used were either relatively old (two versions of the GISS climate model) or were used at a spatial resolution significantly degraded from the IPCC version of the model (CCSM was used at $1.4^\circ \times 1.4^\circ$ in the AR4: Randall et al. (2007)). It is not easy to judge how independent these three model estimates are, or how sensitive these three models are to a land perturbation. A version of the CCSM was used by Koster et al. (2003, 2006) that appears to be relatively sensitive to terrestrial processes (noting that this does not make it wrong). The imposed LULCC change was massive (but typical of earlier work by many authors, see figure 1 of their paper) including a conversion of regions larger than the Amazon Basin, the Congo, and all of South East Asia (including India, southern China, northern Australia, and Indonesia) from forest to grass and shrubs.

Hasler et al. (2009) assess the skill of each model in simulating seasonal rainfall: a very necessary step that is not always undertaken. While the rainfall climatology is not unreasonable, some large biases are clear in all three models. The authors do not show the actual changes in rainfall induced by LULCC as maps of mm d^{-1} in the three models and therefore it is not possible to determine whether the changes are very small or very large relative to the biases in the model. The authors also do not attempt to diagnose whether there is a link between the intensity of the rainfall response to LULCC and the magnitude of the present-day bias. Rather, they focus on the mean ensemble to track the strongest shared response between the models.

Hasler et al. (2009) show where rainfall has either significantly increased or decreased for at least 3 months of the year across an ensemble of the three models, as well as for individual models (figure 5 of their paper). They use a statistical test (Student's t-test) at a 95% significance level, *ignoring autocorrelation*. They used fixed SSTs and this was their rationale for ignoring autocorrelation. They find a strong response to LULCC of two kinds. Firstly, over regions of the tropics where LULCC has been perturbed, they capture a clear impact of LULCC (reduced rainfall rates) over Amazonia, the Congo, and parts of South East Asia. This result is consistent with a mass of literature and is probably robust at these large regional-scales. The result is common to all three models (see figure 5 of their paper) at a magnitude that is large relative to the background variability in each model. They also highlight a fair amount of positive responses remote from the actual perturbations. However, as at a 95% significance level one would expect 5% of all model grid points to give a false positive result (in a statistical sense), it is not convincing that the remote changes found by Hasler et al. (2009) exceed this level. This finding may, therefore, be modified if another statistical method were to be used. However, the authors also provide an analysis of geopotential changes. This builds on earlier analyses by Zhang et al. (1996) and Chase et al. (2000), who linked perturbations in the tropics to remote impacts using changes in atmospheric dynamics as the conduit for these teleconnections. They found seasonally-dependent changes in the northward transport and changes in diabatic heating and potential energy production associated with the changes in rainfall, although these mechanisms differed between the three models. They concluded that they found evidence of a wave-train forced by the changes in the tropics, but that this effect is buried in the natural variability of the climate system and can be more clearly found via multimodel ensembles.

Hasler et al. (2009) concluded that their results suggest that remote impacts of tropical deforestation on rainfall are real, but also note earlier conflicting results (e.g., Findell et al., 2006) and the fact that their results were inconsistent. They correctly note that the use of fixed SSTs tend to dampen internal model variability. While this would hint that using a coupled ocean model would increase the variability and, therefore, reduce the scale of LULCC impact, this is not necessarily the case if LULCC leads to a systematic change in the ocean temperatures (cf. Latif and Park, 2012, this volume).

Despite the fact that interpretation of this work still needs to be undertaken with care, Hasler et al. (2009) remains a rare publication — one that uses several models, long simulations, and an analysis of both local and remote impacts. It strongly reinforces the conclusion that LULCC has a strong regional impact, while noting that the scale of the perturbation imposed was not intended to be realistic. Despite the very sizable perturbation, the conclusion that their results "suggest the existence of a remote effect" is not definitive and may well be the result of their experimental and statistical design. Therefore, it needs to be addressed by many other models, as well as carried out with an additional accounting for oceanic feedbacks.

4.4.2. Findell et al. (2006, 2007, 2009)

In a series of publications, Findell et al. (2006, 2007, 2009) used the GFDL climate model at a $2^\circ \times 2.5^\circ$ resolution. A key development in their analysis was the accounting for autocorrelation and field significance in the statistical testing of the remote responses to LULCC. Simulation lengths were relatively long (around 50 years) and a mix of fixed SSTs and a mixed-layer ocean model was used. The GFDL climate model used in AR4 is among those with a higher sensitivity, but includes a coupled ocean model and is, therefore, only indicative of how the atmosphere-only model would respond. The GFDL model is also relatively strongly coupled to the surface (Koster et al., 2006) and might be expected to show a relatively large effect from LULCC.

Findell et al. (2006) perturbed the LULCC similarly to Hasler et al. (2009) with the broadleaf evergreen forests of South America, central Africa, and Oceania replaced with grasslands. The impact of these changes was a strong reduction in latent heat flux in the tropics, coincident with LULCC. This led to warming in these regions, a result in common with many earlier experiments. Findell et al. (2006) then explored the impact of LULCC on seasonal 200 hPa temperature and 200 hPa geopotential height, showing that there was a significant effect through the tropics in Northern Hemisphere spring and summer. In terms of precipitation, impacts were discernible in the tropics, but not in the extra-tropics. They conclude that their use of long-term averaging and sound statistical methods allow them to conclude that, while LULCC (limited to the tropics, but greatly over-intense in terms of the actual perturbation) has a significant impact on the tropics, it does not lead to significant extra-tropical impacts *in the GFDL climate model.*

Findell et al. (2007) extended the focus of Findell et al. (2006) from the tropics to a global implementation of LULCC using an observed estimate of historical LULCC. In these experiments, therefore, most of the changes in LULCC are in Europe, the eastern US, India, and China, and the intensity of LULCC is smaller than with full-scale clearance, as represented in many earlier studies. The impact on annual air temperature was statistically significant (accounting for autocorrelation and field significance) over India and Europe only and the change in annual precipitation was only significant over parts of Europe and India (Figure 4.5). Overall, the global impact of LULCC was, therefore, negligible, but regionally a significant impact on the climate was simulated. A common criticism of studies of LULCC is that, while there may be regional impacts, these are insignificant in comparison to major modes of variability in the ocean (e.g., El Niño—Southern Oscillation, North Atlantic Oscillation, Southern Annual Mode — see Latif and Park (2012, this volume)). Findell et al. (2009) built on Findell et al. (2007) to explore this issue using an identical model to compare the impacts of ocean modes of variability with the LULCC perturbation.

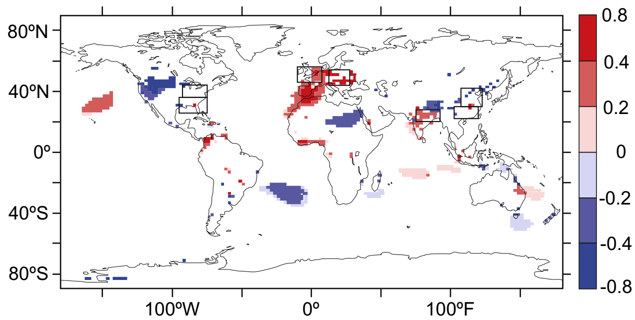

FIGURE 4.5 Annual differences in surface air temperature due to land-cover change. Differences are shaded only where they are statistically significant at the 95% significance level according to a modified t-test. A value of 6.0% passes the 95% significance level. *(Source: Findell et al., 2007. © American Meteorological Society. Reprinted with permission.)*

Their results demonstrated that in terms of global impacts, LULCC was negligible in comparison to the ocean modes tested in terms of air temperature but did have an impact on global precipitation on a par with the ocean modes. Globally, LULCC could be ignored in terms of the non-CO_2 component of climate change. However, over regions of LULCC, the impact of LULCC was notably larger on hydrometeorological quantities than the ocean modes.

In retrospect, this is hardly surprising since the forcing from LULCC is regionally focused, whereas the ocean forcing is by definition remote. However, Findell et al. (2009) demonstrated that, if a climate model is being used to explore surface quantities and the region has undergone significant LULCC, it is necessary to consider LULCC as a significant forcing. Since regions of intense LULCC are coincident with dense human populations and are therefore a key vulnerability to climate, the need to include LULCC in simulations of the regional impacts of (for example) increasing CO_2 is clear. It does not matter whether LULCC has a global teleconnection in this context.

4.4.3. Urban LULCC

Despite the increasing urbanisation in many countries, and the decades of research that point to large urban areas affecting regional climate (Arnfield, 2003), the representation of LULCC in the form of urban landscapes has received little attention in climate modelling or climate projections research. Oleson et al. (2008) are among a small group of authors who have coupled urban landscapes into climate models for innovative research into geo-engineering (Oleson et al., 2010) while several groups have explored the impact of urban landscapes on the scale of individual cities (see Cleugh and Grimmond, 2012, this volume).

McCarthy et al. (2010) used the Hadley Centre climate model and undertook 25-year simulations, including an urban land-surface scheme (Best et al., 2006) linked to the extent of urban coverage (Loveland et al., 2000). McCarthy et al. (2010) also added energy representing urban energy use into the atmosphere above the urban landscapes (either 20 W m^{-2} or 60 W m^{-2}). Their results show strongly regionally-varying impacts of adding in urban landscapes with the effect mainly seen on minimum temperatures and largely caused by the urban land, not the additional energy flux used to represent energy use. The impact on maximum temperatures was small relative to the impact of increasing CO_2 (perhaps an additional 15% of warming) but was significant in minimum temperatures. Expressed as a change in the number of hot nights for major cities, McCarthy et al. (2010) show dramatic increases for all cities included. Figure 4.6 shows a simplified version of McCarthy et al.'s (2010) results omitting estimates of uncertainty and the 60 W m^{-2} simulations. Figure 4.6 shows that the frequency of hot nights is always negligible under $1 \times CO_2$ compared to $2 \times CO_2$. Furthermore, representing urban landscapes increases the likelihood of very hot nights under $1 \times CO_2$ (e.g., from negligible using rural land cover to 10–20 events on average per year over Los Angeles and Tehran). More dramatically, doubling CO_2 greatly increases the frequency of very hot nights in all cities (Alexander and Tebaldi, 2012, this volume). However, representing the cities in the climate model adds an additional and large increase in this frequency, doubling it in cities including Delhi, Beijing, Los Angeles, and Tehran. That is, omitting the representation of urban landscapes in climate models leads to a systematic underestimate of the changes over major cities. Note, Figure 4.6 suggests that it is the addition of the urban landscape, not the 20 W m^{-2}, that dominates the results in agreement with many other similar evaluations (Cleugh and Grimmond, 2012, this volume).

McCarthy et al. (2010) were the first to explore this issue explicitly. It is important to note that single model simulations were utilized, with a single estimate of urban extent. The best way to parameterize urban landscapes in climate models is at best uncertain and is some two decades less well-developed in comparison to natural landscapes (although rapidly resolving some of these uncertainties: Grimmond et al. (2010)). The conclusion of McCarthy et al. (2010) that urban landscapes increase the vulnerability of populations to global warming seems robust, but the scale of this increase remains necessarily uncertain until multiple-model comparisons can be undertaken.

4.4.4. Land-Use and Climate, Identification of Robust Impacts (LUCID): Pitman et al. (2009)

Given the criticism that most LULCC experiments were conducted with a single climate model, it was natural that a project was set up to attempt to resolve this problem. Under the auspices of the International Geosphere Biosphere Programme's Integrated Land Ecosystem–Atmosphere Processes Study (iLEAPS) and the World Climate Research Programme's Global Land Atmosphere System Study (GLASS), the Land-Use and Climate, Identification of Robust Impacts (LUCID) project was launched. The important word in 'LUCID' is 'robust' – the project was designed to identify those impacts that constitute a genuine signal, as opposed to noise generated by model variability. A second level of 'robustness' was to explore which of the apparent signals were consistent across multiple climate models.

Seven climate models were used to perform the LUCID experiments. Two experiments were conducted using prescribed inter-annually and seasonally varying SST and sea-ice extent using data from the C20C project (HadISST1.1, *ftp://www.iges.org/pub/kinter/c20c/HadISST/*). Firstly, present-day simulations, with all

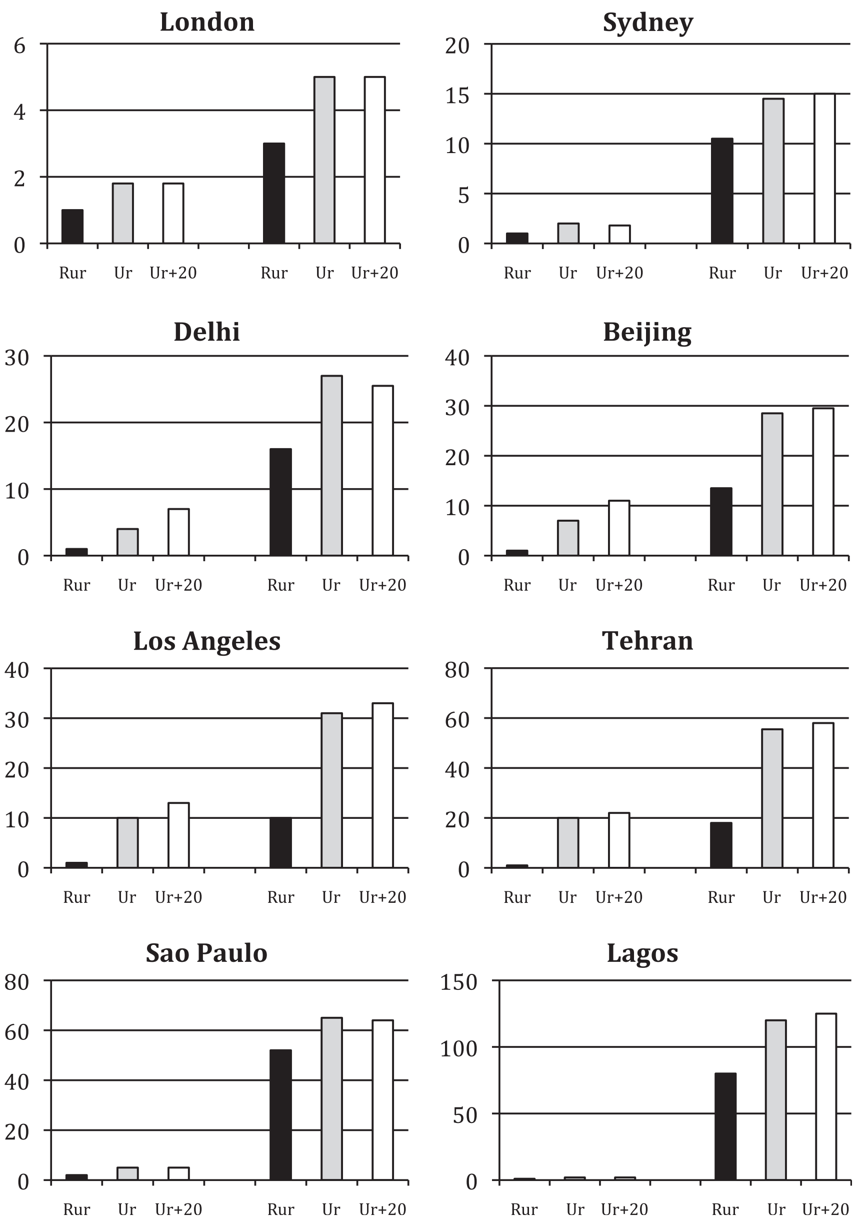

FIGURE 4.6 Annual frequency of hot nights for 1 × CO_2 (left three bars) and 2 × CO_2 (right three bars). The *x*-axis shows rural (Rur), urban (Ur), and urban combined with a 20 W m^{-2} additional heat flux (Ur + 20). *(Source: Modified from McCarthy et al., 2010.)*

GHGs, land cover, and SSTs prescribed at their present-day values were conducted. The land cover was prescribed using a map reflecting 1992 and the period 1972–2002 is simulated. This was then followed by an identical experiment, except that land cover was changed to reflect 1870 conditions. Critically, to address the signal-to-noise ratio, each modelling group was required to conduct at least five independent realizations. The analysis used the Findell et al. (2007) method of a modified Student's t-test (Zwiers and von Storch, 1995) that accounts for autocorrelation within the time series, reducing the rate of false positives.

The preliminary results from LUCID were presented by Pitman et al. (2009). The key result is shown in Figure 4.7. Here, for every climate model, there is a strongly, statistically-significant impact of LULCC on the simulated summer latent heat flux over the regions where LULCC was changed. In Figure 4.7, to be significant, values have to be well in excess of the horizontal line drawn at 5% since, with a 95% significance level, 5% of grid points should appear significant by chance. Clearly, over the regions of LULCC for both the latent heat flux and temperature (2 m air temperature) of every model greatly exceeds this 5% threshold. Given that seven models performed this experiment, the conclusion that LULCC affects the latent heat flux and temperature over regions affected by large-scale LULCC is *robust*. However, the direction of summer temperature change was inconsistent among the models,

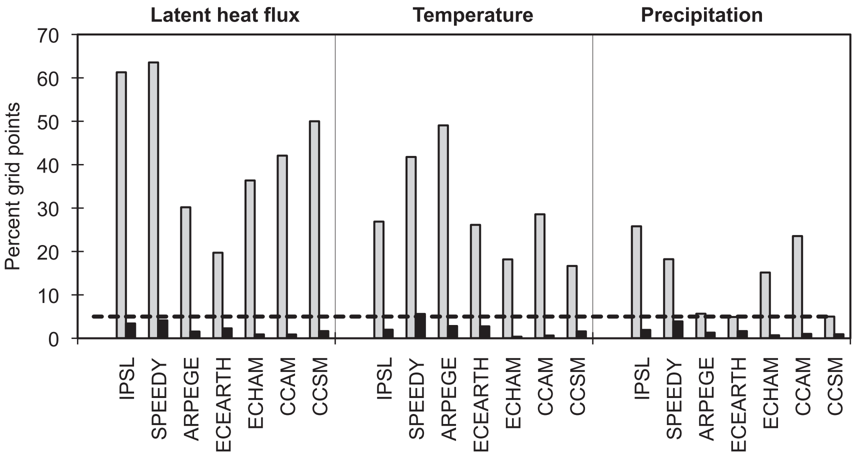

FIGURE 4.7 Percentage of land area that exhibits statistically significant changes in June-July-August (JJA) for latent heat flux, temperature, and precipitation. Light grey bar is the percentage of grid points with statistically significant changes where land cover changes (change in LAI > 0.5) within each climate model. The solid bar is the percentage of grid points with statistically significant changes where land cover is not changed. The horizontal line is the 5% significance level, expected by chance. *(Source: Modified from Pitman et al., 2009.)*

undermining our confidence in the robustness of this result. In the case of rainfall, as noted by Seneviratne et al. (2010) it is less clear whether LULCC affects rainfall. Four of the climate models used in LUCID (IPSL, SPEEDY, ECHAM, and CCAM) show a strongly significant impact on rainfall over regions of LULCC. The other three models do not show impacts above the level expected by chance. In terms of the results from LUCID, therefore, the hypothesis that LULCC affects regional rainfall in statistically significant ways remains unproven.

In Figure 4.7, the percent of land area significantly affected by LULCC *but remote from the actual LULCC* is shown by the solid black bars. For none of latent heat flux, temperature, or rainfall does a single model suggest a scale of impact larger than expected by chance. In terms of LUCID experiments, therefore, there was no evidence to support a hypothesis of a significant impact on these quantities remote from the regions of LULCC.

Is this conclusive? – unfortunately not. Firstly, LUCID used seven climate models and, even if this is a very large sample size for LULCC experiments, it is probably not large enough to form a reliable sample. Secondly, LUCID used fixed SSTs that may have suppressed teleconnections, particularly over the oceans. Thirdly, LUCID used historical LULCC that is mainly focused away from the tropics. Many of the experiments that have identified global teleconnections (e.g., Henderson-Sellers et al., 1993; Zhang et al., 1996; Gedney and Valdes 2000; Werth and Avissar, 2002, 2005) emphasize the global implications of future extensive tropical deforestation. Finally, while LUCID prescribed a common geographical distribution of LULCC, it could not require each model to represent LULCC in identical ways.

However, while LUCID is not conclusive in determining whether LULCC leads to large-scale teleconnections, it is conclusive in several other ways. Firstly, it highlights that any experiments exploring the impact of LULCC on a phenomenon is unreliable if only a few models are used. Thus, while exploration of the impacts of LULCC on the monsoon (Takata et al., 2009) or on extremes (Zhao et al., 2002; Deo et al., 2009; Douglas et al., 2009) are important first steps, the results are not conclusive. Secondly, it demonstrated that precisely how LULCC was implemented in a climate model mattered and that seven modelling groups could choose different, but equally legitimate, ways to do it. This leads to an unresolved research challenge of how best to impose LULCC on a climate model, which is discussed later.

4.4.5. Implications of LULCC for Future Simulations; Feddema et al. (2005)

Several authors have explored the impact of future LULCC under enhanced GHG levels (Costa and Foley 2000; Zhao et al., 2001; Defries et al., 2002; Narisma and Pitman, 2004; Bala et al., 2007; Paeth et al., 2009). Several groups have tried to predict the scale of future LULCC. Figure 4.1g and 4.1h show one estimate derived using the IMAGE model (see *http://www.pbl.nl/en/themasites/image/index.html*). Relative to 2000, the changes estimated to 2050 are mainly focused on the edges of regions yet to be cleared (the Amazon and Congo Basins) and a general intensification of areas already partially cleared. This continues through to 2100, reducing the size of the Amazon and affecting most of the Congo and other tropical forests in the tropics.

Feddema et al. (2005) linked future changes in GHGs consistent with AR4-style emission scenarios with future changes in LULCC. Using a single climate model (the Parallel Climate Model at a resolution not stated) they added the effects of LULCC into simulations conducted using the A2 (relatively high) and B1 (relatively low) emissions scenarios (Nakicenovic et al., 2000). These simulations were then compared with identical future simulations that lacked estimates of future LULCC (i.e., in common with AR4 projections). They conducted transient simulations.

Feddema et al. (2005) found that the impact of LULCC varied depending on the emissions scenario used. They also showed that the impacts of LULCC were negligible on the global average (due to offsetting regional signals) and that most significant climate effects were associated directly with LULCC in tropical and temperate regions, but were indirectly linked to local LULCC at higher latitudes. Feddema et al. (2005) is a key step forward in the study of LULCC and the interactions with climate because it was the first attempt to integrate LULCC with IPCC emission scenarios. However, the statistical methods used, coupled with the single realizations performed and the use of a single climate model, means their results are preliminary. For example, comparing figure 1a in Feddema et al. (2005), which shows the change in land cover for 2050 under a low emissions future, with the first panel in their figure 2 (Northern Hemisphere summer change in temperature) points to a problem. The changes in LULCC included in the model are very small in the Southern Hemisphere with no changes south of about 40°S. Despite this, among the largest regions displaying warming is Antarctica, which warms by 0.5−1.0°C almost everywhere and by up to 2°C (and statistically significantly) at the continental margin. This is quite a common signal in climate model experiments, as are the large changes over the Arctic. These commonly disappear if multiple realizations are performed and statistical tests accounting for autocorrelation are used. Given that these signals are therefore probably noise, it casts doubt on the other apparently statistically significant signals identified by Feddema et al. (2005). To be clear, this does not mean that Feddema et al. (2005) are wrong; it merely means that one set of simulations cannot be conclusive and their leadership in conducting these simulations needs to invigorate other groups to follow. What is clear from their experiments is that over regions of intense LULCC, in common with many earlier studies, a strong and coherent signal is obtained for temperature. Since the impacts of climate change will be felt most strongly at regional-scales and that in many regions LULCC leads to warming that is additive to the CO_2 signal, the conclusion in Feddema et al. (2005) that their results demonstrate the importance of including land-cover change in forcing scenarios for future climate change studies seems entirely defensible.

4.5. LAND USE AND UNDERSTANDING OUR FUTURE CLIMATE

The context of this chapter is the biogeophysical (not biogeochemical) impacts of LULCC (Dickinson (2012, this volume) covers aspects of biochemistry). In the context of the biogeophysical processes related to LULCC, we can address the question, “Is LULCC important?” This can be answered definitively, but the answer depends on the context of the question. At one extreme is the question, “Is LULCC important to the global climate sensitivity to a doubling of CO_2?” Answered relative to CO_2, the answer from available numerical experiments seems to be, “No”. It cannot be entirely ignored due to the release of CO_2 via LULCC, but suggestions that global climate sensitivity would be substantially different if LULCC was accounted for more explicitly is a hypothesis that lacks significant supporting evidence. At the other extreme is the question, “Is LULCC important to the regional climate (or regional climate extremes, or regional climate sensitivity to increasing CO_2)?”. The answer to this question is definitively, “Yes”. LULCC, in this context can include deforestation (e.g., Pitman et al., 2004; Narisma and Pitman, 2003), agriculture (Lyons et al., 1996); irrigation (de Rosnay et al., 2003; Boucher et al., 2004; Douglas et al., 2009; Sacks et al., 2009), urbanisation (Shepherd, 2005; Trusilova et al., 2009), and fire (Görgen et al., 2006). Even where the impact is currently small, it is likely to grow in the future (e.g., Sahel: Taylor et al., 2002). Evidence from almost every region where intensive LULCC has been explored points to a significant impact (China: Gao et al., 2002, 2003; Europe: Heck et al., 2001; United States: Bonan, 1997, 1999, 2001; Australia: Lyons et al., 1996; Narisma and Pitman, 2003; McAlpine et al., 2009).

Between these two extremes are more difficult areas of the science to navigate, where there are no definitive answers. Some of these have become contentious as groups have worked to win an argument over the realism of global teleconnections. Some groups find clear teleconnections (e.g., Henderson-Sellers et al., 1993; Zhang et al., 1996; Gedney and Valdes, 2000; Werth and Avissar, 2002, 2005), while others do not (Findell et al., 2007, 2009; Pitman et al., 2009). There is an imperfect resolution to this disagreement. Most groups that find teleconnections convert the whole of the Amazon (and perhaps other tropical forests) from dense, dark, and lush forest to short bright grass − a sensitivity experiment of value in understanding the potential response of the climate, but not an attempt at a reasonable perturbation based on observations or projections of future land-cover change. Groups that do not find clear teleconnections tend to explore observed changes in LULCC and the scale of that change is small in comparison with the aforementioned tropical deforestation experiments. Thus, Findell et al. (2007) and Pitman et al. (2009) explore the impact of observed LULCC (i.e., little LULCC in the tropics) *and* use statistics that account for autocorrelation to minimize false positives and do not find teleconnections. This does not mean teleconnections will not occur if LULCC becomes intensive in the future, though Feddema et al. (2005) do not really provide strong evidence that they will. A reasonable set of hypotheses is therefore that:

- LULCC directly affects regions under LULCC in terms of temperature, turbulent energy fluxes, soil moisture, runoff, carbon budget, and perhaps rainfall. Importantly, extremes might be quite strongly affected (e.g., heatwaves, floods, and drought; cf. Alexander and Tebaldi, 2012, this volume).
- LULCC does not affect the global climate sensitivity to future (end of the twenty-first century) increases in atmospheric CO_2 significantly (e.g., Harvey, 2012, this volume).
- Historical LULCC does not affect the climate of regions geographically remote from regions of LULCC if the perturbation imposed is realistic; future changes may generate remote changes depending on the scale of LULCC.

The second and third hypotheses are contentious in that they are expressed negatively and many groups would very strongly disagree (although note that they are expressed as hypotheses). Several recent papers have pointed out methods to explore these hypotheses in rigorous ways that would generate a clearer understanding of the science. Building from statements in Findell et al. (2009) and Pitman et al. (2009), one way of addressing the remaining uncertainties would be a coordinated set of experiments designed along the following lines. We assume that these would be conducted using coupled climate models but this would be computationally expensive and a well-designed regional climate modelling initiative could address many of these questions.

1. A rigorous evaluation of LSMs uncoupled from the host climate models (i.e., off-line) needs to be conducted. Do the LSMs capture the contrast between natural and anthropogenic land cover? Can they simulate crops, recovery post-deforestation, and other significant processes? Do they properly simulate the impact of LULCC on fluxes? Off-line skill in capturing the impact of LULCC is clearly a necessity if the model is to capture the impact in a coupled model; it is, however, an insufficient criterion and can only be a first step in any systematic evaluation of the impact of LULCC on climate.
2. A climate model's sensitivity to LULCC needs to be placed in context. If, for example, a model's climate changes by 10°C due to a tweak in a specific parameterization, then it would suggest that it is an overly sensitive model. There are several simple sensitivity experiments that could be performed to determine the models' background response to a perturbation, but the simplest would be to know the impact of a doubling of atmospheric CO_2 in the model. It is important to obtain this signal with *precisely* the same model as is used for an LULCC experiment, and this should be conducted with a fully coupled model to obtain a value of climate sensitivity consistent with IPCC estimates. An alternative would be to determine the climate model's response to a known ENSO-like SST anomaly. Teleconnection patterns from ENSO are quite well-known and a model should be able to broadly capture these. If a climate model captures teleconnections induced by an ENSO anomaly well, given it is a tropical anomaly, then it is hard to assert that teleconnections induced in the same model by a tropical anomaly caused by LULCC are unrealistic.
3. The impact of LULCC is likely to be significantly affected by the strength of the land–atmosphere coupling (Koster et al., 2004; Seneviratne et al., 2006). It is, therefore, helpful to know the coherence between each individual model's pattern of coupling strength and LULCC to understand how a model responds to LULCC.
4. Impose an LULCC anomaly *either* based on observed changes to date, or projected into the future. Any such anomaly should be realistic (e.g., Feddema et al., 2005) not 'scorched earth'.
5. While it is computationally expensive, multiple realizations are required to allow the signal from LULCC to be identified against a given climate model's background (natural) variability. LUCID used five realizations, but also used fixed SSTs. If transient experiments are run then, the number of realizations most likely has to exceed ten.
6. While it is organisationally difficult, multiple models should be used. LUCID used seven climate models but this is not enough to form a reliable sample. The number of models used is defined by community interest and resources and needs to be as inclusive as possible. The models used also have to be current versions of a given climate model to allow the relative role of LULCC to be compared with CO_2.

The effort involved in implementing this set of experiments is very considerable and we recognize this. However, unless we resolve the role of LULCC at regional-scales, the ability to project regional-scale changes resulting from other forcings, including CO_2, will remain limited, at least in regions affected by LULCC. If it can be demonstrated that LULCC does not trigger large-scale teleconnections, regional climate models (see Evans et al., 2012, this volume) may provide an alternative methodology, but this would still require an experimental design that would be computationally demanding.

This brings us to the scientific challenge of imposing a realistic LULCC on a model. When imposing an LULCC, two (or more) maps are used to describe the state of the

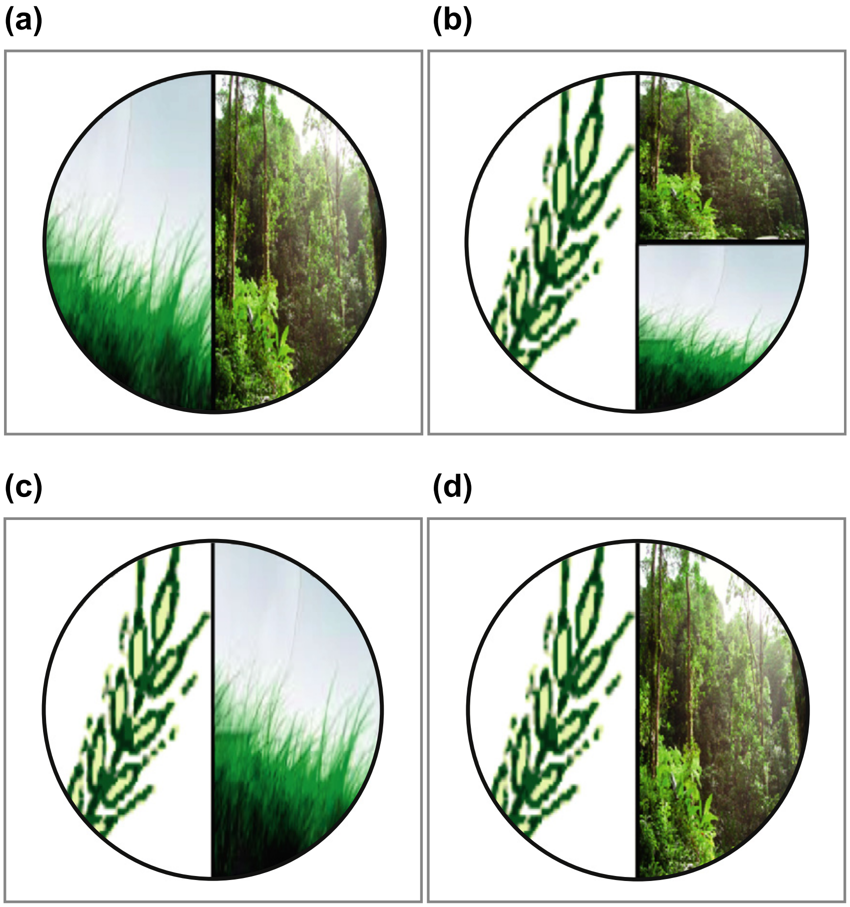

FIGURE 4.8 Illustration of how 50% of a natural landscape (50% forest, 50% grass) can be converted into very different combinations of vegetation via different, but equally valid, decisions: (a) original natural distribution; (b) each original vegetation type has been proportionally reduced; (c) all forests have been cleared and grasslands have remained untouched (for e.g., grazing); and (d) all grasslands have been converted into cropland while forests have remained untouched.

vegetation at various times. It is not straightforward for all modelling groups to use the same map. It is also not straightforward to implement these maps in a homogeneous way. There are multiple issues:

- It is common to reduce the natural vegetation proportionally, hence, if the original landscape was 50% trees and 50% grass (Figure 4.8a) and 50% of this area becomes agricultural then the new fractions would be 25% trees, 25% grass, and 50% crops (Figure 4.8b). But why? It may be as legitimate to remove all the trees and replace just these with crops (Figure 4.8c) on the grounds that a farmer might already use the grasses for grazing and would add productive land by removing the trees. Alternatively, it may be as legitimate to remove all the grasses and replace these with crops. Indeed this would be much easier and quicker for the farmer (Figure 4.8d). A modeller can therefore end up with a 50% tree cover (as in the natural state, Figure 4.8a or the perturbed state, Figure 4.8d), or 25% tree cover (Figure 4.8b) or no tree cover (Figure 4.8c) via equally reasonable judgements. This means that, while common maps might be provided, how the agricultural land is to be created has to be agreed. This is critical in regions of snow (Betts, 2000), but is also likely to be important elsewhere.
- A similar issue is that it is difficult for most modelling groups to change their standard vegetation maps. Climate models are built with a particular vegetation map in mind and with decisions on how many vegetation types will be represented. A climate model may, in some ways, be calibrated to produce a good climate of the twentieth century with a specific vegetation map and a particular number of vegetation types. If one LSM allows for 5 vegetation types and another allows for 20, and maps are provided with 20 classes of vegetation, the implementation of these maps will mean different things in the 2 different models.
- The representation of crop phenology in LSMs is important. Some LSMs prescribe phenology using present-day satellite-derived datasets; others simulate seasonal phenology with implicit or calculated dates for cropping and harvest; and others simply describe crops as natural grassland, but with a different set of parameter values, without representing harvesting. Some

LSMs only represent natural vegetation and describe pasture and crops as a single type of grass. Some simulate bare soil between harvest and sowing, others simulate grass. These all affect the detail of how a given LULCC is implemented in a climate model. These difficulties are further confounded by disturbance to the phenology caused by global warming (e.g., Schneider and Root, 2002).

Pitman et al. (2009) concluded that the expression of LULCC in a climate model depends on how vegetation types are parameterized, how the LSM tiles the surface (there are several approaches), how land covers are actually implemented, which parameters are fixed, which are time-varying, how these differ between LSMs, and how strongly the surface is coupled to the atmosphere (Seneviratne et al., 2006). An identical land-cover perturbation is, therefore, impossible to impose on all models in this context, but a common set of procedures by which a similar LULCC perturbation can be imposed on multiple climate models is possible. This is what has been suggested by a group of scientists for the upcoming IPCC simulations (CMIP5) and would enable a multimodel approach to assess the impact of LULCC on climate.

This leaves us with a problem. Multiple climate models must be used to explore the scale of the impact of LULCC. The imposition of LULCC on these models needs to be as similar as possible, otherwise differences in the impact would be the result of the way LULCC was implemented rather than the impact of a realistic LULCC signal on climate. At present, we cannot do this – there are too many uncertainties in how to model terrestrial processes, how to implement LULCC, and how the land is coupled to the atmosphere to resolve the real signal from LULCC on climate. However, we *know* that LULCC has a large impact on regional climate in regions of intensive LULCC. So, as noted by Feddema et al. (2005) and by Pitman et al. (2009), it is necessary for regional predictions that LULCC is included in AR5 climate projections, and that these changes should include urbanisation (McCarthy et al., 2010; Cleugh and Grimmond, 2012, this volume).

The logical conclusion from this is that including LULCC in AR5 climate models may increase the differences between models *over regions of intensive LULCC*. Clearly, there is a need to design a common numerical experiment, such as the classic 'doubling CO_2' experiment to diagnose the actual sensitivity of individual climate models to LULCC. Such an experiment would encompass, simultaneously, the way the land-atmosphere interactions are parameterized and the intensity of the coupling. Such an experiment would need to include a standardized land forcing that can be applied to all models and avoid potential differences arising from variations in the implementation and/or parameterizations.

In summary, LULCC is a key component of the Earth's climate in terms of the net carbon balance, emissions of CO_2, regional impacts on radiative forcing, and impacts on the partitioning of available energy between sensible and latent heat. Collapsing the role of LULCC into radiative forcing is an invalid simplification (Davin et al., 2007; Davin and de Noblet-Ducoudré, 2010). Radiative forcing is a poor measure because it is only one part of the impact of LULCC. The substantial effects LULCC has on the partitioning of available energy between sensible and latent heat and, the way precipitation is partitioned between evaporation and runoff, is not captured by changes in radiative forcing. Furthermore, global measures of the impact of LULCC are not appropriate since different regions commonly experience impacts of different sign such that a global average is negligible (Feddema et al., 2005). A global measure, such as radiative forcing, is useful for the change in energy due to elevated CO_2 but, as has been convincingly argued by Pielke et al. (2002), it is not for LULCC.

LULCC, while important, is probably only important at specific scales. It is also particularly difficult to resolve in global climate models. There is no doubt that LULCC has a large impact on some regions (but not all) that have been intensively affected by LULCC. It is also very unlikely that it has a strong effect on the global climate sensitivity. In-between, the question of the scale of remote impacts of LULCC remains unresolved because almost all experiments have used modelling approaches that leave difficult problems of statistical significance and the reasonableness of the imposed changes unresolved. Resolving these problems requires considerable effort and most climate modelling groups invest remarkably small amounts of resources in terrestrial modelling. This is surprising since virtually every important reason for developing climate models relates to terrestrial quantities (such as soil moisture, crop yields, biodiversity, drought, streamflow). The ways forward for the climate modelling communities to resolve the scale of the role of LULCC on climate is clear, but it requires a base-level of effort that is not commensurate with current investment in this area and, therefore, represents a real and continuing challenge for the regional prediction of the impacts of various types of forcing on quantities of direct relevance to humans.