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Changes in land cover and terrestrial biogeochemistry

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Abstract

1. Changes in land use and land cover due to human activities produce physical changes in land surface albedo, latent and sensible heat, and atmospheric aerosol and greenhouse gas concentrations. The combined effects of these changes have recently been estimated to account for 40% ± 16% of the human-caused global radiative forcing from 1850 to present day (*high confidence*). As a whole, the terrestrial biosphere (soil and plants) is a net “sink” for carbon (drawing down carbon from the atmosphere), and this sink has steadily increased since 1980 (*very high confidence*). Because of the uncertainty in the trajectory of land cover, the possibility of the land becoming a net carbon source cannot be excluded (*very high confidence*).

2. Climate change and induced changes in the frequency and magnitude of extreme events (e.g., droughts, floods, and heat waves) have led to large changes in plant community structure with subsequent effects on the biogeochemistry of terrestrial ecosystems. Uncertainties about how climate change will affect land cover change make it difficult to project the magnitude and sign of future climate feedbacks from land cover changes (*high confidence*).

3. Since 1901, regional averages of both the consecutive number of frost-free days and the length of the corresponding growing season have increased for the seven contiguous U.S. regions used in this assessment. However, there is important variability at smaller scales, with some locations actually showing decreases of a few days to as much as one to two weeks. Plant productivity has not increased commensurate with the increased number of frost-free days or with the longer growing season due to plant-specific temperature thresholds, plant–pollinator dependence, and seasonal limitations in water and nutrient availability (*very high confidence*). Future consequences of changes to the growing season for plant productivity are uncertain.

4. Recent studies confirm and quantify that surface temperatures are higher in urban areas than in surrounding rural areas for a number of reasons, including the concentrated release of heat from buildings, vehicles, and industry. In the United States, this urban heat island effect results in daytime temperatures 0.9°–7.2°F (0.5°–4.0°C) higher and nighttime temperatures 1.8°– 4.5°F (1.0°–2.5°C) higher in urban areas, with larger temperature differences in humid regions (primarily in the eastern United States) and in cities with larger and denser populations. The urban heat island effect will strengthen in the future as the structure, spatial extent, and population density of urban areas change and grow (*high confidence*).

10. Changes in Land Cover and Terrestrial Biogeochemistry

KEY FINDINGS

1. Changes in land use and land cover due to human activities produce physical changes in land surface albedo, latent and sensible heat, and atmospheric aerosol and greenhouse gas concentrations. The combined effects of these changes have recently been estimated to account for 40% ± 16% of the human-caused global radiative forcing from 1850 to present day (*high confidence*). As a whole, the terrestrial biosphere (soil and plants) is a net "sink" for carbon (drawing down carbon from the atmosphere), and this sink has steadily increased since 1980 (*very high confidence*). Because of the uncertainty in the trajectory of land cover, the possibility of the land becoming a net carbon source cannot be excluded (*very high confidence*).

2. Climate change and induced changes in the frequency and magnitude of extreme events (e.g., droughts, floods, and heat waves) have led to large changes in plant community structure with subsequent effects on the biogeochemistry of terrestrial ecosystems. Uncertainties about how climate change will affect land cover change make it difficult to project the magnitude and sign of future climate feedbacks from land cover changes (*high confidence*).

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10.1 Introduction

Direct changes in land use by humans are contributing to radiative forcing by altering land cover and therefore albedo, contributing to climate change (Ch. 2: Physical Drivers of Climate Change). This forcing is spatially variable in both magnitude and sign; globally averaged, it is negative (climate cooling; Figure 2.3). Climate changes, in turn, are altering the biogeochemistry of land ecosystems through extended growing seasons, increased numbers of frost-free days, altered productivity in agricultural and forested systems, longer fire seasons, and urban-induced thunderstorms (Kunkel 2016; Galloway et al. 2014). Changes in land use and land cover interact with local, regional, and global climate processes (Brown et al. 2014). The resulting ecosystem responses alter Earth's albedo, the carbon cycle, and atmospheric aerosols, constituting a mix of positive and negative feedbacks to climate change (Myhre et al. 2013; Ward et al. 2014; Figure 10.1 and Chapter 2, Section 2.6.2). Thus, changes to terrestrial ecosystems or land cover are a direct driver of climate change and they are further altered by climate change in ways that affect both ecosystem productivity and, through feedbacks, the climate itself. The following sections describe advances since the Third National Climate Assessment (NCA3) (Melillo et al. 2014) in scientific understanding of land cover and associated biogeochemistry and their impacts on the climate system.

[INSERT FIGURE 10.1 HERE]

10.2 Terrestrial Ecosystem Interactions with the Climate System

Other chapters of this report discuss changes in temperature (Ch. 6: Temperature Change), precipitation (Ch. 7: Precipitation Change), hydrology (Chapter 8: Droughts, Floods, and Wildfires), and extreme events (Ch. 9: Extreme Storms). Collectively, these processes affect the phenology, structure, productivity, and biogeochemical processes of all terrestrial ecosystems, and as such, climate change will alter land cover and ecosystem services.

10.2.1 Land Cover and Climate Forcing

Changes in land cover and land use have long been recognized as important contributors to global climate forcing (e.g., Feddema et al. 2005). Historically, studies that account for the contribution of the land cover to radiative forcing have accounted for albedo forcings only and not those from changes in land surface geophysical properties (e.g., plant transpiration, evaporation from soils, plant community structure and function) or in aerosols. Physical climate effects from land-cover or land-use change do not lend themselves directly to quantification using the traditional radiative forcing concept. However, a framework to attribute the indirect contributions of land cover to radiative forcing and the climate system—including effects on seasonal and interannual soil moisture and latent/sensible heat, evapotranspiration, biogeochemical cycle (CO_2) fluxes from soils and plants, aerosol and aerosol precursor emissions, ozone precursor emissions, and snowpack—was reported in NRC (2005). Predicting future consequences of changes in land cover on the climate system will require not only the traditional calculations of surface albedo but also surface net radiation partitioning between latent and sensible heat exchange and the effects of resulting changes in biogeochemical trace gas and aerosol fluxes. Future trajectories of land use and land cover change are uncertain and will depend on population growth, changes in agricultural yield driven by the competing demands for production of fuel (i.e., bioenergy crops), food, feed, and fiber as well as urban expansion. An example of the diversity of future land cover and land use changes is highlighted through the Representative Concentration Pathway (RCPs) and their implementation of land use/land cover to attain target goals of radiative forcing by 2100 (Hurtt et al. 2011). For example, the highest scenario, RCP8.5 (Riahi et al. 2011), features an increase of cultivated land by about 185 million hectares from 2000 to 2050 and another 120 million hectares from 2050 to 2100. In RCP6.0—the Asia Pacific Integrated Model (AIM) (Fujimori et al. 2014), urban land use increases due to population and economic growth while cropland area expands due to increasing food demand. Grassland areas decline while total forested area extent remains constant throughout the century (Hurtt et al. 2011). The Global Change Assessment Model (GCAM), RCP4.5, preserved and expanded forested areas throughout the 21st century. Agricultural land declined slightly due to this afforestation, yet food demand is met through crop yield improvements, dietary shifts, production efficiency, and international trade (Thomson et al. 2011; Hurtt et al. 2011). As with the highest scenario (RCP8.5), the lowest scenario (RCP2.6) (van Vuuren et al. 2011*a*) reallocated agricultural production from developed to developing countries, with increased bioenergy production (Hurtt et al. 2011). Continued land-use change is projected across all RCPs (2.6, 4.5, 6.0, and 8.5) and is expected to contribute between 0.9 and 1.9 W/m^2 to direct radiative forcing by 2100 (Ward et al. 2014). The RCPs demonstrate that land-use management and change combined with policy, demographic, energy technological innovations and change, and lifestyle changes all contribute to future climate (van Vuuren et al. 2011b).

Traditional calculations of radiative forcing by land-cover change yield small forcing values (Ch. 2: Physical Drivers of Climate Change) because they account only for changes in surface albedo (e.g., Myhre and Myhre 2003; Betts et al. 2007; Jones et al. 2015). Recent assessments (Myhre et al. 2013 and references therein) are beginning to calculate the relative contributions of land-use and land-cover change (LULCC) to radiative forcing in addition to albedo and/or aerosols (Ward et al. 2014). Radiative forcing data reported in this chapter are largely from observations (see Table 8.2 in Myhre et al. 2013). Ward et al. (2014) performed an independent modeling study to partition radiative forcing from natural and anthropogenic land use and land cover change and related land management activities into contributions from carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), aerosols, halocarbons, and ozone (O_3).

The more extended effects of land–atmosphere interactions from natural and anthropogenic land-use and land-cover change (LULCC; Figure 10.1) described above have recently been reviewed and estimated by atmospheric constituent (Myhre et al. 2013; Ward et al. 2014; Figure 10.2). The combined albedo and greenhouse gas radiative forcing for land-cover change is estimated to account for 40% ± 16% of the human-caused global radiative forcing from 1850 to 2010 (Ward et al. 2014; Figure 10.2). These calculations for total radiative forcing (from LULCC sources and all other sources) are consistent with Myhre et al. (2013) (2.23 W/m^2 and 2.22 W/m^2 for Ward et al. 2014 and Myhre et al. 2013, respectively). The contributions of CO_2, CH_4, N_2O and aerosols/O_3/albedo effects to total LULCC radiative forcing are about 47%, 34%, 15% and 4%, respectively, highlighting the importance of non-albedo contributions to LULCC and radiative forcing. The net radiative forcing due specifically to fire—after accounting for short-lived forcing agents (O_3 and aerosols), long-lived greenhouse gases, and land albedo change both now and in the future—is estimated to be near zero due to regrowth of forests which offsets the release of CO_2 from fire (Ward and Mahowald 2015).

10.2.2 Land Cover and Climate Feedbacks

Earth system models differ significantly in projections of terrestrial carbon uptake (Lovenduski and Bonan 2017), with large uncertainties in the effects of increasing atmospheric CO_2 concentrations (i.e., CO_2 fertilization) and nutrient downregulation on plant productivity, as well as the strength of carbon cycle feedbacks (Anav et al. 2013; Hoffman et al. 2014; Ch. 2: Physical Drivers of Climate Change). When CO_2 effects on photosynthesis and transpiration are removed from global gridded crop models, simulated response to climate across the models is comparable, suggesting that model parameterizations representing these processes remain uncertain (Rosenzweig et al. 2014).

A recent analysis shows large-scale greening in the Arctic and boreal regions of North America and browning in the boreal forests of eastern Alaska for the period 1984–2012 (Ju and Masek 2016). Satellite observations and ecosystem models suggest that biogeochemical interactions of carbon dioxide (CO_2) fertilization, nitrogen (N) deposition, and land-cover change are responsible for 25%–50% of the global greening of the Earth and 4% of Earth's browning between 1982 and 2009 (Zhu et al. 2016; Mao et al. 2016). While several studies have documented significant increases in the rate of green-up periods, the lengthening of the growing season (Section 10.3.1) also alters the timing of green-up (onset of growth) and brown-down (senescence); however, where ecosystems become depleted of water resources as a result of lengthening growing season, the actual period of productive growth can be truncated (Adams et al. 2015).

Large-scale die-off and disturbances resulting from climate change have potential effects beyond the biogeochemical and carbon cycle effects. Biogeophysical feedbacks can strengthen or reduce climate forcing. The low albedo of boreal forests provides a positive feedback, but those albedo effects are mitigated in tropical forests through evaporative cooling; for temperate forests, the evaporative effects are less clear (Bonan 2008). Changes in surface albedo, evaporation, and surface roughness can have feedbacks to local temperatures that are larger than the feedback due to the change in carbon sequestration (Jackson et al. 2008). Forest management frameworks (e.g., afforestation, deforestation, and avoided deforestation) that account for biophysical (e.g., land surface albedo and surface roughness) properties can be used as climate protection or mitigation strategies (Anderson et al. 2011).

[INSERT FIGURE 10.2 HERE]

10.2.3 Temperature Change

Interactions between temperature changes, land cover, and biogeochemistry are more complex than commonly assumed. Previous research suggested a fairly direct relationship between increasing temperatures, longer growing seasons (see Section 10.3.1), increasing plant productivity (e.g., Walsh et al. 2014), and therefore also an increase in CO_2 uptake. Without water or nutrient limitations, increased CO_2 concentrations and warm temperatures have been shown to extend the growing season, which may contribute to longer periods of plant activity and carbon uptake, but do not affect reproduction rates (Reyes-Fox et. al. 2014). However, there are other processes that offset benefits of a longer growing season, such as changes in water availability and demand for water (e.g., Georgakakos et al. 2014; Hibbard et al. 2014). For instance, increased dry conditions can lead to wildfire (e.g., Hatfield et al. 2014; Joyce et al. 2014; Ch. 8: Droughts, Floods and Wildfires) and urban temperatures can contribute to urban-induced thunderstorms in the southeastern United States (Ashley et al. 2012). Temperature benefits of early onset of plant development in a longer growing season can be offset by 1) freeze damage caused by late-season frosts; 2) limits to growth because of shortening of the photoperiod later in the season; or 3) by shorter chilling periods required for leaf unfolding by many plants (Fu et al. 2015; Gu et al. 2008). MODIS data provided insight into the coterminous U.S. 2012 drought, when a warm spring reduced the carbon cycle impact of the drought by inducing earlier carbon uptake (Wolf et al. 2016). New evidence points to longer temperature-driven growing seasons for grasslands that may facilitate earlier onset of growth, but also that senescence is typically earlier (Fridley et al. 2016). In addition to changing CO_2 uptake, higher temperatures can also enhance soil decomposition rates, thereby adding more CO_2 to the atmosphere. Similarly, temperature, as well as changes in the seasonality and intensity of precipitation, can influence nutrient and water availability, leading to both shortages and excesses, thereby influencing rates and magnitudes of decomposition (Galloway et al. 2014).

10.2.4 Water Cycle Changes

The global hydrological cycle is expected to intensify under climate change as a consequence of increased temperatures in the troposphere. The consequences of the increased water-holding capacity of a warmer atmosphere include longer and more frequent droughts and less frequent but more severe precipitation events and cyclonic activity (see Ch. 9: Extreme Storms for an in-depth discussion of extreme storms). More intense rain events and storms can lead to flooding and ecosystem disturbances, thereby altering ecosystem function and carbon cycle dynamics. For an extensive review of precipitation changes and droughts, floods, and wildfires, see Chapters 7 and 8 in this report, respectively.

From the perspective of the land biosphere, drought has strong effects on ecosystem productivity and carbon storage by reducing photosynthesis and increasing the risk of wildfire, pest infestation, and disease susceptibility. Thus, droughts of the future will affect carbon uptake and storage, leading to feedbacks to the climate system (Chapter 2, Section 2.6.2; also see Chapter 11 for Arctic/climate/wildfire feedbacks; Schlesinger et al. 2016). Reduced productivity as a result of extreme drought events can also extend for several years post-drought (i.e., drought legacy effects; Frank et al. 2015; Reichstein et al. 2013; Anderegg et al. 2015). In 2011, the most severe drought on record in Texas led to statewide regional tree mortality of 6.2%, or nearly nine times greater than the average annual mortality in this region (approximately 0.7%) (Moore et al. 2016). The net effect on carbon storage was estimated to be a redistribution of 24–30 TgC from the live to dead tree carbon pool, which is equal to 6%–7% of pre-drought live tree carbon storage in Texas state forestlands (Moore et al. 2016). Another way to think about this redistribution is that the single Texas drought event equals approximately 36% of annual global carbon losses due to deforestation and land-use change (Ciais et al. 2013). The projected increases in temperatures and in the magnitude and frequency of heavy precipitation events, changes to snowpack, and changes in the subsequent water availability for agriculture and forestry may lead to similar rates of mortality or changes in land cover. Increasing frequency and intensity of drought across northern ecosystems reduces total observed organic matter export, has led to oxidized wetland soils, and releases stored contaminants into streams after rain events (Szkokan-Emilson et al. 2017).

10.2.5 Biogeochemistry

Terrestrial biogeochemical cycles play a key role in Earth's climate system, including by affecting land–atmosphere fluxes of many aerosol precursors and greenhouse gases, including carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O). As such, changes in the terrestrial ecosphere can drive climate change. At the same time, biogeochemical cycles are sensitive to changes in climate and atmospheric composition.

Historically, increased atmospheric CO_2 concentrations have led to increased plant production (known as CO_2 fertilization) and longer-term storage of carbon in biomass and soils. Whether increased atmospheric CO_2 will continue to lead to long-term storage of carbon in terrestrial ecosystems depends on whether CO_2 fertilization simply intensifies the rate of short-term carbon cycling (for example, by stimulating respiration, root exudation, and high turnover root growth) or whether the additional carbon is used by plants to build more wood or tissues that, once senesced, decompose into long-lived soil organic matter. Under increased CO_2 concentrations, plants have been observed to optimize water use due to reduced stomatal conductance, thereby increasing water-use efficiency (Keenan et al. 2013). This change in water-use efficiency can affect plants' tolerance to stress and specifically to drought (Swann et al. 2016). Due to the complex interactions of the processes that govern terrestrial biogeochemical cycling, terrestrial ecosystem responses to increasing CO_2 levels remains one of the largest uncertainties in long-term climate feedbacks and therefore in predicting longer-term climate change (Ch. 2: Physical Drivers of Climate Change).

Nitrogen is a principal nutrient for plant growth and can limit or stimulate plant productivity (and carbon uptake), depending on availability. As a result, increased nitrogen deposition and natural nitrogen-cycle responses to climate change will influence the global carbon cycle. For example, nitrogen limitation can inhibit the CO_2 fertilization response of plants to elevated atmospheric CO_2 (e.g., Norby et al. 2005; Zaehle et al. 2010). Conversely, increased decomposition of soil organic matter in response to climate warming increases nitrogen mineralization. This shift of nitrogen from soil to vegetation can increase ecosystem carbon storage (Melillo et al. 2011; Ciais et al. 2013). While the effects of increased nitrogen deposition may counteract some nitrogen limitation on CO_2 fertilization, the importance of nitrogen in future carbon–climate interactions is not clear. Nitrogen dynamics are being integrated into the simulation of land carbon cycle modeling, but only two of the models in CMIP5 included coupled carbon–nitrogen interactions (Knutti and Sedlacek 2013).

Many factors, including climate, atmospheric CO_2 concentrations, and nitrogen deposition rates influence the structure of the plant community and therefore the amount and biochemical quality of inputs into soils (Jandl et al. 2007; McLauchlan 2006; Smith et al. 2007). For example, though CO_2 losses from soils may decrease with greater nitrogen deposition, increased emissions of other greenhouse gases, such as methane (CH_4) and nitrous oxide (N_2O), can offset the reduction in CO_2 (Liu and Greaver 2009). The dynamics of soil organic carbon under the influence of climate change is poorly understood and therefore not well represented in models. As a result, there is high uncertainty in soil carbon stocks in model simulations (Todd-Brown et al. 2013; Tian et al. 2015).

Future emissions of many aerosol precursors are expected to be affected by a number of climate-related factors, in part because of changes in aerosol and aerosol precursors from the terrestrial biosphere. For example, volatile organic compounds (VOCs) are a significant source of secondary organic aerosols, and biogenic sources of VOCs exceed emissions from the industrial and transportation sectors (Guenther et al. 2006). Isoprene is one of the most important biogenic VOCs, and isoprene emissions are strongly dependent on temperature and light, as well as other factors like plant type and leaf age (Guenther et al. 2006). Higher temperatures are expected to lead to an increase in biogenic VOC emissions. Atmospheric CO_2 concentration can also affect isoprene emissions (e.g., Rosenstiel et al. 2003). Changes in biogenic VOC emissions can impact aerosol formation and feedbacks with climate (Ch. 2: Physical Drivers of Climate Change, Section 2.6.1; Feedbacks via changes in atmospheric composition). Increased biogenic VOC emissions can also impact ozone and the atmospheric oxidizing capacity (Pyle et al. 2007). Conversely, increases in nitrogen oxide (NO_x) pollution produce tropospheric ozone (O_3), which has damaging effects on vegetation. For example, a recent study estimated yield losses for maize and soybean production of up to 5% to 10% due to increases in O_3 (McGrath et al. 2015).

10.2.6 Extreme Events and Disturbance

This section builds on the physical overview provided in earlier chapters to frame how the intersections of climate, extreme events, and disturbance affect regional land cover and biogeochemistry. In addition to overall trends in temperature (Ch. 6: Temperature Change) and precipitation (Ch. 7: Precipitation Change), changes in modes of variability such as the Pacific Decadal Oscillation (PDO) and the El Niño–Southern Oscillation (ENSO) (Ch. 5: Circulation and Variability) can contribute to drought in the United States, which leads to unanticipated changes in disturbance regimes in the terrestrial biosphere (e.g., Kam et al. 2014). Extreme climatic events can increase the susceptibility of ecosystems to invasive plants and plant pests by promoting transport of propagules into affected regions, decreasing the resistance of native communities to establishment, and by putting existing native species at a competitive disadvantage (Diez et al. 2012). For example, drought may exacerbate the rate of plant invasions by non-native species in rangelands and grasslands (Moore et al. 2016). Land-cover changes such as encroachment and invasion of non-native species can in turn lead to increased frequency of disturbance such as fire. Disturbance events alter soil moisture, which, in addition to being affected by evapotranspiration and precipitation (Ch. 8: Droughts, Floods, and Wildfires), is controlled by canopy and rooting architecture as well as soil physics. Invasive plants may be directly responsible for changes in fire regimes through increased biomass, changes in the distribution of flammable biomass, increased flammability, and altered timing of fuel drying, while others may be "fire followers" whose abundances increase as a result of shortening the fire return interval (e.g., Lambert et al. 2010). Changes in land cover resulting from alteration of fire return intervals, fire severity, and historical disturbance regimes affect long-term carbon exchange between the atmosphere and biosphere (e.g., Moore et al. 2016). Recent extensive diebacks and changes in plant cover due to drought have interacted with regional carbon cycle dynamics, including carbon release from biomass and reductions in carbon uptake from the atmosphere; however, plant regrowth may offset emissions (Vose et al. 2016). The 2011–2015 meteorological drought in California (described in Ch. 8: Droughts, Floods, and Wildfires), combined with future warming, will lead to long-term changes in land cover, leading to increased probability of climate feedbacks (e.g., drought and wildfire) and in ecosystem shifts (Diffenbaugh et al. 2015). California's recent drought has also resulted in measureable canopy water losses, posing long-term hazards to forest health and biophysical feedbacks to regional climate (Anderegg et al. 2015; Asner et al. 2016; Mann and Gleick 2015). Multiyear or severe meteorological and hydrological droughts (see Ch. 8: Droughts, Floods, and Wildfires for definitions) can also affect stream biogeochemistry and riparian ecosystems by concentrating sediments and nutrients (Vose et al. 2016).

Changes in the variability of hurricanes and winter storm events (Ch. 9: Extreme Storms) also affect the terrestrial biosphere, as shown in studies comparing historic and future (projected) extreme events in the western United States and how these translate into changes in regional water balance, fire, and streamflow. Composited across 10 global climate models (GCMs) summer (June–August) water-balance deficit in the future (2030–2059) increases compared to that under historical (1916–2006) conditions. Portions of the Southwest that have significant monsoon precipitation and some mountainous areas of the Pacific Northwest are exempt from this deficit (Littell et al. 2016). Projections for 2030–2059 suggest that extremely low flows that have historically occurred (1916–2006) in the Columbia Basin, upper Snake River, southeastern California, and southwestern Oregon are less likely to occur. Given the historical relationships between fire occurrence and drought indicators such as water-balance deficit and streamflow, climate change can be expected to have significant effects on fire occurrence and area burned (Littell et al. 2016, 2011; Elsner et al. 2010).

Climate change in the northern high latitudes is directly contributing to increased fire occurrence (Ch. 11: Arctic Changes); in the coterminous United States, climate-induced changes in fires, changes in direct human ignitions, and land-management practices all significantly contribute to wildfire trends. Wildfires in the western United States are often ignited by lightning, but management practices such as fire suppression contribute to fuels and amplify the intensity and spread of wildfire. Fires initiated from unintentional ignition, such as by campfires, or intentional human-caused ignitions are also intensified by increasingly dry and vulnerable fuels, which build up with fire suppression or human settlements (See also Ch. 8: Droughts, Floods, and Wildfires).

10.3 Climate Indicators and Agricultural and Forest Responses

Recent studies indicate a correlation between the expansion of agriculture and the global amplitude of CO_2 uptake and emissions (Zeng et al. 2014; Gray et al. 2014). Conversely, agricultural production is increasingly disrupted by climate and extreme weather events, and these effects are expected to be augmented by mid-century and beyond for most crops (Lobell and Tebaldi 2014; Challinor et al. 2014. Precipitation extremes put pressure on agricultural soil and water assets and lead to increased irrigation, shrinking aquifers, and ground subsidence.

10.3.1 Changes in the Frost-Free and Growing Seasons

The concept that longer growing seasons are increasing productivity in some agricultural and forested ecosystems was discussed in the Third National Climate Assessment (NCA3; Melillo et al. 2014). However, there are other consequences to a lengthened growing season that can offset gains in productivity. Here we discuss these emerging complexities as well as other aspects of how climate change is altering and interacting with terrestrial ecosystems. The growing season is the part of the year in which temperatures are favorable for plant growth. A basic metric by which this is measured is the frost-free period. The U.S. Department of Agriculture Natural Resources Conservation Service defines the frost-free period using a range of thresholds. They calculate the average date of the last day with temperature below 24°F (-4.4°C), 28°F (-2.2°C), and 32°F (0°C) in the spring and the average date of the first day with temperature below 24°F, 28°F, and 32°F in the fall, at various probabilities. They then define the frost-free period at three index temperatures (32°F, 28°F, and 24°F), also with a range of probabilities. A single temperature threshold (for example, temperature below 32°F) is often used when discussing growing season; however, different plant cover-types (e.g., forest, agricultural, shrub, and tundra) have different temperature thresholds for growth, and different requirements/thresholds for chilling (Zhang et al. 2011; Hatfield et al. 2014). For the purposes of this report, we use the metric with a 32°F (0°C) threshold to define the change in the number of "frost-free" days, and a temperature threshold of 41°F (5°C) as a first-order measure of how the growing season length has changed over the observational record (Zhang et al. 2011).

The NCA3 reported an increase in the growing season length of as much as several weeks as a result of higher temperatures occurring earlier and later in the year (e.g., Walsh et al. 2014; Hatfield et al. 2014; Joyce et al. 2014). NCA3 used a threshold of 32°F (0°C) (i.e., the frost-free period) to define the growing season. An update to this finding is presented in Figures 10.3 and 10.4, which show changes in the frost-free period and growing season, respectively, as defined above. Overall, the length of the frost-free period has increased in the contiguous United States during the past century (Figure 10.3). However, growing season changes are more variable: growing season length increased until the late 1930s, declined slightly until the early 1970s, increased again until about 1990, and remained quasi-stable thereafter (Figure 10.4). This contrasts somewhat with changes in the length of the frost-free period presented in NCA3, which showed a continuing increase after 1980. This difference is attributable to the temperature thresholds used in each indicator to define the start and end of these periods. Specifically, there are now more frost-free days (32°F threshold) in winter than the growing season (41°F threshold).

The lengthening of the growing season has been somewhat greater in the northern and western United States, which experienced increases of 1–2 weeks in many locations. In contrast, some areas in the Midwest, Southern Great Plains, and the Southeast had decreases of a week or more between the periods 1986–2015 and 1901–1960 (Kunkel 2016). These differences reflect the more general pattern of warming and cooling nationwide (Ch. 6: Temperature Changes). Observations and models have verified that the growing season has generally increased plant productivity over most of the United States (Mao et al. 2016).

Consistent with increases in growing season length and the coldest temperature of the year, plant hardiness zones have shifted northward in many areas (Daly et al. 2012). The widespread increase in temperature has also impacted the distribution of other climate zones in parts of the United States. For instance, there have been moderate changes in the range of the temperate and continental climate zones of the eastern United States since 1950 (Chan and Wu 2015) as well as changes in the coverage of some extreme climate zones in the western United States. In particular, the spatial extent of the "alpine tundra" zone has decreased in high-elevation areas (Diaz and Eischeid 2007), while the extent of the "hot arid" zone has increased in the Southwest (Grundstein 2008).

The period over which plants are actually productive, that is, their true growing season, is a function of multiple climate factors, including air temperature, number of frost-free days, and rainfall, as well as biophysical factors, including soil physics, daylight hours, and the biogeochemistry of ecosystems (EPA 2016). Temperature-induced changes in plant phenology, like flowering or spring leaf onset, could result in a timing mismatch (phenological asynchrony) with pollinator activity, affecting seasonal plant growth and reproduction and pollinator survival (Yang and Rudolf 2010; Rafferty and Ives 2011; Kudo and Ida 2013; Forrest 2015). Further, while growing season length is generally referred to in the context of agricultural productivity, the factors that govern which plant types will grow in a given location are common to all plants whether they are in agricultural, natural, or managed landscapes. Changes in both the length and the seasonality of the growing season, in concert with local environmental conditions, can have multiple effects on agricultural productivity and land cover.

In the context of agriculture, a longer growing season could allow for the diversification of cropping systems or allow multiple harvests within a growing season. For example, shifts in cold hardiness zones across the contiguous United States suggest widespread expansion of thermally suitable areas for the cultivation of cold-intolerant perennial crops (Parker and Abatzoglou 2016) as well as for biological invasion of non-native plants and plant pests (Hellmann et al. 2008). However, changes in available water, conversion from dry to irrigated farming, and changes in sensible and latent heat exchange associated with these shifts need to be considered. Increasingly dry conditions under a longer growing season can alter terrestrial organic matter export and catalyze oxidation of wetland soils, releasing stored contaminants (for example, copper and nickel) into streamflow after rainfall (Szkokan-Emilson et al. 2017). Similarly, a longer growing season, particularly in years where water is limited, is not due to warming alone, but is exacerbated by higher atmospheric CO_2 concentrations that extend the active period of growth by plants (Reyes-Fox et al. 2014). Longer growing seasons can also limit the types of crops that can be grown, encourage invasive species encroachment or weed growth, or increase demand for irrigation, possibly beyond the limits of water availability. They could also disrupt the function and structure of a region's ecosystems and could, for example, alter the range and types of animal species in the area.

A longer and temporally shifted growing season also affects the role of terrestrial ecosystems in the carbon cycle. Neither seasonality of growing season (spring and summer) nor carbon, water, and energy fluxes should be interpreted separately when analyzing the impacts of climate extremes such as drought (Sippel et al. 2016; Wolf et al. 2016; Ch. 8: Droughts, Floods, and Wildfires). Observations and data-driven model studies suggest that losses in net terrestrial carbon uptake during record warm springs followed by severely hot and dry summers can be largely offset by carbon gains in record-exceeding warmth and early arrival of spring (Wolf et al. 2016). Depending on soil physics and land cover, a cool spring, however, can deplete soil water resources less rapidly, making the subsequent impacts of precipitation deficits less severe (Sippel et al. 2016). Depletion of soil moisture through early plant activity in a warm spring can potentially amplify summer heating, a typical lagged direct effect of an extremely warm spring (Frank et al. 2015). Ecosystem responses to the phenological changes of timing and extent of growing season and subsequent biophysical feedbacks are therefore strongly dependent on the timing of climate extremes (Sippel et al. 2016; Ch. 8: Droughts, Floods, and Wildfires; Ch. 9: Extreme Storms).

The global Coupled Model Intercomparison Project Phase 5 (CMIP5) analyses did not explicitly explore future changes to the growing season length. Many of the projected changes in North American climate are generally consistent across CMIP5 models, but there is substantial inter-model disagreement in projections of some metrics important to productivity in biophysical systems, including the sign of regional precipitation changes and extreme heat events across the northern United States (Maloney et al. 2014).

[INSERT FIGURES 10.3 AND 10.4 HERE]

10.3.2 Water Availability and Drought

Drought is generally parameterized in most agricultural models as limited water availability and is an integrated response of both meteorological and agricultural drought, as described in Chapter 8: Droughts, Floods, and Wildfires. However, physiological as well as biophysical processes that influence land cover and biogeochemistry interact with drought through stomatal closure induced by elevated atmospheric CO_2 levels (Keenan et al. 2013; Swann et al. 2016). This has direct impacts on plant transpiration, atmospheric latent heat fluxes, and soil moisture, thereby influencing local and regional climate. Drought is often offset by management through groundwater withdrawals, with increasing pressure on these resources to maintain plant productivity. This results in indirect climate effects by altering land surface exchange of water and energy with the atmosphere (Marston et al. 2015).

10.3.3 Forestry Considerations

Climate change and land-cover change in forested areas interact in many ways, such as through changes in mortality rates driven by changes in the frequency and magnitude of fire, insect infestations, and disease. In addition to the direct economic benefits of forestry, unquantified societal benefits include ecosystem services, like protection of watersheds and wildlife habitat, and recreation and human health value. United States forests and related wood products also absorb and store the equivalent of 16% of all CO_2 emitted by fossil fuel burning in the United States each year (Melillo et al., 2014). Climate change is expected to reduce the carbon sink strength of forests overall.

Effective management of forests offers the opportunity to reduce future climate change—for example, as given in proposals for Reduced Emissions from Deforestation and forest Degradation (REDD+; https://www.forestcarbonpartnership.org/what-redd) in developing countries and tropical ecosystems (see Ch. 14: Mitigation)—by capturing and storing carbon in forest ecosystems and long-term wood products (Lippke et al. 2011). Afforestation in the United States has the potential to capture and store 225 million tons of additional carbon per year from 2010 to 2110 (EPA 2005; King et al. 2006). However, the projected maturation of United States forests (Wear and Coulston 2015) and land-cover change, driven in particular by the expansion of urban and suburban areas along with projected increased demands for food and bioenergy, threaten the extent of forests and their carbon storage potential (McKinley et al. 2011).

Changes in growing season length, combined with drought and accompanying wildfire are reshaping California's mountain ecosystems. The California drought led to the lowest snowpack in 500 years, the largest wildfires in post-settlement history, greater than 23% stress mortality in Sierra mid-elevation forests, and associated post-fire erosion (Asner et al., 2016). It is anticipated that slow recovery, possibly to different ecosystem types, with numerous shifts to species' ranges will result in long-term changes to land surface biophysical as well as ecosystem structure and function in this region (Asner et al. 2016; http://www.fire.ca.gov/treetaskforce/).

While changes in forest stocks, composition, and the ultimate use of forest products can influence net emissions and climate, the future net changes in forest stocks remain uncertain (Bonan 2008; Pan et al. 2011; Hurtt et al. 2011; Hansen et al. 2013; Williams et al. 2013). This uncertainty is due to a combination of uncertainties in future population size, population distribution and subsequent land-use change, harvest trends, wildfire management practices (for example, large-scale thinning of forests), and the impact of maturing U.S. forests.

10.4 Urban Environments and Climate Change

Urban areas exhibit several characteristics that affect land-surface and geophysical attributes, including building infrastructure (rougher, more uneven surfaces compared to rural or natural systems), increased emissions and concentrations of aerosols and other greenhouse gasses, and increased anthropogenic heat sources (Grimmond et al. 2016; Mitra and Shepherd 2016). The understanding that urban areas modify their surrounding environment has been accepted for over a century, but the mechanisms through which this occurs have only begun to be understood and analyzed for more than 40 years (Landsberg 1970; Mitra and Shepherd 2016). Prior to the 1970s, the majority of urban climate research was observational and descriptive (Mills 2007), but since that time, more importance has been given to physical dynamics that are a function of land surface (for example, built environment and change to surface roughness); hydrologic, aerosol, and other greenhouse gas emissions; thermal properties of the built environment; and heat generated from human activities (Seto et al. 2016 and references therein).

There is now strong evidence that urban environments modify local microclimates, with implications for regional and global climate change (Mills 2007; Mitra and Shepherd 2016). Urban systems affect various climate attributes, including temperature, rainfall intensity and frequency, winter precipitation (snowfall), and flooding. New observational capabilities—including NASA's dual polarimetric radar, advanced satellite remote sensing (for example, the Global Precipitation Measurement Mission-GPM), and regionalized, coupled land–surface–atmospheric modeling systems for urban systems—are now available to evaluate aspects of daytime and nighttime temperature fluctuations; urban precipitation; contribution of aerosols; how the urban built environment impacts the seasonality and type of precipitation (rain or snow) as well as the amount and distribution of precipitation; and the significance of the extent of urban metropolitan areas (Shepherd 2013; Seto and Shepherd 2009; Grimmond et al. 2016; Mitra and Shepherd 2016).

The urban heat island (UHI) is characterized by increased surface and canopy temperatures as a result of heat-retaining asphalt and concrete, a lack of vegetation, and anthropogenic generation of heat and greenhouse gasses (Shepherd 2013). The heat gain due to the storage capacity of urban built structures, reductions in local evapotranspiration, and anthropogenically generated heat alter the spatio-temporal pattern of temperature and leads to the UHI phenomenon. The UHI physical processes that affect the climate system include generation of heat storage in buildings during the day, nighttime release of latent heat storage by buildings, and sensible heat generated by human activities, include heating of buildings, air conditioning, and traffic (Hidalgo et al. 2008).

The strength of the effect is correlated with the spatial extent and population density of urban areas; however, because of varying definitions of urban vs. non-urban, impervious surface area is a more objective metric for estimating the extent and intensity of urbanization (Imhoff et al. 2010). Based on land surface temperature measurements, on average, the UHI effect increases urban temperature by 5.2°F (2.9°C), but it has been measured at 14.4°F (8°C) in cities built in areas dominated by temperate forests (Imhoff et al. 2010). In arid regions, however, urban areas can be more than 3.6°F (2°C) cooler than surrounding shrublands (Bounoua et al. 2015). Similarly, urban settings lose up to 12% of precipitation through impervious surface runoff, versus just over 3% loss to runoff in vegetated regions. Carbon losses from the biosphere to the atmosphere through urbanization account for almost 2% of the continental terrestrial biosphere total, a significant proportion given that urban areas only account for around 1% of land in the United States (Bounoua et al. 2015). Similarly, statistical analyses of the relationship between climate and urban land use suggest an empirical relationship between the patterns of urbanization and precipitation deficits during the dry season. Causal factors for this reduction may include changes to runoff (for example, impervious-surface versus natural-surface hydrology) that extend beyond the urban heat island effect and energy-related aerosol emissions (Kaufmann et al. 2007).

The urban heat island effect is more significant during the night and during winter than during the day, and it is affected by the shape, size, and geometry of buildings in urban centers as well as by infrastructure along gradients from urban to rural settlements (Seto and Shepherd 2009; Grimmond et al. 2016; Seto et al. 2016). Recent research points to mounting evidence that urbanization also affects cycling of water, carbon, aerosols, and nitrogen in the climate system (Seto and Shepherd 2009).

Coordinated modeling and observational studies have revealed other mechanisms by which the physical properties of urban areas can influence local weather and climate. It has been suggested that urban-induced wind convergence can determine storm initiation; aerosol concentrations and composition then influence the amount of cloud water and ice present in the clouds. Aerosols can also influence updraft and downdraft intensities, their life span, and surface precipitation totals (Shepherd 2013). A pair of studies investigated rainfall efficiency in sea-breeze thunderstorms and found that integrated moisture convergence in urban areas influenced storm initiation and mid-level moisture, thereby affecting precipitation dynamics (Shepherd et al. 2001; van Den Heever and Cotton 2007).

According to the World Bank, over 81% of the United States population currently resides in urban settings (World Bank 2017). Climate mitigation efforts to offset UHI are often stalled by the lack of quantitative data and understanding of the specific factors of urban systems that contribute to UHI. A recent study set out to quantitatively determine contributors to the intensity of UHI across North America (Zhao et al. 2014). The study found that population strongly influenced nighttime UHI, but that daytime UHI varied spatially following precipitation gradients. The model applied in this study indicated that the spatial variation in the UHI signal was controlled most strongly by impacts on the atmospheric convection efficiency. Because of the impracticality of managing convection efficiency, results from Zhao et al. (2014) support albedo management as an efficient strategy to mitigate UHI on a large scale.

TRACEABLE ACCOUNTS

Key Finding 1

Changes in land use and land cover due to human activities produce physical changes in land surface albedo, latent and sensible heat, and atmospheric aerosol and greenhouse gas concentrations. The combined effects of these changes have recently been estimated to account for 40% ± 16% of the human-caused global radiative forcing from 1850 to present day (*high confidence*). As a whole, the terrestrial biosphere (soil and plants) is a net "sink" for carbon (drawing down carbon from the atmosphere), and this sink has steadily increased since 1980 (*very high confidence*). Because of the uncertainty in the trajectory of land cover, the possibility of the land becoming a net carbon source cannot be excluded (*very high confidence*).

Description of evidence base

Traditional methods that estimate albedo changes for calculating radiative forcing due to land-use change were identified by NRC (2005). That report recommended that indirect contributions of land-cover change to climate-relevant variables, such as soil moisture, greenhouse gas (e.g., CO_2 and water vapor) sources and sinks, snow cover, and aerosol and aerosol and ozone precursor emissions also be considered. Several studies have documented physical land surface processes such as albedo, surface roughness, sensible and latent heat exchange, and land-use and land-cover change that interact with regional atmospheric processes (e.g., Marotz et al. 1975; Barnston and Schickendanz 1984; Alpert and Mandel 1986; Pielke and Zeng 1989; Feddema et al. 2005; Pielke et al. 2007), however, traditional calculations of radiative forcing by land-cover change in global climate model simulations yield small forcing values (Ch. 2: Physical Drivers of Climate Change) because they account only for changes in surface albedo (e.g., Myhre and Myhre 2003; Betts et al. 2007; Jones et al. 2015).

Recent studies that account for the physical as well as biogeochemical changes in land cover and land use radiative forcing estimated that these drivers contribute 40% of present radiative forcing due to land-use/land-cover change (0.9 W/m^2) (Ward et al. 2014; Myhre et al. 2013). These studies utilized AR5 and follow-on model simulations to estimate changes in land-cover and land-use climate forcing and feedbacks for the greenhouse gases—carbon dioxide, methane, and nitrous oxide—that contribute to total anthropogenic radiative forcing from land-use and land-cover change (Myhre et al., 2013; Ward et al., 2014). This research is grounded in long-term observations that have been documented for over 40 years and recently implemented into global Earth system models (Myhre et al. 2013; Anav et al 2013). For example, IPCC, 2013: Summary for Policymakers states: "From 1750 to 2011, CO_2 emissions from fossil fuel combustion and cement production have released 375 [345 to 405] GtC to the atmosphere, while deforestation and other land-use changes are estimated to have released 180 [100 to 260] GtC. This results in cumulative anthropogenic emissions of 555 [470 to 640] GtC." (IPCC 2013). IPCC 2013, Working Group 1, Chapter 14 states for North America: "In summary, it is very likely that by mid-century the anthropogenic warming signal will be large compared to natural variability such as that stemming from the NAO, ENSO, PNA, PDO, and the NAMS in all North America regions throughout the year" (Christensen et al. 2013).

Major uncertainties

Uncertainty exists in the future land-cover and land-use change as well as uncertainties in regional calculations of land-cover change and associated radiative forcing. The role of the land as a current sink has *very high confidence;* however, future strength of the land sink is uncertain (Wear and Coulston 2015; McKinley et al. 2011). The existing impact of land systems on climate forcing has *high confidence* (Myhre et al. 2013). Based on current RCP scenarios for future radiative forcing targets ranging from 2.6 to 8.5 W/m^2, the future forcing has lower confidence because it is difficult to estimate changes in land cover and land use into the future (van Vuuren et al. 2011*b*). Compared to 2000, the RCP8.5 CO_2-eq. emissions more than double by 2050 and increase by three by 2100 (Riahi et al. 2011). About one quarter of this increase is due to increasing use of fertilizers and intensification of agricultural production, giving rise to the primary source of N_2O emissions. In addition, increases in livestock population, rice production, and enteric fermentation processes increase CH_4 emissions (Riahi et al. 2011). Therefore, if existing trends in land-use and land-cover change continue, the contribution of land cover to forcing will increase with *high confidence*. Overall, future scenarios from the RCPs suggest that land-cover change based on policy, bioenergy, and food demands could lead to significantly different distribution of land cover types (forest, agriculture, urban) by 2100 (Hurtt et al. 2011; Riahi et al. 2011; Thomson et al. 2011; van Vuuren et al. 2011*a,b*; Fujimori et al. 2014).

Summary sentence or paragraph that integrates the above information

The key finding is based on basic physics and biophysical models that have been well established for decades with regards to the contribution of land albedo to radiative forcing (NRC 2005). Recent assessments specifically address additional biogeochemical contributions of land-cover and land-use change to radiative forcing (NRC 2005; Myhre et al. 2013). The role of current sink strength of the land is also uncertain (Wear and Coulston 2015; McKinley et al. 2011). The future distribution of land cover and contributions to total radiative forcing are uncertain and depend on policy, energy demand and food consumption, dietary demands (van Vuuren et al. 2011b).

Key Finding 2

Climate change and induced changes in the frequency and magnitude of extreme events (e.g., droughts, floods, and heat waves) have led to large changes in plant community structure with subsequent effects on the biogeochemistry of terrestrial ecosystems. Uncertainties about how climate change will affect land cover change make it difficult to project the magnitude and sign of future climate feedbacks from land cover changes (*high confidence*).

Description of evidence base

From the perspective of the land biosphere, drought has strong effects on ecosystem productivity and carbon storage by reducing microbial activity and photosynthesis and by increasing the risk of wildfire, pest infestation, and disease susceptibility. Thus, future droughts will affect carbon uptake and storage, leading to feedbacks to the climate system (Schlesinger et al. 2016). Reduced productivity as a result of extreme drought events can also extend for several years post-drought (i.e., drought legacy effects; Frank et al. 2015; Reichstein et al. 2013; Anderegg et al. 2015). Under increased CO_2 concentrations, plants have been observed to optimize water use due to reduced stomatal conductance, thereby increasing water-use efficiency (Keenan et al. 2013). This change in water-use efficiency can affect plants' tolerance to stress and specifically to drought (Swann et al. 2016).

Recent severe droughts in the western United States (Texas and California) have led to significant mortality and carbon cycle dynamics. (Moore et al., 2016, Asner et al., 2016; http://www.fire.ca.gov/treetaskforce/). Carbon redistribution through mortality in the Texas drought was around 36% of global carbon losses due to deforestation and land use change (Ciais et al. 2013).

Major uncertainties

Major uncertainties include how future land-use/land-cover changes will occur as a result of policy and/or mitigation strategies in addition to climate change. Ecosystem responses to phenological changes are strongly dependent on the timing of climate extremes (Sippel et al. 2016). Due to the complex interactions of the processes that govern terrestrial biogeochemical cycling, terrestrial ecosystem response to increasing CO_2 levels remains one of the largest uncertainties in long-term climate feedbacks and therefore in predicting longer-term climate change effects on ecosystems (e.g., Swann et al. 2016).

Summary sentence or paragraph that integrates the above information

The timing, frequency, magnitude, and extent of climate extremes strongly influence plant community structure and function, with subsequent effects on terrestrial biogeochemistry and feedbacks to the climate system. Future interactions between land cover and the climate system are uncertain and depend on human land-use decisions, the evolution of the climate system, and the timing, frequency, magnitude, and extent of climate extremes

Key Finding 3

Since 1901, regional averages of both the consecutive number of frost-free days and the length of the corresponding growing season have increased for the seven contiguous U.S. regions used in this assessment. However, there is important variability at smaller scales, with some locations actually showing decreases of a few days to as much as one to two weeks. Plant productivity has not increased commensurate with the increased number of frost-free days or with the longer growing season due to plant-specific temperature thresholds, plant–pollinator dependence, and seasonal limitations in water and nutrient availability (*very high confidence*). Future consequences of changes to the growing season for plant productivity are uncertain.

Description of evidence base

Data on the lengthening and regional variability of growing season since 1901 were updated by Kunkel (2016). Many of these differences reflect the more general pattern of warming and cooling nationwide (Ch. 6: Temperature Changes). Without nutrient limitations, increased CO_2 concentrations and warm temperatures have been shown to extend the growing season, which may contribute to longer periods of plant activity and carbon uptake, but do not affect reproduction rates (Reyes-Fox et al. 2014). However, other confounding variables that coincide with climate change (for example, drought, increased ozone, and reduced photosynthesis due to increased or extreme heat) can offset increased growth associated with longer growing seasons (Adams et al. 2015) as well as changes in water availability and demand for water (e.g., Georgakakos et al. 2014; Hibbard et al. 2014). Increased dry conditions can lead to wildfire (e.g., Hatfield et al. 2014; Joyce et al. 2014; Ch. 8: Droughts, Floods and Wildfires) and urban temperatures can contribute to urban-induced thunderstorms in the southeastern United States (Ashley et al. 2012). Temperature benefits of early onset of plant development in a longer growing season can be offset by 1) freeze damage caused by late-season frosts; 2) limits to growth because of shortening of the photoperiod later in the season; or 3) by shorter chilling periods required for leaf unfolding by many plants (Fu et al. 2015; Gu et al. 2008).

Major uncertainties

Uncertainties exist in future response of the climate system to anthropogenic forcings (land use/land cover as well as fossil fuel emissions) and associated feedbacks among variables such as temperature and precipitation interactions with carbon and nitrogen cycles as well as land-cover change that impact the length of the growing season (Reyes-Fox et al. 2014, Hatfield et al. 2014, Adams et al. 2015; Ch. 6: Temperature Changes and Ch. 8: Droughts, Floods and Wildfires).

Summary sentence or paragraph that integrates the above information

Changes in growing season length and interactions with climate, biogeochemistry and land cover were covered in 12 chapters of NCA3 (Melillo et al. 2014), but with sparse assessment of how changes in the growing season might offset plant productivity and subsequent feedbacks to the climate system. This key finding provides an assessment of the current state of the complex nature of the growing season.

Key Finding 4

Recent studies confirm and quantify higher surface temperatures in urban areas than in surrounding rural areas, for a number of reasons including the concentrated release of heat from buildings, vehicles, and industry. In the United States, this urban heat island effect results in daytime temperatures 0.9°–7.2°F (0.5°–4.0°C) higher and nighttime temperatures 1.8°– 4.5°F (1.0°–2.5°C) higher in urban areas, with larger temperature differences in humid regions (primarily in the eastern United States) and in cities with larger and denser populations. The urban heat island effect will strengthen in the future as the structure, spatial extent, and population density of urban areas change and grow (*high confidence*).

Description of evidence base

Urban interactions with the climate system have been investigated for more than 40 years (Landsberg 1970; Mitra and Shepherd 2016). The heat gain due to the storage capacity of urban built structures, reduction in local evapotranspiration, and anthropogenically generated heat alter the spatio-temporal pattern of temperature and leads to the well-known urban heat island (UHI) phenomenon (Seto and Shepherd 2009; Grimmond et al. 2016; Seto et al. 2016). The urban heat island (UHI) effect is correlated with the extent of impervious surfaces, which alter albedo or the saturation of radiation (Imhoff et al. 2010). The urban-rural difference that defines the UHI is greatest for cities built in temperate forest ecosystems (Imhoff et al. 2010). The average temperature increase is 2.9°C, except for urban areas in biomes with arid and semiarid climates (Imhoff et al. 2010; Bounoua et al. 2015).

Major uncertainties

The largest uncertainties about urban forcings or feedbacks to the climate system are how urban settlements will evolve and how energy consumption and efficiencies, and their interactions with land cover and water, may change from present times (Riahi et al. 2011; van Vuuren et al. 2011*b*; Hibbard et al. 2014; Seto et al. 2016)

Summary sentence or paragraph that integrates the above information

Key Finding 4 is based on simulated and satellite land surface measurements analyzed by Imhoff et al. (2010). Bounoua et al. (2015), Shepherd (2013), Seto and Shepherd (2009), Grimmond et al. (2016), Seto et al. (2016) provide specific references with regards to how building materials and spatio-temporal patterns of urban settlements influence radiative forcing and feedbacks of urban areas to the climate system.

FIGURES

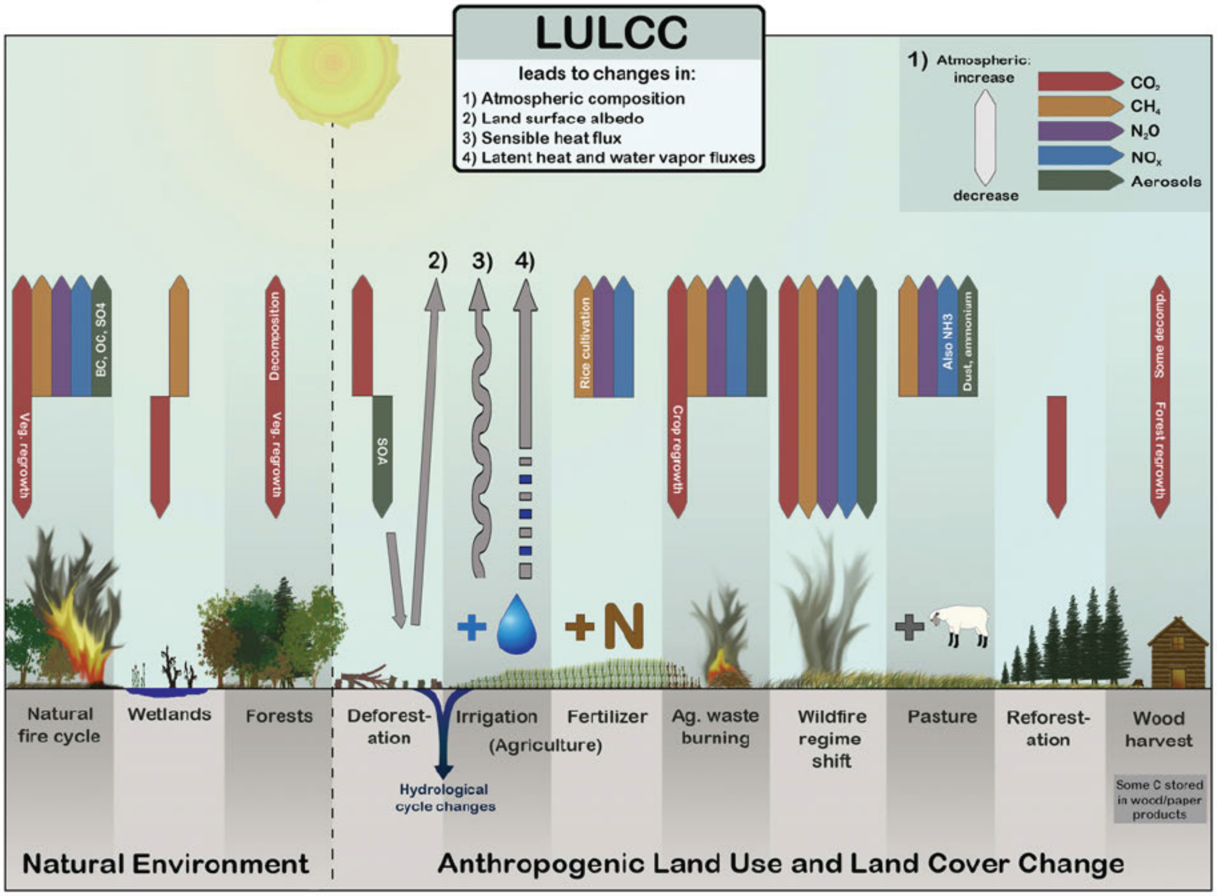


Figure 10.1. This graphical representation summarizes land–atmosphere interactions from natural and anthropogenic land-use and land-cover change (LULCC) contributions to radiative forcing. Emissions and sequestration of carbon and fluxes of nitrogen oxides, aerosols, and water shown here were used to calculate net radiative forcing from LULCC. (Figure source: Ward et al. 2014).

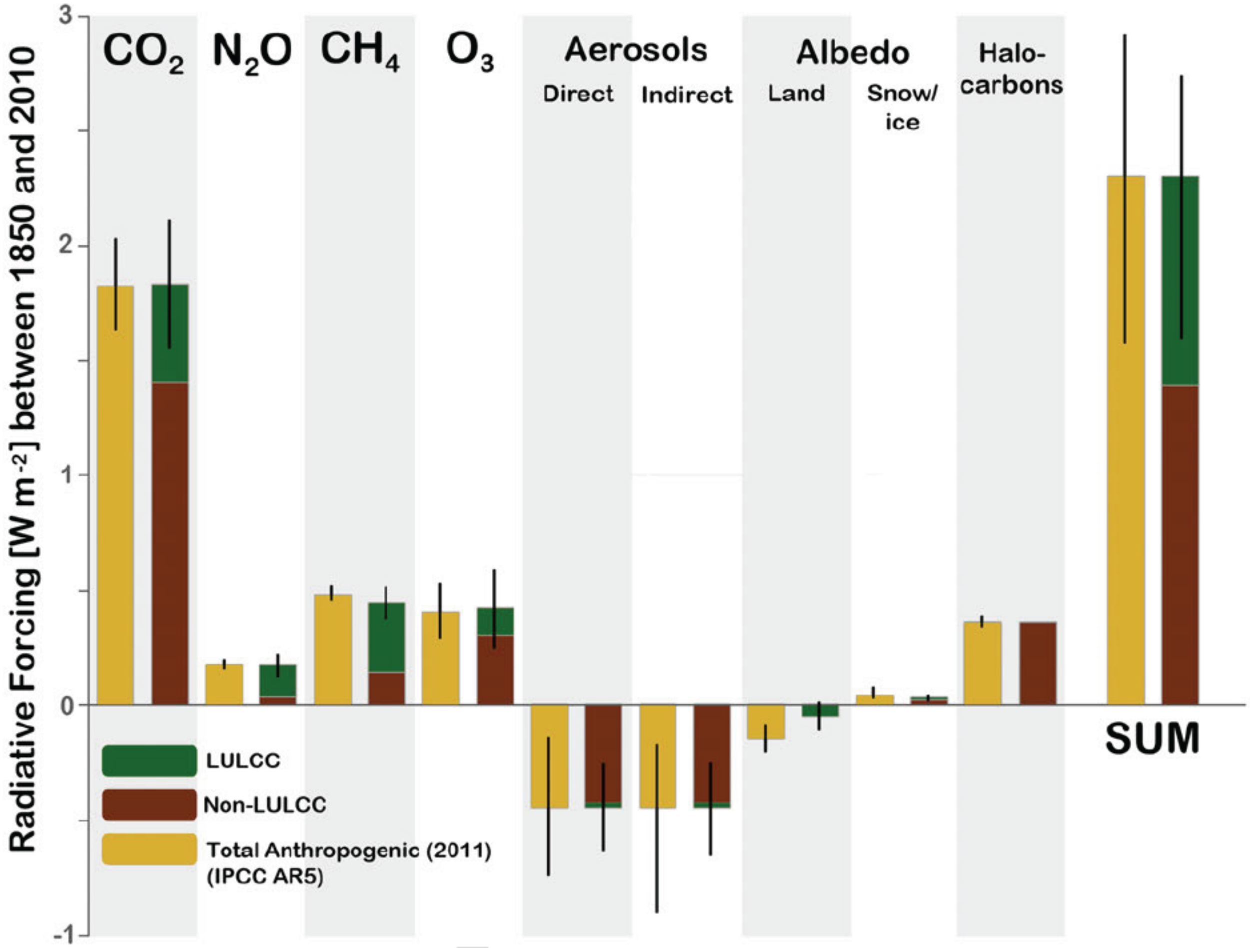
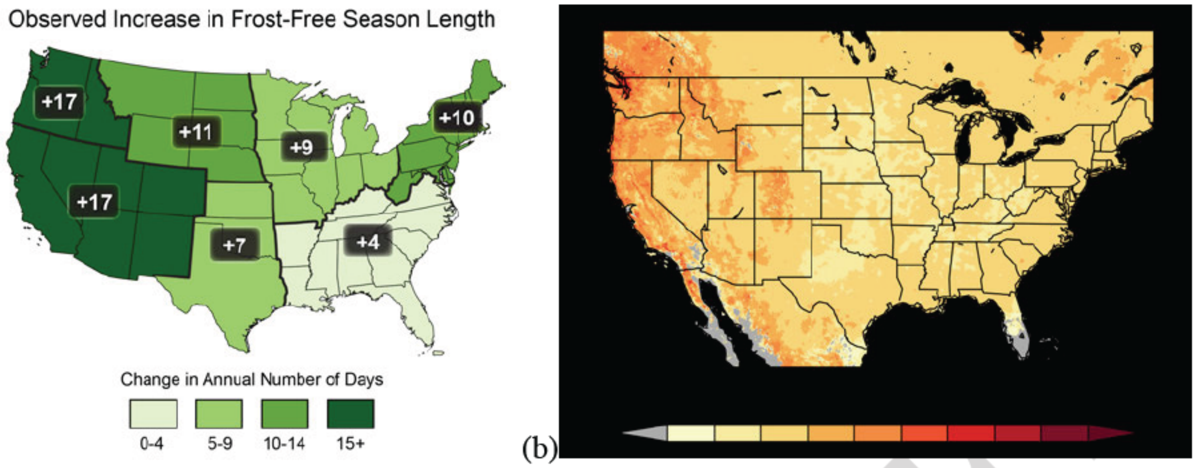


Figure 10.2. Anthropogenic radiative forcing (RF) contributions, separated by land-use and land-cover change (LULCC) and non-LULCC sources (green and maroon bars, respectively), are decomposed by atmospheric constituent to year 2010 in this diagram, using the year 1850 as the reference. Total anthropogenic RF contributions by atmospheric constituent (Myhre et al. 2013; see also Figure 2.3) are shown for comparison (yellow bars). Error bars represent uncertainties for total anthropogenic RF (yellow bars) and for the LULCC components (green bars; Ward et al. 2014). The SUM bars indicate the net RF when all anthropogenic forcing agents are combined. (Figure source: Ward et al. 2014).



(a) (b)

Figure 10.3. (a) Observed changes in the length of the frost-free season by region, where the frost-free season is defined as the number of days between the last spring occurrence and the first fall occurrence of a minimum temperature at or below 32°F. This change is expressed as the change in the average number of frost-free days in 1986–2015 compared to 1901–1960. (b) Projected changes in the length of the frost-free season at mid-century (2036–2065 as compared to 1976–2005) under the RCP8.5 scenario. Gray indicates areas that are not projected to experience a freeze in more than 10 of the 30 years (Figure source: (a) updated from Walsh et al. 2014; (b) NOAA NCEI / CICS-NC, data source: LOCA dataset).

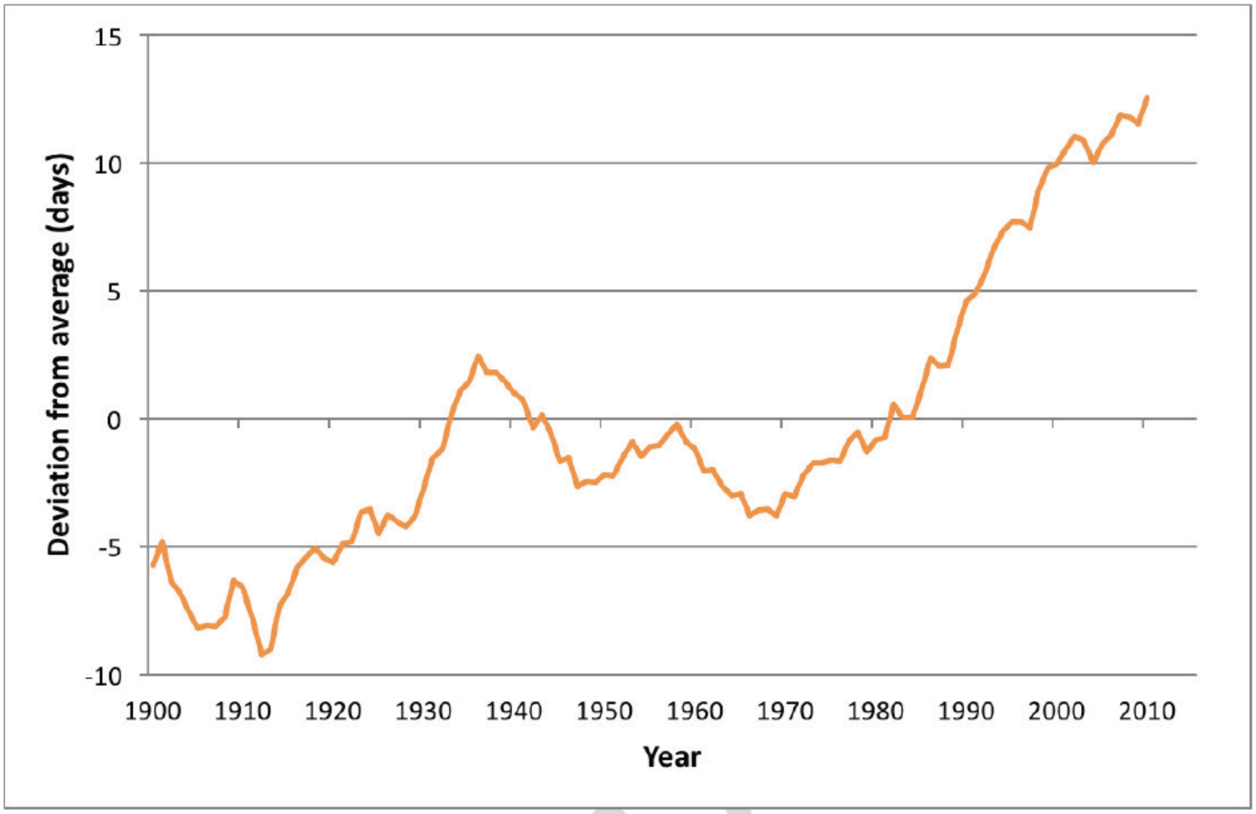


Figure 10.4. The length of the growing season in the contiguous 48 states compared with a long-term average (1895–2015), where "growing season" is defined by a daily minimum temperature threshold of 41°F. For each year, the line represents the number of days shorter or longer than the long-term average. The line was smoothed using an 11-year moving average. Choosing a different long-term average for comparison would not change the shape of the data over time. (Figure source: Kunkel 2016).

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