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Review

Land cover changes and their biogeophysical effects on climate

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ABSTRACT: Land cover changes (LCCs) play an important role in the climate system. Research over recent decades highlights the impacts of these changes on atmospheric temperature, humidity, cloud cover, circulation, and precipitation. These impacts range from the local- and regional-scale to sub-continental and global-scale. It has been found that the impacts of regional-scale LCC in one area may also be manifested in other parts of the world as a climatic teleconnection. In light of these findings, this article provides an overview and synthesis of some of the most notable types of LCC and their impacts on climate. These LCC types include agriculture, deforestation and afforestation, desertification, and urbanization. In addition, this article provides a discussion on challenges to, and future research directions in, assessing the climatic impacts of LCC.

KEY WORDS land cover change; climate; biogeophysical impacts

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1. Introduction

Land cover change (LCC) has significant impacts on the earth's climate, hydrology, water resources, soils, and biota (Foley *et al.*, 2003b; Lambin *et al.*, 2003; DeFries *et al.*, 2004; Twine *et al.*, 2004; Scanlon *et al.*, 2005, 2007, Zhang and Schilling, 2006; Cotton and Pielke, 2007; Pereira *et al.*, 2010). Despite some uncertainties in the magnitude of the impacts, it is increasingly recognized as an important forcing of local (Landsberg, 1970; Balling, 1988; Segal *et al.*, 1989b, Rabin *et al.*, 1990; Balling *et al.*, 1998; Arnfield, 2003; Campra *et al.*, 2008; NRC, 2012), regional (Barnston and Schickedanz, 1984; Zheng *et al.*, 2002; Foley *et al.*, 2003a; Mohr *et al.*, 2003;

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Oleson *et al.*, 2004; Voldoire and Royer, 2004; Gero *et al.*, 2006; Ray *et al.*, 2006; Betts *et al.*, 2007; Costa *et al.*, 2007; Abiodun *et al.*, 2008; Klingman *et al.*, 2008; Lee *et al.*, 2008; Nuñez *et al.*, 2008; Kvalevåg *et al.* 2010; Hu *et al.*, 2010), and global climate (Franchito and Rao, 1992; Wu and Raman, 1997; DeFries *et al.*, 2002; Kabat *et al.*, 2004; Avissar and Werth, 2005; Feddema *et al.*, 2005; *NRC*, 2005; Gordon *et al.*, 2005; Cui *et al.*, 2006; Ramankutty *et al.*, 2006; Takata *et al.*, 2009; Sacks *et al.*, 2009; Puma and Cook, 2010; Davin and Noblet-Ducoudré, 2010; Strengers *et al.*, 2010; Lee *et al.*, 2011; Lawrence *et al.*, 2012). As with carbon dioxide (CO_2), LCC affects the climate system on multi-decadal time scales and longer. In a recent global-scale modelling study, Avila *et al.* (2012) demonstrated that impacts of LCC on indices of temperature extreme were equal to the impacts of doubling of CO_2. In some regions, impacts were similar to forcing of CO_2 while in others they were opposite. Hence, LCC can dampen or enhance the impacts of increasing CO_2 and as a result, it would not be prudent to explain future changes in temperature extremes and other climatic metrics only by increasing CO_2. LCC is, thus, of primary concern in any assessment of climate processes, and involves land surface conversions such as the following: forest to agriculture, reforestation of formerly agricultural areas, afforestation, grassland to irrigated agriculture, rural to suburban, and suburban to fully built-up. Ramankutty and Foley (1999) noted that approximately 12 million km^2 of forests and woodlands have been removed globally since 1700 AD. They estimated that about 18 million km^2, or 11% of the global land area, is currently under farming. This is approximately the size of the entire South American continent (Ramankutty and Foley, 1999). An example of agricultural expansion over the last 500 years can be found in Figure 1. In 2000, livestock grazing represented 22% or 28 million km^2 of the global land area (Ramankutty *et al.*, 2008), which can trigger desertification in semi-arid regions (NRC, 1992). Moreover, Hansen *et al.* (2008; 2010) estimated that there has been 0.27 million km^2 of humid tropical forest loss and 1.10 million km^2 global gross forest cover loss between 2000 and 2005.

These transformations of the Earth's surface fundamentally alter the fluxes of solar and thermal infrared radiation, sensible, and latent heat, the movement of water between the sub-surface and atmosphere, and the exchange of momentum between the land-surface and atmosphere. Alterations such as these occur on spatial scales ranging from the patch or micro- (10^{-2} to 10^3 m) to sub-regional (10^4 to 2×10^5 m) scales (e.g. Anthes, 1984; Oke, 1987) (Figure 2). They may result in modifications of surface albedo, which also alters the near surface energy balance (Zeng and Neelin, 1999; Hoffman and Jackson, 2000; Berbet and Costa, 2003; Zhang *et al.*, 2009b) and the thermal climate (e.g. Oke, 1987; Bonan, 2001, 2008a; Juang *et al.*, 2007). The physical climate modifications manifest as spatial heterogeneities of temperature, humidity, and wind speed. High or low albedo may result in lowered or increased temperature, respectively, due to greater reflection of shortwave radiation or, conversely, higher amounts of shortwave radiation absorption (e.g. Otterman, 1974; Otterman *et al.*, 1984; Lofgren, 1995; Sailor, 1995, Bonan, 2008a). Increased transpiration from a vegetated area also means increased and decreased fluxes of latent and sensible energy, respectively, and a resultant lowering of surface maximum temperatures (e.g. Barnston and Schickedanz, 1984; Geerts, 2002; Ter Maat *et al.*, 2006; Kueppers *et al.*, 2007; Biggs *et al.*, 2008; Ozdogan *et al.*, 2010). In addition, modified climatic phenomena have been observed along and near the boundaries of land cover type transitions, with horizontal gradients of climate variables intensifying, and an alteration of mesoscale vertical circulations within the planetary boundary layer (PBL) that enhance the vertical movement of air (e.g. Segal and Arritt, 1992; Weaver and Avissar, 2001). These greater upwards vertical motions in the PBL may be realized through convective cloud development, and even precipitation, given favourable larger-scale atmospheric conditions. The latter include weak stability and slow background synoptic winds or winds that blow parallel to landscape boundaries (Carleton *et al.*, 2001; Pielke, 2001; Weaver and Avissar, 2001).

Given this context, the primary aim of this article is to review the role of LCC in the climate system. We particularly focus on biogeophysical impacts of LCC. Examples of biogeophysical properties include surface roughness, leaf area index (LAI), vegetation stomatal resistance, and albedo. LCC leads to modifications of these properties resulting in changes in energy, moisture and momentum fluxes. Highlighted examples of impacts include both long-term systematic changes (e.g. agricultural land use change, deforestation, reforestation and afforestation), and short-term abrupt changes (e.g. rapid urbanization). The literature reviewed includes both observational and model-based studies. Finally, we provide a synthesis of results from these studies and discuss critical challenges in LCC-climate research, and make a series of recommendations related to better detecting LCC from observed climatic records and improving modelling approaches for understanding climate impacts of LCC.

2. The role of LCC within the climate system

As indicated above, changes in land cover result in alterations to surface moisture, heat, and momentum fluxes, as well as trace gas exchanges such as CO_2. These changes result in a different PBL structure, cloud cover regime, and indeed all other aspects of local and regional weather and climate (Pielke and Avissar, 1990; Rabin *et al.*, 1990; Pielke, 2001; Fu, 2003; Wang *et al.*, 2009). If sufficiently large areas are affected, then changes in climate occur not only locally, but also in regions remote from the original landscape modification (e.g. NRC, 2005; Cui *et al.*, 2006; Niyogi *et al.*, 2010; Snyder, 2010).

Given this context, the surface energy and moisture budgets for bare and vegetated soils are critical to

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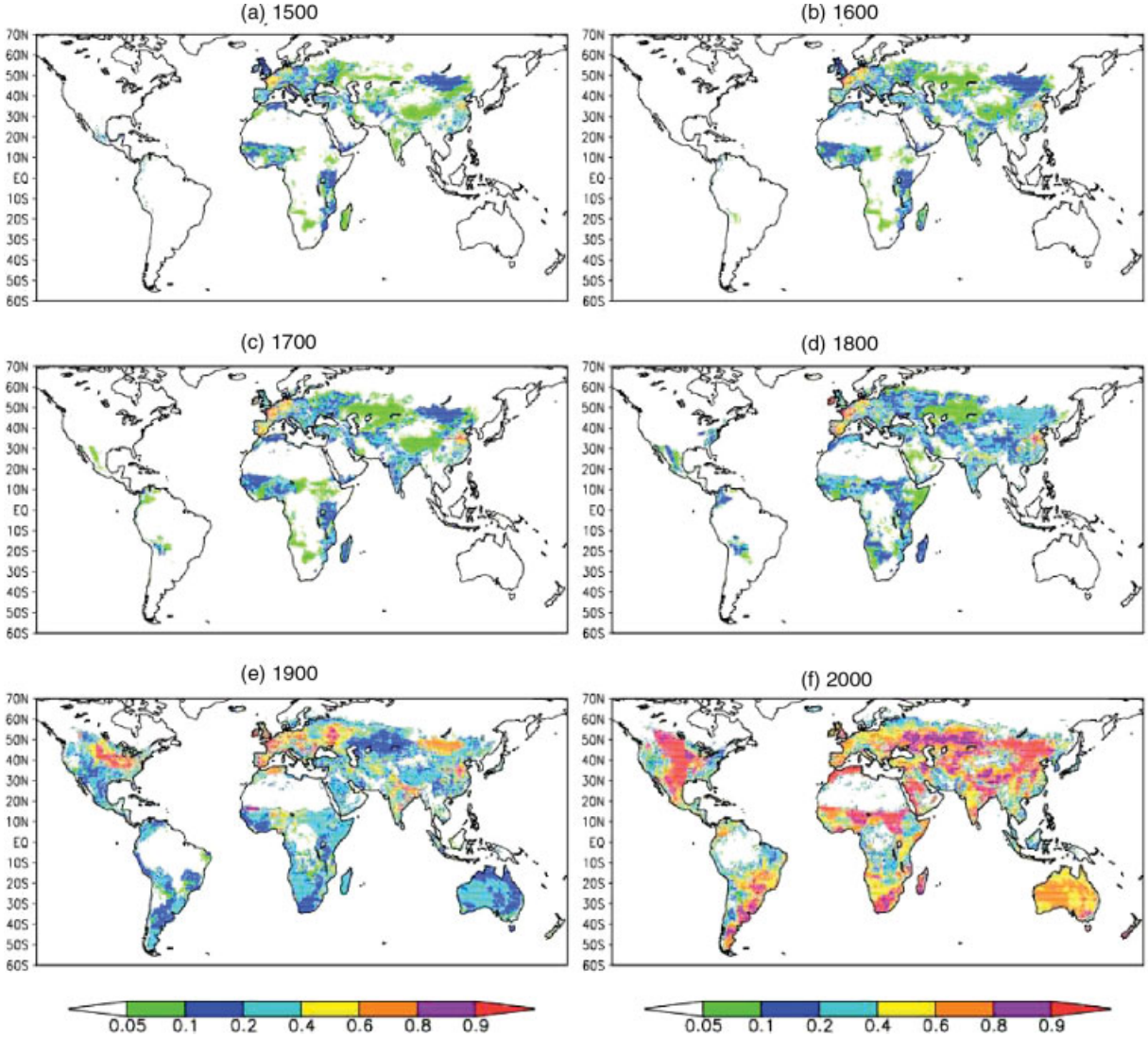

Figure 1. LCC for various time periods. Pasture or crop lands are presented as a fraction. Source of the data is http://luh.unh.edu. Refinement of the data has continued (e.g. much of central Australia is ungrazed or low density grazing and shown as pasture) (Source: Pielke *et al.*, 2011).

understand the impacts of LCC, and can be written after Pielke (2001):

$$R_N = Q_G + H + L(E + T) \tag{1}$$

$$P = E + T + RO + I, \tag{2}$$

where R_N represents the net radiative fluxes $= Q_s(1 - A) + Q_{LW}^{\downarrow} - Q_{LW}^{\uparrow}$; Q_G is the soil heat flux; H is the turbulent sensible heat flux; $L(E + T)$ is the turbulent latent heat flux; L is the latent heat of vapourization; E is physical evaporation (conversion of liquid water into water vapour by non-biophysical processes, such as from the soil surface and from the surfaces of leaves and branches); T is transpiration (the phase conversion to water vapour, by biological processes, through stoma of plants); P is the precipitation; RO is runoff; I is infiltration; Q_s is insolation; A is albedo; $Q_{LW}^{\downarrow}$ is the downwelling longwave radiation; $Q_{LW}^{\uparrow}$ is upwelling longwave radiation $= (1 - \varepsilon)\, Q_{LW}^{\downarrow} + \varepsilon\sigma\, T_s{}^4$; ε is the surface emissivity; σ is the Stefan-Boltzmann constant ($5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$); and T_s is the surface temperature. The direction of the fluxes is conventionally defined such that receipt at the surface is positive and loss from the surface is negative. Equation (1) is a budget equation, however, the sources and sinks need to be of opposite sign.

Equations (1) and (2) are not independent of each other. A reduction in E and T in Equation (2), e.g. increases Q_G and/or H in (1) when R_N does not change. Reduced E and T can occur, e.g. through clear-cutting of a forest and the subsequent increase in runoff. The precipitation rate and type also influence how water is distributed between runoff, infiltration, and interception by plant surfaces.

Any LCC that alters one or more of the variables in Equations (1) and (2) has the potential to affect the climate directly. For instance, a decrease in albedo (i.e. a darkening of the surface) by afforestation or irrigated agriculture, increases R_N and thus makes more energy

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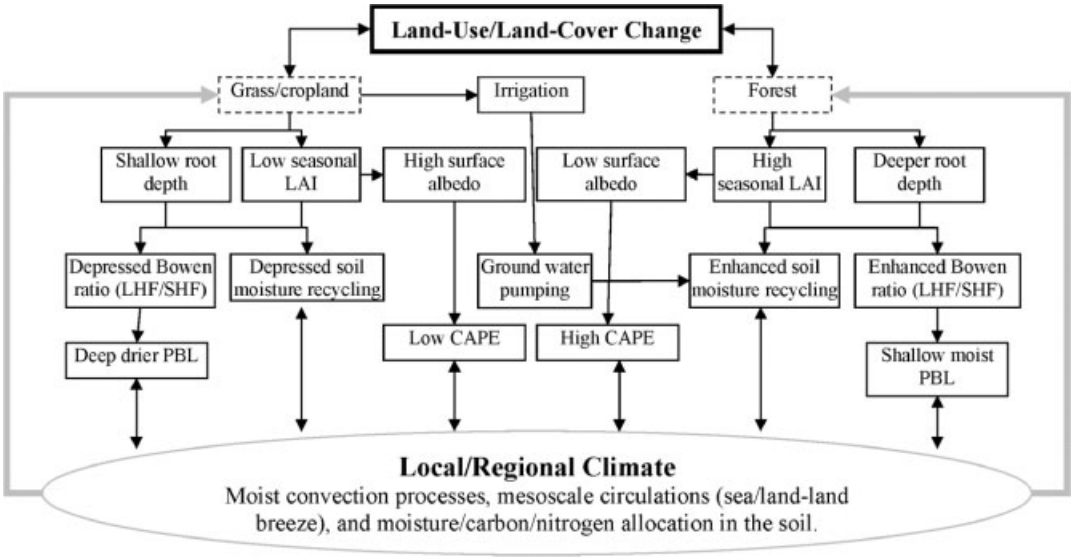

Figure 2. Conceptual model of the impacts of LCC on local and regional climate (Source: Pielke *et al.*, 2007).

available for Q_G, H, E, and T. These changes also modify energy partitioning amount into H versus E and T or Bowen Ratio $(=H/[L(E+T)]$, Bowen, 1926; Oke, 1987). In other words, lower Bowen ratio refers to a moist environment. Once the surface energy budget is altered, fluxes of heat, moisture, and momentum within the PBL are directly affected (Segal *et al.*, 1989a; Douglas *et al.*, 2006). Local (meso-scale) and regional wind and other weather patterns can subsequently be affected due to horizontal variations in H and PBL depth (Segal and Arritt, 1992; Leeper *et al.*, 2009). In addition to albedo and partitioning of surface fluxes, the surface biophysical characteristics can also impact the thermal inertia/heat capacity of the land surface. It is noted that nighttime temperatures are more sensitive to heat capacity (see McNider *et al.*, 2005; Shi *et al.*, 2005). High water contents in soils, such as from irrigation, can increase heat capacity as can highly vegetated areas.

Similar budget equations (comparable to Equations (1) and (2)) can be written for carbon and nitrogen fluxes (e.g. Parton *et al.*, 1987; Running and Coughlan, 1988). The carbon budget involves the assimilation of CO_2 into carbohydrates within vegetation, the respiration of CO_2 from plants and animals, decay of animal and plants, industrial and vehicular combustion processes, outgassing from oceans and other water bodies and volcanic emissions. The nitrogen budget has also been segmented into its different components, e.g. by Galloway *et al.* (2004) and Lamarque *et al.* (2005). Each of these budgets will be changed if any characteristic of the land surface is altered. These include both land management caused changes (e.g. deforestation or alterations in the type of agriculture) and phenological changes due to drought and other environmental stresses to vegetation (e.g. increased temperature, attacks by pests and disease, etc.).

The surface energy and moisture budgets and the carbon and nitrogen budgets are closely coupled. Changes in the energy and moisture budget alter the carbon and nitrogen budget (and that of other trace constituents), while alterations in carbon and nitrogen (and other trace gases and aerosols) change the surface energy and moisture budgets. The primary link in the coupling of these fluxes is the transpiration of water vapour through the stoma of plants, which influences changes in the energy budget, and is also involved in the assimilation of carbon into plant leaves, roots, and stems. If the amount of actively growing plant biomass changes, this alters the transpiration of water vapour into the atmosphere and thus the amount of carbon that is assimilated. The amount of nitrogen compounds and other trace nutrients affect plant growth and vitality, as determined from parameters such as biomass, leaf-area index, and photosynthesis. The intimate coupling by feedback processes of the surface budgets is a fundamental regulator of the climate system. In short, land management practices and resulting LCCs that alter any one of these budgets necessarily alter all of them.

There are a number of model based studies conducted addressing comparative biogeophysical and biogeochemical impacts of LCC (e.g. Brovkin *et al.*, 2004; Matthews *et al.*, 2004; Lawrence *et al.*, 2012). Brovkin *et al.* (2004) and Matthews *et al.* (2004) found cooling due to biogeophysical impacts while warming due to biogeochemical impacts. Brovkin *et al.* (2004) have reported that globally averaged biogeochemical change related warming is 0.18 °C while biogeophysical change related cooling is 0.26 °C (net cooling 0.08 °C). However, they have also noted that regional impacts can be significant. On the other hand, Matthews *et al.* (2004) noted a net global warming of 0.15 °C for the combined impact.

The following sections highlight the biogeophysical climatic impacts of some of the most notable types of LCC. We include discussion of the impacts of agricultural land use, deforestation and afforestation, desertification, and urbanization. The discussion within each subsection flows from smaller to larger scales. In addition, modelling

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studies are identified and the rest represent observational data based research.

3. Meso-, regional-, sub-continental- and global-scale impacts

3.1. Changes in fluxes and precipitation

At the meso-scale, LCC-driven urbanization impacts energy fluxes and balance (see the review by Arnfield, 2003). In most of the cases, urbanization results in replacement of natural vegetation with a built environment. As a result, energy flux is dominated by sensible energy flux that leads to development of the Urban Heat Island (UHI). In urban areas the maximum sensible energy flux can be several orders of magnitude higher than latent energy flux (e.g. Grimmond and Oke, 1995; Grossman-Clarke *et al.*, 2010; Hanna *et al.*, 2011). However, the relative magnitude of partitioning into sensible energy flux varies with season, geographical location of the urban area, and within urban area land use variations. The latter can be the central business district (nearly free of vegetation) versus residential area (can be substantially vegetated) (e.g. Masson *et al.*, 2002; Lemonsu *et al.*, 2004; Offerle *et al.*, 2005, 2006; Kawai *et al.*, 2009; Grossman-Clarke *et al.*, 2010; Hanna *et al.*, 2011; Loridan and Grimmond, 2012). In humid temperate regions, the removal of the natural forest and wetlands has resulted in a reduction of transpiration and evaporation and an increase in sensible energy fluxes and the Bowen ratio (e.g. Shepherd, 2006; Caldwell *et al.*, 2012). In arid and semi-arid regions, by contrast, urban areas typically have irrigated landscapes such that the latent energy flux is much larger than the natural desert or steppe landscape (e.g. Segal *et al.*, 1988).

These changes in convective fluxes associated with urban land use also influence meso-scale atmospheric dynamics and stability profiles in such a manner that precipitation is affected (Figure 3). A wealth of historical and contemporary literature shows that UHI-destabilization, canopy-related surface roughness, and/or pollution can independently or synergistically modify, amplify, reduce, or initiate precipitating cloud systems (e.g. Landsberg, 1970; Changnon *et al.*, 1981; Bornstein and Lin, 2000; Shepherd *et al.*, 2002, 2010a, 2010b; Niyogi *et al.*, 2006; Shepherd, 2006; Kaufmann *et al.*, 2007; Mote *et al.*, 2007; van den Heever and Cotton, 2007; Rose *et al.*, 2008; Stallins and Rose, 2008; Trusilova *et al.*, 2008; Hand and Shepherd, 2009; Shem and Shepherd, 2009; Ashley *et al.*, 2012; Mitra *et al.*, 2011; Niyogi *et al.*, 2011). The overwhelming majority of these studies reveal a link between urban areas, convection enhancement, and increased precipitation.

Lei *et al.* (2008), Kishtawal *et al.* (2010) and Niyogi *et al.* (2010) suggested that the heavy rainfall trend is greater over the urban regions of India compared to non-urban areas and this can be verified by both *in situ* and satellite datasets. Mitra *et al.* (2011) and Niyogi *et al.* (2011) noted that increased sensible heat flux, convergence, atmospheric destabilization, and resultant modified atmospheric flow patterns play an important role in enhancing precipitation. However, details of the mechanisms and pathways are not fully understood (Shepherd *et al.*, 2010b).

In a modelling study, Trusilova *et al.* (2008) found statistically significant increases in winter rainfall in

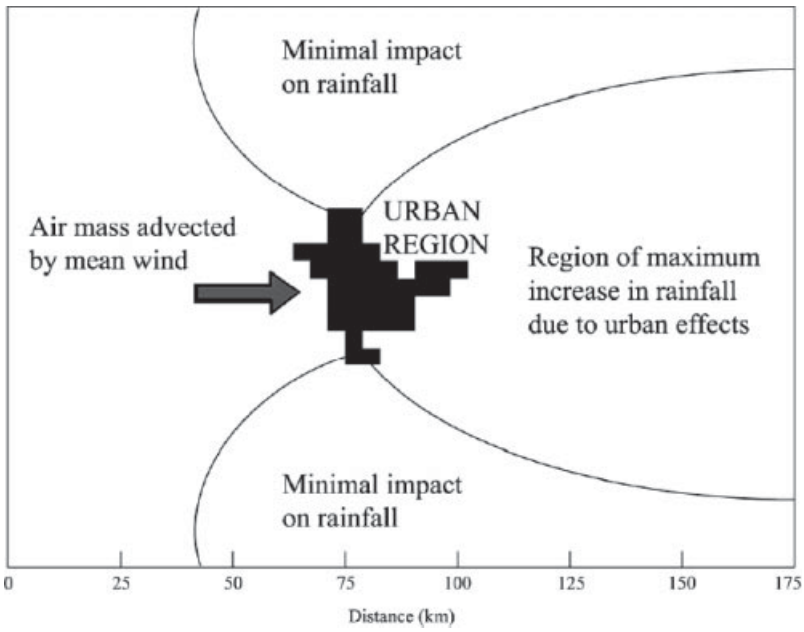

Figure 3. Idealized diagram showing the region of maximum expected rainfall increases due to urban effects located downwind of the city centre. Minimal impact of urban land use on precipitation is observed in the regions perpendicular to the mean wind vector (an adaptation from Shepherd *et al.*, 2002).

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Europe in their urban simulations as compared to the pre-urban settlement simulations. UHI-induced enhancement of convection was partly responsible for increased winter precipitation. Three observational and model-based studies in China examining urban effects on rainfall found decreased cumulative rainfall (Guo *et al.*, 2006; Kaufmann *et al.*, 2007; Zhang *et al.*, 2009a). However, these studies did not consider the effects of atmospheric pollution. Smaller cloud droplet size distributions and suppressed rainfall have been linked to increased aerosol concentrations from anthropogenic sources over and downwind of urban areas (Rosenfeld, 1999, 2000; Givati and Rosenfeld, 2004; Lensky and Drori, 2007). The above results suggest that urban forcing on atmospheric dynamics, thermodynamics, energy exchanges, cloud microphysics, and composition need to be explicitly represented in future modelling systems from local to global scales (Jin *et al.*, 2007; Shepherd *et al.*, 2010b).

Also on the meso-scale, except for vegetated surfaces, the replacement of one land cover type (e.g. grassland, forest) by agriculture (rainfed and irrigated) alters not only the radiative and thermal climates noted earlier, but also the moisture budget (Adegoke *et al.*, 2003, 2007). In particular, the substitution of crops that readily lose moisture to the atmosphere, such as corn (maize) and soybeans in the Midwest United States, may cool the surface sufficiently during daytime hours in summer to promote a downward flux of sensible heat, in contrast to nearby natural vegetation areas which tend to better conserve water (e.g. Hatfield *et al.*, 2007). This 'oasis effect' has been implicated in recent observed increases in summertime extreme dew-point temperatures, and reduced maximum temperatures during the 20th century, in parts of the Midwest United States (Bonan, 2001; Sandstrom *et al.*, 2004). This condition may also promote a greater incidence of severe weather (Pielke and Zeng, 1989) due to increased destabilizing impacts of water vapour on the PBL compared to stabilizing effects of evaporation-induced cooling.

A propensity for increased convective cloud and precipitation development in agricultural areas is not solely a function of replacing a natural surface with one that evapotranspires more readily. The crop phenology, ambient atmospheric moisture content, and background synoptic-scale atmospheric circulation (surface winds and free-atmosphere winds) are also critical. For example, Rabin *et al.* (1990) showed, in the southern Great Plains, that when the atmosphere was dry, convective clouds tended to develop first over the dry wheat stubble and later over more moist – and actively photosynthesizing – surfaces that have higher net radiation values. Conversely, when the atmosphere was more humid, convective clouds tended to develop first over the moister vegetated surfaces, and later over drier surfaces.

A similar dependence of the land cover-convection relationship on surface and atmospheric moisture conditions has been observed in the rain-fed corn and soybean areas of the Midwest United States (Carleton *et al.*, 2001; Allard and Carleton, 2010). Moreover, there is a synoptic (i.e. regional-scale) circulation influence on these local land surface-atmosphere impacts that occurs via the advection of moisture by low-level winds, the sign and magnitude of the free-atmosphere vertical motion of air, and the extent to which moisture is trapped within the PBL (e.g. Bentley and Stallins, 2008; Carleton *et al.*, 2008a; Carleton *et al.*, 2008b; Allard and Carleton, 2010). An observational study found that convective precipitation was enhanced in association with the major crop-forest boundaries in the Midwest Corn Belt (Carleton *et al.*, 2008a, 2008b). The findings of these studies highlight the possible role of contrasting phenology and PBL circulations between crop and tree areas in the vegetation boundary–precipitation relationship.

Numerous other studies document LCC induced changes in surface fluxes around the world. For example, Douglas *et al.* (2006) compared modelled water vapour fluxes in India from a pre-agricultural and a contemporary land cover and found that mean annual vapour fluxes have increased by 17% with a 7% increase in the wet season and a 55% increase in the dry season. In a model sensitivity study, Sen Roy *et al.* (2011) found that latent and sensible heat fluxes could be up to about 40 and 80 $\mathrm{Wm^{-2}}$ higher due to increased and decreased soil moisture, respectively, related to irrigated and non-irrigated conditions in India (Figure 4).

Tuinenburg *et al.* (2011) concluded that large scale irrigation in southern and eastern India may increase local precipitation as a result of land-atmosphere feedbacks. Many other observational and modelling studies for this region that further document changes in precipitation due to LCC include Lohar and Pal (1995), Saeed *et al.* (2009), Niyogi *et al.* (2010), Kishtawal *et al.* (2010), Douglas *et al.* (2009), Lei *et al.* (2008), Lee *et al.* (2009) and Sen Roy *et al.* (2011). Other regions where these effects on regional and sub-continental climate have been shown including the United States (e.g. DeAngelis *et al.*, 2010), Australia (e.g. Nair *et al.*, 2011) and Southeast Asia (e.g. Takahashi *et al.*, 2010). In a detailed observational study, DeAngelis *et al.* (2010) reported that irrigation in the Ogallala aquifer region of the United States has resulted in a 15% increase of July precipitation several hundred miles downwind, including, Indiana and western Kentucky.

Land cover change in southwest Australia impacts boundary layer cloud formation (Lyons *et al.*, 1993; Lyons, 2002; Ray *et al.*, 2003), and micro- (Lyons *et al.*, 2008), meso- and synoptic-scale circulations (Nair *et al.*, 2011). Utilizing observations of surface energy fluxes, Lyons *et al.* (2008) showed a higher possibility of dust devil formation over cleared agricultural landscapes leading to decreases in cloud particle size and reduced probability of rainfall (Junkermann *et al.*, 2009). A modelling study for southeast Australia, a major agricultural region, simulated mean summer rainfall decrease by 4–12% (Figure 3(c) in McAlpine *et al.*, 2009). The authors attributed this change to a significant decrease in evapotranspiration (6.8%), latent heating (7.3%), and total cloud cover, especially low clouds and convective

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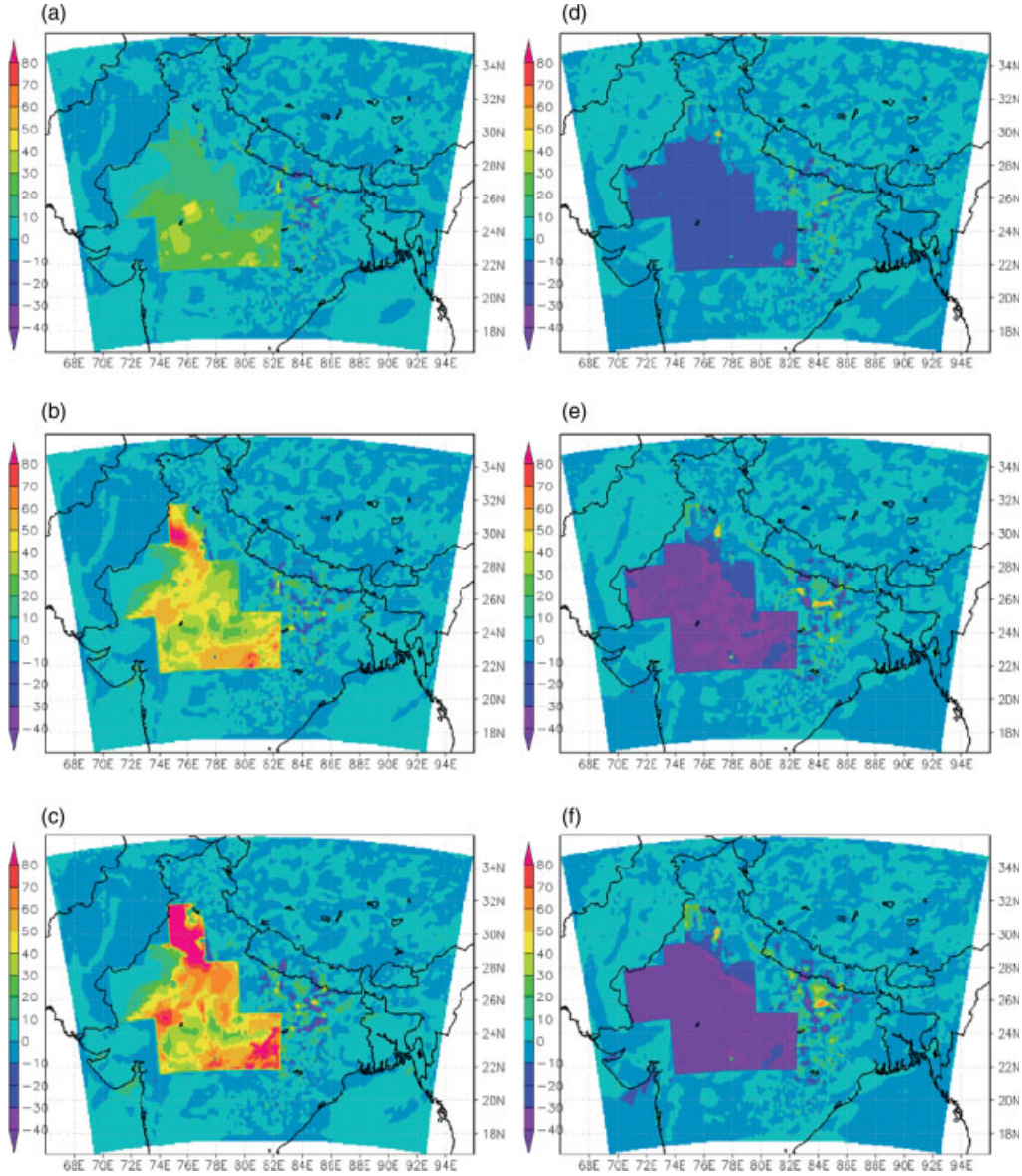

Figure 4. Latent energy flux differences ($W\,m^{-2}$) for 03/23/2000 for relatively drier and wetter (irrigated) conditions. Soil moisture was systematically decreased and increased by 5, 10, and 15% from the current state over irrigated areas. Differences were calculated as: (a) current minus 5% drier, (b) current minus 10% drier, (c) current minus 15% drier, (d) current minus 5% wetter, (e) current minus 10% wetter, and (f) current minus 15% wetter. Various shades of red, orange, and green (a–c) suggest lowering of latent energy flux with increased drying while shades of blue and purple suggest increasing latent energy flux with increased wetting (Source: Sen Roy *et al.*, 2011).

clouds. Analysis of daily rainfall events indicates an increase in the number of dry and hot days, the drought duration, and decreases in daily rainfall intensity and wet day rainfall amounts in southeast Australia, (Figure 2(b); see Deo *et al.*, 2009 and McAlpine *et al.*, 2009 for further details) (Figure 5).

At the large-scale, a global modelling study by Puma and Cook (2010) found increases in precipitation primarily downwind of the major irrigation areas. However, this study also reported that precipitation in parts of India decreased due to a weaker summer monsoon. Similar global effects were found by Guimberteau *et al.* (2011) who noted that irrigation began to significantly increase precipitation starting around 1950 over the Northern Hemisphere mid-latitudes and in the tropics.

Studies of tropical deforestation suggest a decrease in surface evapotranspiration, usually leading to a net decrease in rainfall over the area of deforestation. For example, in a modelling experiment over eastern Amazonia, Sampaio *et al.* (2007) found up to 31 (i.e. 449 mm) and 25% (491 mm) reductions in annual average ET and precipitation, respectively. However, shallow clouds occur more often over deforested areas whereas deep convective clouds favour forested areas. This feature is most evident over the Amazon basin where there is an observed significant climatic shift in shallow cloudiness patterns associated with deforestation during the dry season, when the thermal lifting mechanism is the dominant factor in convective development (Chagnon *et al.*, 2004; Wang *et al.*, 2009). Recently, Spracklen *et al.* (2012) reported

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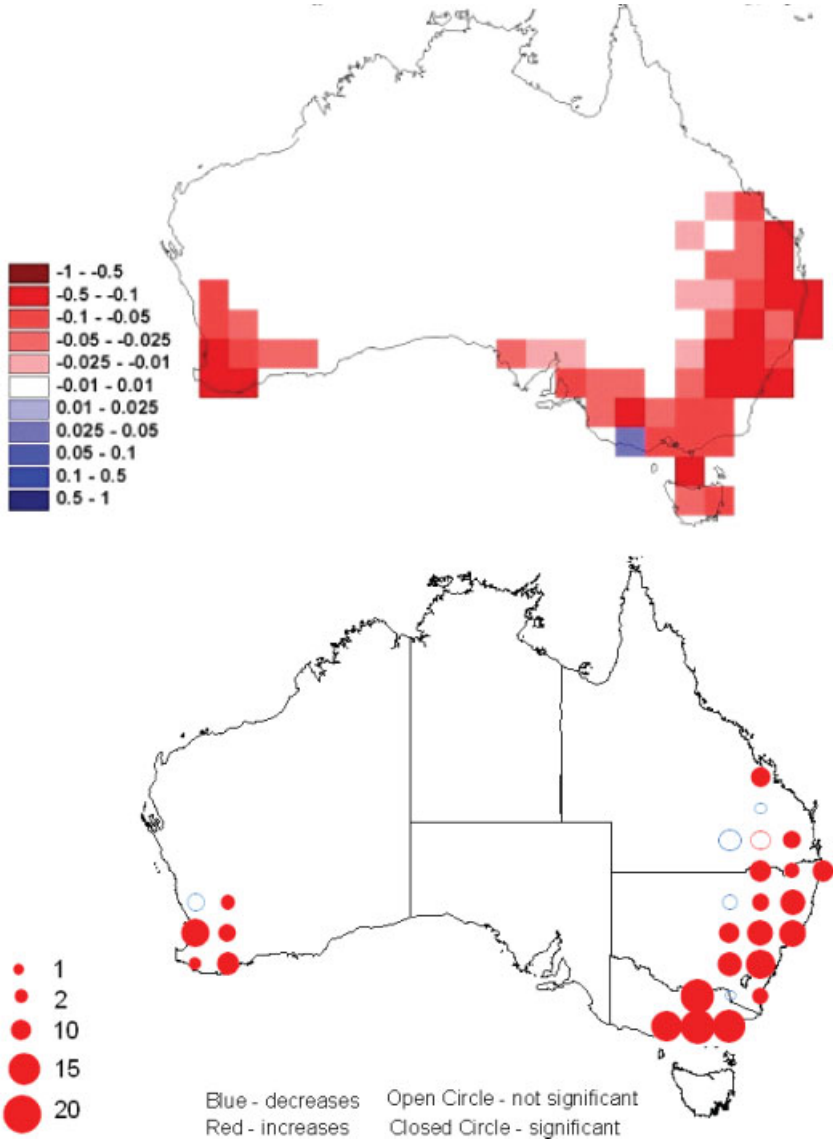

Figure 5. (a) Change in vegetation fraction as represented in the CSIRO Mark 3.5 climate model. (b) Change in the number of consecutive dry days between pre-European and modern-day conditions. Symbols: red for increase, blue for decrease, closed symbols are statistically significant, open symbols are not statistically significant at 90% confidence (Source: McAlpine *et al.*, 2007; Deo *et al.*, 2009).

on the important role of air passage over tropical forests in increasing rainfall. They found that air that has passed over extensive vegetation in the preceding few days produces at least twice as much rain as air that has passed over little vegetation.

Additional examples of tropical deforestation and its climatic impacts, including fluxes and precipitation, can be found in a series of model based research by Roy and Avissar (2002), Snyder *et al.* (2004), Pongratz *et al.* (2006), Werth and Avissar (2002), Abiodun *et al.* (2008), Da Silva *et al.* (2008), Hasler *et al.* (2009), and Davin and Noblet-Ducoudré (2010). Note that variations in the albedo or roughness change can affect the net response to deforestation (Dirmeyer and Shukla, 1996; Sud *et al.*, 1996). Modelling studies have also found similar results for deforestation over tropical Africa (e.g. Kitoh *et al.*, 1988; Xue and Shukla, 1993; Polcher and Laval, 1994a; Sud *et al.*, 1996) and Southeast Asia (e.g. Henderson-Sellers *et al.*, 1993; Polcher and Laval, 1994b; Sen *et al.*, 2004b).

In a model with ideal topography, Dirmeyer (1994) showed that differences in vegetation characteristics affected drought occurrence in mid-latitudes. Restoring vegetation in a GCM to distributions known in Europe, Asia, and North Africa within the Roman Empire (~2000 YBP) caused summer rainfall in the model to increase in southern Europe and the Atlas Mountains, the lower Nile Valley and the Levant (Reale and Dirmeyer, 2000; Reale and Shukla, 2000). Subsequently, comparison of historical and modern-era LCC modelling studies have been carried out for Europe (e.g. Heck *et al.*, 2001), East Asia (e.g. Xue, 1996; Fu *et al.*, 2004) and Australia (e.g. Narisma *et al.*, 2003; Narisma and Pitman, 2004; McAlpine *et al.*, 2007). It was found that the model responses to LCC in the mid-latitudes are complex. This is because agriculture displaces both forest and prairie/steppes,

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with opposite effects on climate from the typical changes in albedo and roughness, making the response highly sensitive to the details of the experiment implementation (Pitman *et al.*, 2009).

In general, afforestation and reforestation scenarios show precipitation increases in modelling experiments. Beltrán-Przekurat *et al.* (2012) found, from a modelling study, that in afforested areas over southern South America absolute mean values were higher, up to 0.5 mm day^{-1} in spring and 1.0 mm day^{-1} in summer, compared to current conditions. Using a global circulation model, Xue and Shukla (1996) found a 0.8 mm day^{-1} (or 27%) increases in Sahel precipitation over afforested areas and decreases south of the region.

Using a projected afforestation scenario within a regional climate model over the United States, Jackson *et al.* (2005) noted that changes in summer precipitation were not uniform and depended on location. A general decrease in rainfall was found in afforested areas located in the northern states. Precipitation increased in a few areas such as in Florida and southern Georgia and in other areas, not directly affected by the LCC.

3.1.1. Summary

LCC leads to changes in energy fluxes and their partitioning. At the meso-scale urbanization produces the most dramatic modification in energy partitioning, dominated by sensible energy flux. It is also apparent that these changes, along with other surface biophysical properties, including roughness and albedo, have led to development of convection and or precipitation. However, it is also becoming evident that impacts of urbanization on climate vary widely and are dependent on season, latitude, relative geographical location, and ecological setting. Nonetheless, impacts of urbanization on regional precipitation still need further investigation.

Irrigated agriculture also produces notable changes in surface energy partitioning. In contrast to the impacts of urbanization, the energy flux is dominated by increased latent heating (i.e. lower Bowen ratio). Research on irrigation suggests local, regional, and continental-scale impacts on precipitation. It is also possible that large-scale adoption of irrigation could impact inter-annual precipitation patterns. However, the irrigation-precipitation relationship is complex and the related science is still emerging. Studies on large-scale afforestation also suggest increases in precipitation while tropical deforestation results in lowering of evapotranspiration and precipitation.

3.2. Changes in temperature

One of the most well-known meso-scale features associated with LCC and temperature increase is the UHI (Eliasson and Homer, 1990; Arnfield, 2003; Souch and Grimmond, 2006; Jansson *et al.*, 2007; Yow, 2007; Hidalgo *et al.*, 2008; Trusilova *et al.*, 2009; Georgakis *et al.*, 2010; McCarthy *et al.*, 2010), recognized since at least 1820 (Howard, 1820). Higher surface skin, air, and canopy temperatures relative to the surrounding rural area, typically define the UHI. It is very common that urban development occurs at the expense of existing vegetated area. This type of modification from vegetated permeable surfaces to non-permeable materials such as brick, concrete, and asphalt results in lower latent heat flux and increased sensible heat flux, and hence increased air temperature. For example, Fall *et al.* (2010) found that almost all areas in the continental United States have experienced urbanization-related warming, with values ranging from 0.103 °C (conversion from agriculture to urban) to 0.066 °C (from forest to urban). These results agree with findings from studies by Kukla *et al.* (1986), Arnfield (2003), Kalnay and Cai (2003), Zhou and Shepherd (2009), and Hale *et al.* (2006, 2008).

The UHI-related temperature changes are complex and depend on time of day and year (seasons), latitude, climate regime, circulation feedbacks, surrounding land cover, and size (e.g. Arnfield, 2003; Yow, 2007; Zhou and Shepherd, 2009; Stone *et al.*, 2010). In the mid-latitudes the UHI temperature signal is most pronounced during summer (e.g. Philandras *et al.*, 1999). However, in high-latitude areas it is best developed in the winter months where urban temperatures can be up to 6 °C higher than surrounding rural regions (Hinkel and Nelson, 2007). With negligible to no solar radiation during the winter, this high-latitude UHI is largely due to anthropogenic heat released by maintaining internal building temperatures. In mid-latitude areas, a recent study by Imhoff *et al.* (2010), noted that ecological context may influence the amplitude of the summer daytime UHI. For example, cities built in biomes dominated by mixed forest and temperate broadleaf forest observed up to 8 °C urban–rural temperature difference. These authors (Imhoff *et al.*, 2010) found that urban–rural temperature differences were largest during mid-day in summer. Moreover, urban areas that replaced forest, temperate grasslands and tropical grasslands and savannah experienced 6.5-9.0, 6.3, and 5.0 °C urban–rural temperature difference, respectively.

The UHI-related temperature gradients can be dependent on both land use and urban parameters (e.g. built-up ratio, green surface ratio, sky view factor, etc.) (Oke, 1987). A net surplus of surface energy over urban regions is explained by enhanced ground heat storage and anthropogenic heating, as well as reduced evapotranspirational cooling. Smaller values of the sky view factor below roof level result in decreased longwave radiative loss and turbulent heat transfer, and add to the UHI anomaly (Unger, 2004). Additional discussion of UHI and UHI-related aspects can be found in Satoh *et al.* (1996), Ohashi *et al.* (2009), Fujibe (2010), Murata *et al.* (2012), Aoyagi *et al.* (2012), and Sachiho *et al.* (2012).

Despite UHI's status as a well studied climatological feature, uncertainties related to various processes within UHI have remained. For example, challenges are inherent when considering the multiple-scale interactions between broader global climate change and urban environments. Efforts to mitigate urban biases in the climatological

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record (Karl *et al.*, 1988; Peterson, 2003) also overlook potential natural climate signals in urban environments.

While some alterations to the land surface, such as UHI, have increased regional temperatures, regional cooling can also occur as a consequence of LCC. In particular, it is found that LCC related to rainfed agriculture has reduced regional temperatures (e.g. Bonan, 1997; Bonan, 2001; Kalnay and Cai, 2003; McPherson *et al.*, 2004; Gameda *et al.*, 2007; Fall *et al.*, 2010; Ge, 2010; Beltrán-Przekurat *et al.*, 2012). On the basis of a large-scale modelling study, Bonan (1997) reported up to 2 °C cooling of summer temperature over the central United States and up to 1.5 g kg^{-1} increase in atmospheric moisture content in much of the United States. He suggested that lowered surface roughness and stomatal resistance, and increased albedo due to replacement of forests with modern vegetation (largely agriculture), resulted in these changes. In a follow-up study, Bonan (2001) found a temporal correlation between expansion of agriculture and lowering of the daily maximum temperature. Subsequently, McPherson *et al.* (2004) and Ge (2010) found low anomalies of observed maximum temperatures for the rainfed winter wheat growing area of Oklahoma and Kansas. McPherson *et al.* (2004) also reported higher dew point temperatures over wheat growing areas compared to the surrounding native grasslands of Oklahoma. These findings were further supported by Sandstrom *et al.* (2004) who reported an increased frequency of days experiencing extreme dew point temperature (≥22 °C) in the central United States and suggested that increased evapotranspiration from croplands (i.e. LCC) to be the primary cause of this increase.

In an observational study, Gameda *et al.* (2007) found significant reductions in mid-June to mid-July maximum air temperatures, diurnal temperature range, and solar radiation of 1.7 °C decade^{-1}, 1.1 °C decade^{-1} and 1.2 M J m^{-2} decade^{-1}, respectively, in the Canadian Prairies. They attributed these changes to the increased latent heating associated with increased area under crop cultivation that resulted in lowering of temperature. Moreover, Campra *et al.* (2008) found a 0.30 °C cooling in semi arid Almeira, Spain, associated with changing of pastureland to green house farming, and resultant strong negative radiative forcing (up to −34 W m^{-2}). In summary, it is evident that the adoption of agriculture and resultant LCC has modified the energy balance, albedo, surface roughness, and radiation balance, and led to these changes in temperature.

LCC-related temperature reductions can be further amplified by irrigation. Over recent years, a series of observation and model based studies have been conducted over the key irrigated areas of the United States (Mahmood *et al.*, 2004, 2006; Christy *et al.*, 2006; Lobell *et al.*, 2006a, 2006b; Bonfils and Lobell, 2007; Kueppers *et al.*, 2007, 2008; Lobell and Bonfils, 2008; Jin and Miller, 2011; Sorooshian *et al.*, 2011), India (Sen Roy *et al.*, 2007; Biggs *et al.*, 2008), Australia (Geerts, 2002), and globally (Guimberteau *et al.*, 2011) and have reported lowered growing season temperature in these areas. For example, an observational study by Christy *et al.* (2006) estimated a 0.26 °C cooling trend decade^{-1} in the daily maximum temperature in California during the growing season while Bonfils and Lobell (2007) reported a 3.2 °C lower daily average temperature. In subsequent modelling studies Kueppers *et al.* (2007, 2008) and Sorooshian *et al.* (2011) reported up to a 7.5 °C cooling of surface temperatures due to irrigation in California. Kueppers *et al.* (2007) noted increased latent energy flux and atmospheric humidity along with these lowered temperatures.

In an observational data-based study, Mahmood *et al.* (2006) found up to a 1.41 °C lowering of maximum temperatures during the post-1945 period over irrigated locations in Nebraska (Figure 6) (Mahmood *et al.*, 2006). Moreover, a cooling trend in long-term extreme maximum temperatures was observed for irrigated locations (Mahmood *et al.*, 2004). This is further supported by increased growing-season dew point temperatures up to 2.17 °C over irrigated areas (Mahmood *et al.*, 2008). Sen Roy *et al.* (2007) reported up to 0.34 °C lowering of growing season maximum temperatures over irrigated areas in India, with individual growing season months showing up to a 0.53 °C decrease. Long-term temperatures in both regions (Nebraska and northwestern India) also showed a negative trend.

As shown in Figure 4, irrigation allows more energy to be partitioned into latent heat than sensible heat (i.e. smaller Bowen ratio), because of increased evaporative cooling, thereby lowering near-surface temperatures (Mahmood and Hubbard, 2002; Mahmood *et al.*, 2004; Kueppers *et al.*, 2007; Sen Roy *et al.*, 2011). The higher soil moisture also lowers albedo and thus increases net radiation and evaporation rate. A recent climate modelling study (Cook *et al.*, 2011) showed that the cooling effects from irrigation exist across the globe and the magnitude of the effects may remain the same or intensify over most irrigated regions under the higher greenhouse gas scenario.

Deforestation and afforestation also impacts temperatures. The consensus parameterization of the tropical deforestation studies was that surface albedo increases and roughness length decreases (Kitoh *et al.*, 1988; Mylne and Rowntree, 1992; Sud *et al.*, 1993). Although these changes have opposite impacts on near-surface air temperature, most studies suggested a net warming (see *Garratt* (1993) for a review of early studies) (Figure 7). A more robust result was a decrease in surface evapotranspiration and significant increase in annual mean temperature (Sampaio *et al.*, 2007). These authors noted up to 4.2 °C and 25 Wm^{-2} increases in temperature and sensible heat flux, respectively. Impacts of tropical deforestation on temperature can also be found in previously noted modelling studies by Shukla *et al.* (1990), Nobre *et al.* (1991), Snyder *et al.* (2004), Pongratz *et al.* (2006), Werth and Avissar (2002), Abiodun *et al.* (2008), Da Silva *et al.* (2008), and Davin and Noblet-Ducoudré (2010). Further overview of the

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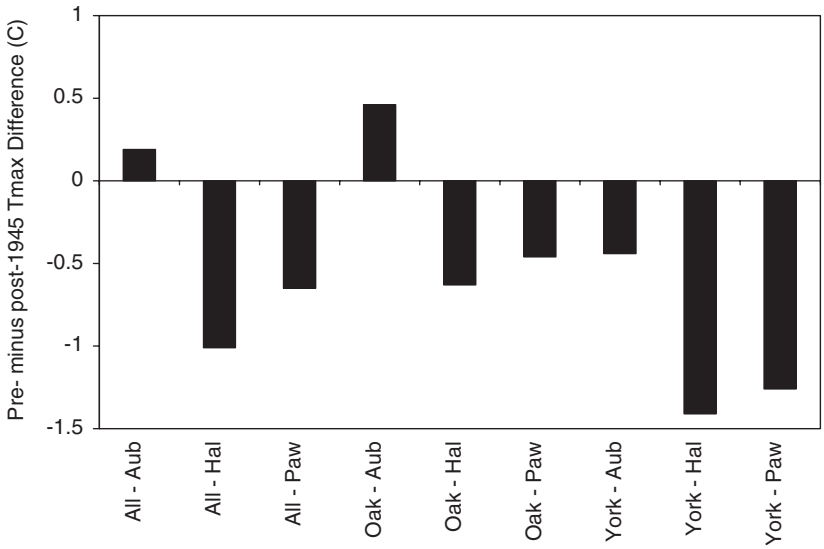

Figure 6. Cooling at irrigated locations in Nebraska, USA during the post-1945 period. Negative values show cooling. Alliance (All), Oakland (Oak), and York are irrigated locations while Halsey (Hal), Auburn (Aub), and Pawnee City (Paw) are non-irrigated locations (Source: Mahmood *et al.*, 2006).

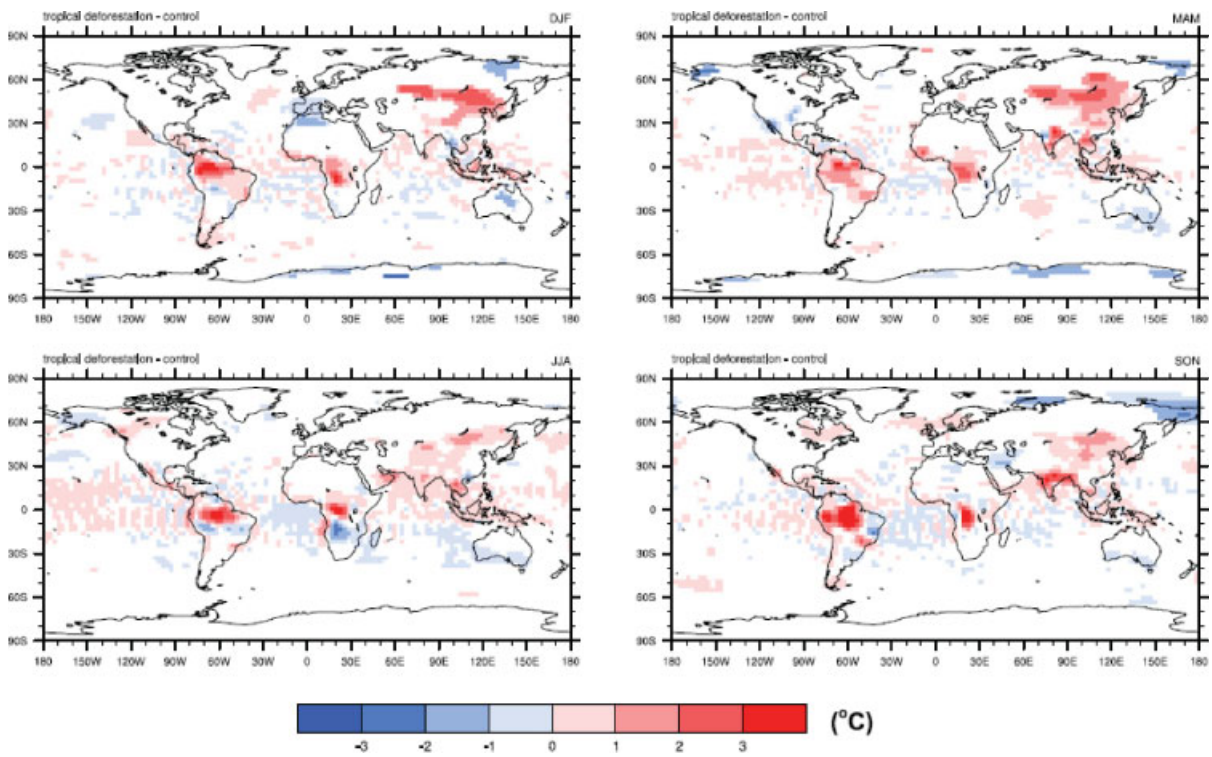

Figure 7. Tropical forest removal and its impacts on seasonal surface temperatures. These changes can also be seen over distant regions away from the location of deforestation (Source: Snyder, 2010).

links between global forests, LCC and climate change can be found in Bonan (2008b).

Despite the large area of deforestation globally, there is a recent positive trend in forest regrowth, afforestation and reforestation (Nagendra and Southworth, 2010). Little published literature is available about the effects of these processes on climate, although in general, observations and modelling studies agree that afforestation and reforestation decrease near-surface temperature due to increases in latent heat, LAI, roughness length, and rooting depth (Nosetto *et al.*, 2005; Pielke *et al.*, 2007; Strack *et al.*, 2008; Beltrán-Przekurat *et al.*, 2012).

LCC can also change the global average temperature. Various modelling studies provide ranges of estimated changes in global-scale temperature due to LCC. For example, Davin and Noblet-Ducoudré (2010) noted that global-scale deforestation may result in a 1 °C cooling of global mean temperature due to the resultant albedo

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change. This estimate exceeds the results of several previous model-based global-scale studies that reported a 0.03–0.82 °C cooling due to LCC (Betts, 2001; Claussen *et al.*, 2001; Matthews *et al.*, 2003, 2004; Brovkin *et al.*, 2006; Betts *et al.*, 2007; Davin *et al.*, 2007; Pongratz *et al.*, 2010). Findell *et al.* (2007), and Pitman *et al.* (2009) noted, respectively, a small amount of cooling in global average temperature, but disagreements among global models are a result of their variable sensitivity to LCC. On the other hand, Bounoua *et al.* (2002) found a 0.2 °C global surface warming if all forests and grasslands were replaced with croplands.

3.2.1. Summary

LCC impacts on temperature depend on the type of conversion. In particular, urbanization leads to significant warming while agriculture often leads to cooling. Agriculture-related cooling is further magnified if irrigation is introduced. On the other hand, tropical deforestation may lead to net warming, while in the mid-latitudes it may lead to cooling. Globally, deforestation may result in cooling. Also, afforestation may lead to low-latitude cooling and high-latitude warming.

Overall, the majority of observationally-based studies on agriculturally-driven LCC provided a fairly robust assessment of, and support to, the theoretical understanding of land surface-atmosphere interactions (e.g. Bonan, 2001; Kalnay and Cai, 2003; Christy *et al.*, 2006; Mahmood *et al.*, 2006; Bonfils *et al.*, 2008). Challenges associated with these studies include potential uncertainty due to shortcomings in observational data: station siting, exposure of instruments, maintenance, and lack of detailed metadata, among others. Some of the studies reviewed above used robust techniques to minimize these uncertainties. However, it is unclear whether it is possible to remove all of the biases.

A second approach to addressing LCC impacts summarized in the aforementioned studies includes observational analyses in support of hypothesized processes and mechanisms identified in regional models and established theories of land-surface atmosphere interactions (e.g. Adegoke *et al.*, 2003; Sen Roy *et al.*, 2007). Applications of biogeophysically based regional models provide an improved understanding of the causes of temperature changes and have helped refine the theories. However, many of these sensitivity studies – particularly on irrigation impacts – were based on limited atmospheric scenarios due to limited computational capabilities in the past. With improved computing resources, experiments can now be designed to include a broad range of atmospheric settings that investigate how LCC influences temperature.

3.3. Changes in atmospheric circulations and PBL

As has been described above, LCC alters the surface physical processes of albedo, net radiation, Bowen ratio, and momentum flux, and these are expressed in the climate variables of near-surface air temperature, humidity, wind speed, and soil moisture. LCC also affects the temperature, moisture and stability in the PBL that becomes evident, e.g. in convective cloud development and precipitation. In general, the top of the PBL is located closer to the ground at night over surfaces having low aerodynamic roughness (e.g. grasslands, crops) but is elevated due to daytime solar heating over rougher surfaces (forests and urban areas). Along and near steep horizontal gradients in LC types, the associated strong contrasts in albedo and convective fluxes can promote 'non-classical meso-scale circulations' (NCMCs) – so called because they resemble but are different in cause from the classical circulations of sea and land breezes – that may produce convective clouds and even precipitation in proximity to the LC boundaries (Segal *et al.*, 1988; Segal and Arritt, 1992). During daytime hours, an NCMC is characterized by vertical motion either along the boundary or displaced towards the surface having higher Bowen ratio values (i.e. strong sensible heat flux), but sinking air over the adjacent surface having lower Bowen ratio (i.e. where evaporation is greater). Accordingly, NCMCs become most evident between strongly contrasting LC types (e.g. Weaver *et al.*, 2002).

Because urban environment contains some of the above land surface characteristics, it modifies the PBL in particularly significant ways. Boundary layer changes include enhanced low-level convergence of air during the daytime ("country breeze"). The UHI is typically strongest during the nocturnal part of the diurnal cycle, but the UHI circulation is more evident during the daytime because of the urban–rural pressure gradient and vertical mixing during the daytime hours (Shreffler, 1978; Fujibe and Asai, 1984). This process explains why urban-forced convection and associated precipitation anomalies are not simply a night–early morning phenomenon. Vukovich and Dunn (1978) used a 3-D model to show that UHI intensity and PBL stability play dominant roles in UHI circulation. Huff and Vogel (1978) associated the urban circulation with increased sensible heat fluxes and surface roughness of the urban area.

Baik *et al.* (2007) and Han and Baik (2008) employed analytical and numerical models to show that PBL destabilization over the UHI leads to a region of enhanced vertical motion. They also argued that during the daytime, stability conditions were more conducive to stronger UHI-related circulations. In a model based assessment, Rozoff *et al.* (2003) found that non-linear interactions associated with friction, momentum drag, and heating could cause downwind convergence. Subsequently, Niyogi *et al.* (2006) showed that urban morphology affects both temperature and wind flow. Shem and Shepherd (2009) revealed that urban-induced convergence associated with urban circulation on the periphery of Atlanta's impervious land surface and increased sensible heat flux led to enhanced convection downwind of the city. Observational study by Shepherd *et al.* (2010a) and modelling experiments by, Carter *et al.* (2012), Lo *et al.* (2007), Yoshikado (1994), and Ohashi and Kida (2002) found similar results for Houston, Hong Kong, and cities in Japan.

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In southwest Australia, analysis of radiosonde observations (the 2005–2007 Bunny Fence Experiment – BuFex) showed increased vigour of PBL development over native vegetation, leading to higher PBL heights during both winter and summer seasons, compared to agricultural areas. On average, the noontime PBL height was higher by ~260 m over the native vegetation during summertime, while during winter it was ~189 m (Nair *et al.*, 2011) (please see introductory sections for further explanation). Energy fluxes determined from aircraft show that the enhanced PBL development was driven by heat fluxes and were consistently higher over the native vegetation areas, with peak differences of ~200 W m^{-2} and ~100 W m^{-2} observed during the summer and winter seasons, respectively. Based on modelling studies, McPherson *et al.* (2004) and Mahmood *et al.* (2011) showed changes in meso-scale wind circulation, particularly, along the LC transitions, and PBL heights due to LCC in Oklahoma and Kentucky.

A number of modelling studies found expected changes in large-scale atmospheric circulations due to LCC (e.g. Zheng and Eltahir, 1998; Chase *et al.*, 2000; Sen *et al.*, 2004b; Sen *et al.*, 2004a; Feddema *et al.*, 2005; D'Almeida *et al.*, 2007; Abiodun *et al.*, 2008; Jonko *et al.*, 2010; Snyder, 2010; Lee *et al.*, 2011). For example, in a global modelling study, Chase *et al.* (2000) demonstrated that the modifications in tropical vegetation resulted in a northward-displaced westerly jet and reduction in its maximum intensity. Similarly, over the tropical Pacific basin, the strength of the low-level easterlies was also reduced. Feddema *et al.* (2005) found that LCC could lead to weakening of the Hadley circulations and large-scale changes in the strength and timing of Asian monsoon circulations. Recently, Lee *et al.* (2011) reported changes in the large-scale Asian monsoonal circulation due to irrigation. They noted that the cooling led to significant lowering of the tropospheric geopotential height over the irrigated regions, and also modified the upper level atmospheric circulation. These changes eventually led to weakening of the upper level Asian mid-latitude jet, through a series of feedback loops.

3.3.1. Summary

On meso- and regional-scales, different land cover types (e.g. trees versus crops, rainfed versus irrigated agriculture) and land cover conversions through time (e.g. deforestation, urbanization) alter the surface and near-surface climate variables of albedo, net radiation, the Bowen ratio of convective fluxes of sensible to latent heat, and aerodynamic roughness and momentum flux. These alterations are expressed as increased spatial variability of near-surface air temperature, atmospheric humidity, and PBL characteristics of depth and stability. Accordingly, sea breeze-like meso-scale circulations within the PBL (i.e. NCMCs) can develop along and near the boundaries or transition zones of LC types (urban–rural interface, dryland-irrigated agriculture), and potentially promoting preferred areas of convective cloud formation and precipitation. Despite the progress in our understanding of these processes, the role of urbanization on meso-, and potentially regional-scale atmospheric circulation needs significant additional research over the coming years. This assertion also applies to other LCC-driven land-surface boundaries.

3.4. Teleconnections

The above assessment shows that biogeophysical impacts of LCC on local and regional-scale are significant, undeniable and discernible. Scientific research has also indicated possible teleconnections between regional LCC and climate over remote areas (e.g. Hasler *et al.*, 2009; Snyder, 2010). However, impacts of LCC on global climate and its variations are still under investigation and not fully understood (Pielke *et al.*, 2011). In addition, the question of whether LCC global impacts could be as prominent as El Niño or La Niña, or greenhouse gases, remain to be fully explored using adequate land surface representations in global models. It should be emphasized that El Niño and La Niña are large-scale coupled ocean-atmospheric oscillations, and are dynamic and cyclical over inter-annual and multi-decadal time-scales, while LCC becomes relatively 'static' after completion of the change process. LCC impacts behave more like a trend similar to greenhouse gas effects, but with great regional variability.

The lack of a persistent global climate response to LCC is partly because of the opposing and offsetting signals of local and regional impacts, and the fact that global averaging cancels and minimizes these climatic responses (e.g. Feddema *et al.*, 2005; Kvalevåg *et al.* 2010; Lawrence and Chase, 2010). For example, the modelling study of Claussen *et al.* (2001) found that low-latitude deforestation leads to regional warming and extratropical cooling via its effects on the energy and water cycles, as well as global average warming due to impacts on the carbon cycle. On the other hand, high-latitude deforestation showed a cooling effect, dominated by impacts on the surface energy balance. Modelling studies by Snyder *et al.* (2004) and Davin and Noblet-Ducoudré (2010) corroborate these results, finding cooling for both boreal and temperate deforestation, and warming as a result of tropical deforestation. However, afforestation appears to produce low-latitude cooling and high-latitude warming (Claussen *et al.*, 2001; Bala *et al.*, 2007). Despite strong regional effects, the global response to large-scale deforestation has been shown to be a slight net cooling in model simulations (Claussen *et al.*, 2001; Bala *et al.*, 2007).

Similarly, Sacks *et al.* (2009) suggested that the impacts of irrigation were significant on the regional-scales but global-scale averaged impacts were not noticeable. However, the modelling study by Puma and Cook (2010) found that regional cooling effects were already significant over southern and eastern Asia early in the 20th century, but became significant across the middle-latitude croplands after the mid-20th century. They noted that Asia and parts of North America experienced winter

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season warming during the last part of the 20th century due to increased irrigation. Puma and Cook (2010) also found a weakening of the Indian monsoon. Their results suggest a significant lowering of temperature, negative temperature trends, and increased precipitation in the tropics and the Northern Hemisphere middle-latitudes starting around the 1950s and thus, irrigation is an important component of LCC and the future global climate.

A number of modelling studies have found teleconnections between regional-scale LCC and climates of distant regions (Chase *et al.*, 2000, 2001; Gedney and Valdes, 2000; Zhao *et al.*, 2001; Werth and Avissar, 2002). Snyder *et al.* (2004) conducted a detailed global modelling study to investigate the impacts of removal of each of the six biomes. They found significant changes, not only in global average temperature and precipitation, but also in responses of these variables in regions away from the LCC. For example, it was found that tropical deforestation may result in up to a 2.5 mm d^{-1} reduction in precipitation over both oceanic and land areas distant from regions of LCC. Air temperature in remote regions also showed similar changes (Snyder *et al.*, 2004; Snyder, 2010) (Figure 7). In several subsequent modelling experiments Cui *et al.* (2006) and Hasler *et al.* (2009) found similar teleconnections between LCC and the climate of remote areas. Cui *et al.* (2006) noted that LCC in Tibet impacts East Asian atmospheric circulation and monsoon precipitation while Takata *et al.* (2009) also found significant alterations in the Asian monsoon from landscape change. The pattern and intensity of the Asian monsoon circulation impact circulations elsewhere in both the northern and southern hemispheres. Recently, Snyder (2010) further investigated impacts of LCC, reporting strong relationships between tropical deforestation and the Northern Hemisphere atmospheric circulation changes. This model based study indicates, e.g. that removal of tropical forest weakens deep convective activity and eventually impacts the northern extratropics by modifying the strength of the westerlies. Moreover, LCC in the tropics may change European storm tracks and shift the Ferrel cell northward. Amazon deforestation may modulate remote tropical ocean and climate variability (Voldoire and Royer, 2005; Schneider *et al.*, 2006; Nobre *et al.*, 2009), while the deforestation signal on weather patterns may vary in strength with the phase of El Niño (Da Silva *et al.*, 2008). Nobre *et al.* (2009), in a modelling study, found that the replacement of Amazonian tropical forest with grassland produces an ENSO-like response over the tropical Pacific Ocean, which further reduces rainfall over Amazonia. Sen *et al.* (2004b) also reported distant responses of deforestation over Indochina, including weakening of the monsoon flow over the Tibetan Plateau and eastern China.

As Findell *et al.* (2006) noted, that, it would be difficult to differentiate the extratropical response to LCC from natural climate variability. Voldoire and Royer (2005) also reported weak remote impacts of tropical deforestation. Based on a detailed model intercomparison study, Pitman *et al.* (2009) found no common remote responses to LCC. However, they noted, that this could be a function of the various model parameterizations, the use of prescribed fixed sea surface temperature and inclusion of a relatively small tropical LCC signal. It has been demonstrated that tropical LCC would produce the most significant global teleconnection response (e.g. Snyder, 2010). In addition, we note that Pitman *et al.* (2009) have used more muted global LCC (1870 *vs* current) than the other sensitivity studies (e.g. Snyder *et al.*, 2004; Snyder, 2010).

A key question is whether these drastic LCCs will occur in the future. On the basis of the past history of human modification of the land, these cannot be ruled-out. Current estimates of land cover change in the coming decades indicate a continuation from the last century, particularly in developing countries with large population growth, but also in developed countries as different land uses are implemented (e.g. production of biofuels). Moreover, such drastic changes demonstrate teleconnections between LCC of one region and the climate of distant regions. Therefore, it cannot be concluded that the climate of the distant regions will not respond to LCC of another region.

4. Incompletely understood issues

Despite progress over the recent decades, currently there remains a lack of comprehensive understanding of irrigation impacts on the regional-scale atmosphere and climate (e.g. via precipitation recycling). Bagley *et al.* (2012) show the potential impact on regional atmospheric water budgets of evaporated water from major croplands, hinting at the potential impacts of LCC and irrigation. A modelling study by Lobell *et al.* (2009) demonstrates varying regional response of temperature due to irrigation. These could be linked, in addition of model uncertainty and experimental design, to any or all of the following factors: varying levels of irrigation adoption and application, regional extent of irrigation, regional atmospheric feedback loops, and interactions between regional and large-scale circulations. In the broader sense, Keys *et al.* (2012) show the potential vulnerability of different regions to local and remote disruptions to evapotranspiration via processes like LCC or irrigation that may alter the atmospheric water supply to an area using the concept of 'precipitation-sheds'.

4.1. Deforestation

Deforestation rates vary by country and fluctuate over time in reaction to economic and political pressures. For example, the rate has increased within Latin American countries from 1.8% per year during the 1990s to over 3.2% per year in the most recent decade (FAO, 2011). These changes are dynamic, involving multiple land use pathways often leading to some form of forest regrowth, which is an important trend in Latin America (Grau and Aide, 2008). However, re-burning is often

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practiced to maintain cleared lands, especially for grazing (Fisch *et al.*, 1994). Precise knowledge of transition types of LCCs is critical in areas with dynamic agricultural frontier expansion.

Sampaio *et al.* (2007) suggested a tipping point of around 40% deforestation for the Amazon, after which climate impacts accelerate and a new equilibrium of reduced forest is reached. These effects are stronger for the transition to crops (soybean) than to pasture (Costa *et al.*, 2007). In their early modelling studies, Nobre *et al.* (1991) first hinted at the possibility of multiple equilibria in the Amazon Basin. Other factors may reinforce the forest retreat, including climate change (Salazar *et al.*, 2007) and the effect of natural fires from lightning (Hirota *et al.*, 2010). Tropical forests provide a number of biogeophysical feedbacks to the global climate system, and tropical deforestation works against mitigation of global climate change (Bonan, 2008b). However, the CO_2 fertilization effect could encourage tropical forest growth (Lapola *et al.*, 2009; Salazar and Nobre, 2010).

4.2. Benefits of reforestation?

Another important unknown issue is the capacity of reforestation to mitigate the biogeophysical climate impacts of LCC at a regional scale and which spatial configurations of vegetation might help enhance recycling of water vapour to the atmosphere through the regulation of energy fluxes, wind and surface water availability. Latitude-specific deforestation modelling experiments conducted by Bala *et al.* (2007) indicate clearly that reforestation projects in tropical regions would be beneficial in mitigating global-scale warming. However, it would be counterproductive if implemented at high latitudes and would offer only marginal benefits in temperate regions. The evidence assembled in Bala *et al.* (2007) demonstrates that deforestation in tropical and sub-tropical regions can have a significant warming and drying effect on regional climate, with teleconnections to regions remote from where the deforestation occurs. Large-scale reforestation has the potential to ameliorate regional climate changes associated with deforestation while providing other ecological services such as biodiversity, clean air and water. However, we currently do not know the extent to which such actions will modify temperature and rainfall patterns directly (McAlpine *et al.*, 2009). A related question thus is: How much vegetation is required and where should it be located? Should vegetation be configured in large blocks or in linear strips/vegetation bands, which are more amenable to integrating the production functions of landscapes (Ryan *et al.*, 2010)?

4.3. Coupling ecosystem dynamics with climate feedbacks

Most research to date on the climate impacts of LCC has focused on anthropogenic modification. However, terrestrial ecosystems and the climate system are closely coupled, with multiple interactions and feedbacks occurring across a range of scales (e.g. Chapin *et al.*, 2008). Extreme climatic events such as heatwaves, droughts, and floods can have disproportionate effects on ecosystems relative to the scale at which they occur. The timing of these events has a critical influence on their impact. For example, synergisms between heatwave and drought aggravate the negative effect on plant growth and function (De Boeck *et al.*, 2011). There is growing evidence of committed ecosystem changes due to climate change, with predicted northern expansion of boreal forests with lower net primary productivity and increased risk of forest die-off in the Amazon (Jones *et al.*, 2009). A recent example is the widespread dieback of Amazon forest due to severe drought in 2005 (Phillips *et al.*, 2009). Drought-induced forest die-off in the region may increase the biogeophysical climate impacts of deforestation, and constitutes a large uncertainty in regional climate–ecosystem interactions, and also carbon-cycle feedbacks within global climate (McAlpine *et al.*, 2010). A recent special report on extreme events by Working Groups I and II of the Intergovernmental Panel on Climate Change (IPCC) predicts that droughts will intensify in some regions such as the Mediterranean, central North America, South Africa and Australia (IPCC, 2012). This possibility highlights the importance of coupling dynamic vegetation models with regional and global climate models when investigating the biogeophysical feedbacks of LCC on the climate system.

In the context of the above discussion and the emergence of a clearer picture of regional-scale climatic response to LCC and a complex global-scale response, we propose a series of tasks. These are presented in the following subsections.

5. Recommendations

5.1. Broadening the scope of IPCC

An important lesson to be drawn from this article is the need to broaden the current global climate change agenda to recognize that climate change results from multiple forcings, and that LCC must be included in global and regional strategies to effectively mitigate climate change (Feddema *et al.*, 2005; NRC, 2005). The forthcoming IPCC's Fifth Assessment currently lacks a comprehensive evaluation of the relative impact of biogeophysical feedbacks of LCC on regional climate. Mahmood *et al.* (2010) argues that the development of suitable regional policies to adapt to the impacts of climate change, including LCC effects, must be assessed in detail as part of the IPCC Fifth Assessment, in order for them to be scientifically complete. Current risk assessments such as the special report on extreme events (IPCC, 2012) fail to account for these feedbacks. A coordinated research effort is required to address this problem, as the biogeophysical LCC forcing of climate may, in some regions, be of similar magnitude or larger than that of greenhouse gas-induced climate change (Bonan, 2008b; Avila *et al.*, 2012).

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5.2. Detection of climatic impacts of LCC with in situ measurements

A number of observational platforms should be used for better detection of climate impacts of LCC. These include *in situ* observing networks and satellites. High quality *in situ* measurements can play an important role in detecting the signals of LCC impacts. Note that the improved observational data are also necessary to fully drive the models at the resolution necessary for more accurate simulations of LCC-driven atmospheric processes. Urban micronets, such as the one in Oklahoma City (Basara *et al.*, 2009), need to be established. This type of network could be further expanded to study the role of urban morphology, shape, and form on precipitation (particularly winter precipitation), urban cloud climatologies, synergistic mechanisms and on when they are most dominant (e.g. diurnally, seasonally, and or as function of meteorological regime). High quality mesonets such as those in Oklahoma, Kentucky, Nebraska, and Delaware should also be used and expanded to other regions. Some of these mesonets could be located in regions experiencing significant LCC or where climate response could be significant (e.g. the Amazon, African tropical forests, and Boreal forests). Analysis of long-term data collected by the mesonets, and the knowledge of LCC in the vicinity, allow for linking LCC to its impacts on climate. The US Regional Climate Reference Network (USRCRN) could be used along with these mesonets to detect regional LCC-forced climate signals. The US Climate Reference Network (USCRN) can also be a useful observation platform to help in this detection.

In addition to its well-known effect on air temperature, LCC also adversely affects the measurement of precipitation. When averaged across the entire globe and for all seasons, the underestimation bias associated with precipitation measurement results in a decrease of about 8% due to the wind, 2% to wetting losses on the internal walls of the gauge and on the collector during its emptying, and 1% resulting from evaporation losses of storage gauges (Legates, 1987; Legates and Willmott, 1990). LCC can affect these biases, thereby introducing artificial trends or masking real trends in the precipitation time-series (Legates, 1995). In particular, the growth of trees and urban sprawl near a precipitation gauge can alter the wind speed and/or temperature (thereby affecting the distribution of solid versus liquid precipitation) across the gauge orifice, which systematically decreases the bias in precipitation gauge measurement. This bias decrease results in an artificial increase in precipitation that may be indistinguishable from the true precipitation amounts. In particular, checks for discontinuities in the data are not likely to identify such changes in the mean bias as they are slow, gradual and indistinguishable from true precipitation signals. As a result, when attempting to detect LCC impacts on observed precipitation, the data should be carefully evaluated for such biases.

Data from global networks such as FluxNet should also be used to detect responses of regional climate to LCC. These networks provide rich data sets that include, in addition to standard meteorological measurements, energy, water and carbon flux observations. Currently, the length of the time series for some stations within these networks is nearly two decades. As a result, they provide an excellent opportunity to assess the potential response of the regional atmosphere linked to LCC.

The large-scale adoption of irrigation in many parts of the world and its reported impacts on weather and climate, means that extensive field experiments should be undertaken to better understand the role of irrigation in the structure, evolution and modulation of the PBL at meso- and regional scales. These efforts may also be carried out in the context of severe weather impacts, and should include modelling activities to complement field campaigns and better identify the associated physical processes.

5.3. Detection of LCC Using Satellite Data

LC data collected by satellites have been explicitly used over the last several decades to monitor changes (Townshend *et al.*, 1991; DeFries and Townshend, 1994) and to establish links between LCC and the climate response (Rabin *et al.*, 1990; Carleton *et al.*, 1994; McPherson *et al.*, 2004; Jin *et al.*, 2007). Normalized Difference Vegetation Index (NDVI) and similar indices derived from optical remote sensing data have been widely used in LCC detection and are expected to continue to be used for the foreseeable future. In addition, passive microwave sensors, such as Scanning Multichannel Microwave Radiometer (SMMR) and Special Sensor Microwave Imager (SSM/I) offer valuable land cover information over the relatively long term, based on the Microwave Polarization Difference Temperature (MPDT) by which the leaf water content can be derived (e.g. Justice *et al.*, 1989). In addition to leaf water content, the MPDT is also related to soil moisture, surface roughness, and canopy structure. SMMR and SSM/I also allow for the characterization of land cover categories (Townshend *et al.*, 1989; Neale *et al.*, 1990). Advanced Microwave Scanning Radiometer-Earth Observing System (AMSR-E), a successor in technology to SMMR and SSM/I, provides vegetation water content and surface soil moisture in addition to surface temperature (Njoku, 1999; Du, 2012). AMSR-E also provides derived indices such as a microwave vegetation index (MVI) and global vegetation/roughness maps (Shi *et al.*, 2008). Vegetation conditions and soil moisture can also be estimated from the European Space Agency (ESA) L-band Soil Moisture and Ocean Salinity (SMOS) mission (Kerr *et al.*, 2010), and NASA's planned Soil Moisture Active Passive (SMAP) mission (Entekhabi *et al.*, 2010). Active microwave sensors, such as the European Remote-Sensing Satellite (ERS) scatterometer, provide land cover classification in addition to meso-scale surface geophysical parameters such as vegetation cover, surface roughness, and surface soil-moisture content. Polarimetric and interferometric (PolInSAR), multi-angle optical remote sensing, and Light Detection And Ranging (LiDAR) should also

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continue to be used for measurements of the vertical structure of vegetation (e.g. Lefsky *et al.*, 2002; Harding and Carabajal, 2005; Lefsky *et al.*, 2005; Lefsky *et al.*, 2007).

Indirectly, satellites have detected the hydrologic fingerprint of changing land use practices. Rodell *et al.* (2009) used a decade's worth of data from the Gravity Recovery And Climate Experiment (GRACE) to detect a significant depletion of groundwater over northwestern India related to greatly expanded irrigation-fed agriculture in the region. Similar GRACE assessments of groundwater depletion by agriculture have been performed for the Sacramento and San Joaquin valleys of California (Famiglietti *et al.*, 2011).

5.4. Modelling

The previously described modelling studies provide important clues as to how LCC impacts climate. They lend further support for additional research and improvement in modelling, including more realistic representations of the land surface, as highlighted in the model inter-comparison study by Pitman *et al.* (2009). These authors have noted that the sources of some limitations and uncertainties in their experiments originated from '*lack of consistency in*: *1) the implementation of LCC despite agreed maps of agricultural land*, *2) the representation of crop phenology*, *3) the parameterization of albedo*, *and 4) the representation of evapotranspiration for different land cover types*' (p. 1). Future modelling work should consider addressing these challenges for improved assessment of the climatic impacts of LCC.

Both Puma and Cook (2010) and Gordon *et al.* (2005) demonstrated the importance of irrigation as LCC and its impacts on global climate. Thus, in addition to the regional-scale, the role of irrigation in global climate should be further investigated. Large-scale global climate model-based studies should be conducted to improve understanding of physical processes and to quantify the climatic impacts of irrigation.

More accurate vegetation and management data are also needed if the goal is to continually improve the simulations focusing on climatic impacts of LCC. In the recent decades, a number of global and regional data sets for LCC (Ramankutty and Foley, 1999; Goldewijk, 2001; Waisenan and Bliss, 2002; Brown *et al.*, 2005; Pongratz *et al.*, 2008; Ramankutty *et al.*, 2008; Steyaert and Knox, 2008), fertilizer application (Potter *et al.*, 2010) and irrigation (Siebert *et al.*, 2005; Wisser *et al.*, 2010; http://www.iwmigiam.org/, accessed in July 2012) have been produced. Despite considerable progress, there is still significant room for improving the accuracy of these data sets.

Interactions among scales need to be assessed within the current modelling framework. Steyaert and Knox (2008) introduced a new analysis of LC in the eastern United States for several periods since 1650. Their data set is unique in that they present values of surface properties in the parameter format used by the modelling community, and at a reasonably fine-scale spatial resolution. They have found, when examining the large temporal changes in LC (on a fine spatial scale) for this region, that LCCs played a major role in local and regional climates and in attributing observed temperature trends. The other examples could be how meso- and regional-scale climate, modified by the LCC, interacts with large-scale climate.

The scientific community needs a more complete and coordinated investigation addressing LCC and teleconnections. In our opinion, we should build upon previous studies (e.g. Voldoire and Royer, 2004, 2005; Pitman *et al.*, 2009; Snyder, 2010) and conduct more robust and realistic multi-model global-scale simulations and analyses. This line of research should include the forcing of LCCs on global climate, in addition to their interactions with the large-scale coupled ocean–atmosphere oscillations.

Model applications are needed to examine LCC impacts on more extreme weather and climate conditions (e.g. severe thunderstorms, flash floods, floods, drought, and seasonal wetness and dryness). Detailed studies of meso- and synoptic-scale interactions of urbanization with climate are also needed. This is because urbanization represents one of the most intense and multifaceted alterations of landscape for its comparable spatial scale. Some of the challenges for these studies have been the absence of land use data to properly characterize urban physical properties and their representation in models. Recently, Oleson *et al.* (2008a, 2008b, 2010) and Jackson *et al.* (2010) have made some important progress to this end. However, additional research needs to be conducted to overcome challenges related to improved characterization and parameterization of urban surfaces so that the urban meso- and synoptic-scale interactions can be modelled realistically.

The review and synthesis comprising this article have demonstrated that LCC plays an important and spatio-temporally varied role in modifying climate. We conclude that climate change metrics of LCC should become part of any climate assessment. In addition, there are other metrics to be considered such as the magnitude of moist enthalpy changes, magnitude of the spatial redistribution of land surface latent and sensible heating (i.e. Bowen ratio), the magnitude of the spatial redistribution of precipitation and moisture convergence, and the normalized gradient of regional radiative heating changes (Mahmood *et al.*, 2010). In summary, humans are changing the face of the planet at an accelerated rate and the findings from LCC studies for all spatial scales should be incorporated into developing climate change and variability metrics that address impacts on atmospheric circulations, hydrologic cycles, and water resources.

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