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

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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

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(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

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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

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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

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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).

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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

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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

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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

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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.

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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).

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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).

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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.

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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

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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

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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

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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

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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.

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FIGURES

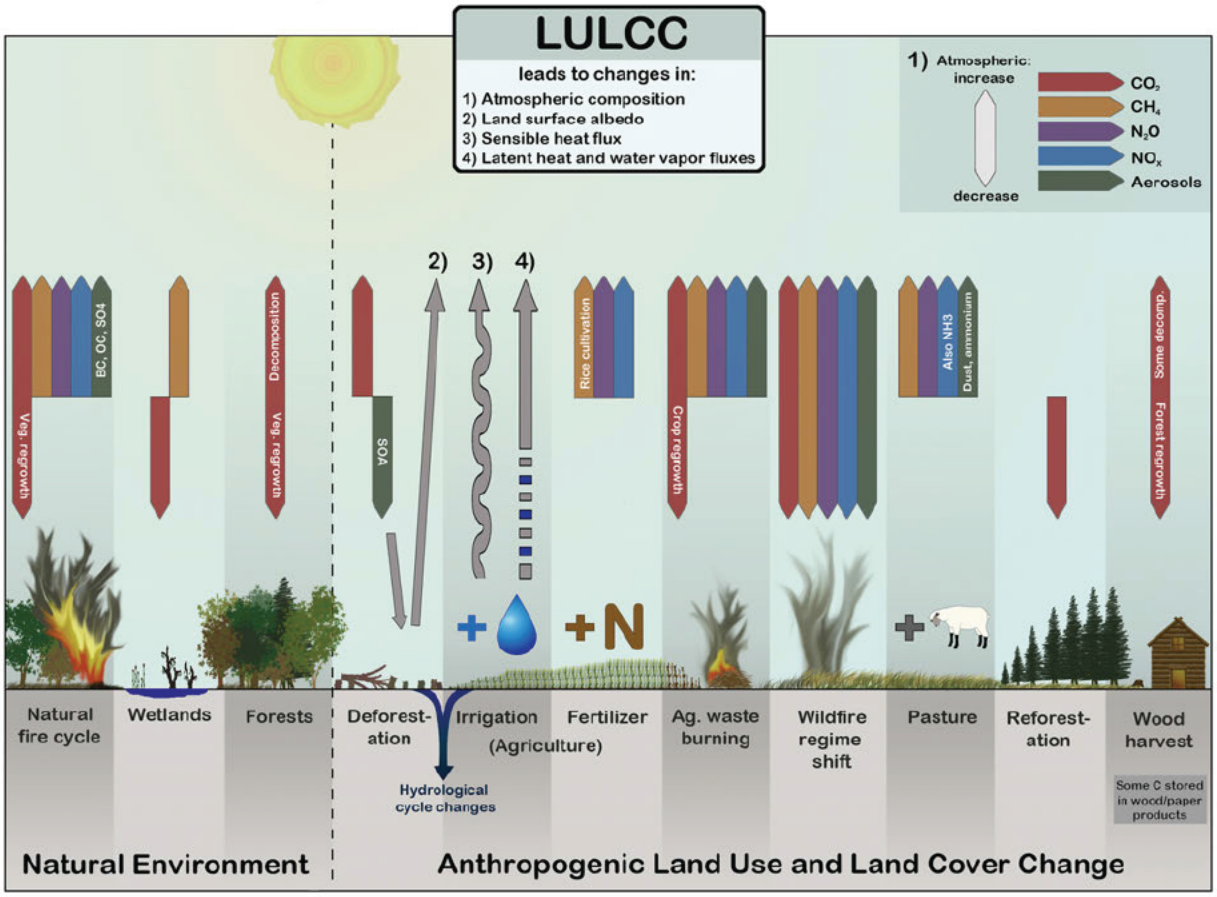

Figure 10.1. This graphical representation summarizes land–atmosphere interactions from natural and anthropogenic land-use and land-cover change (LULCC) contributions to radiative forcing. Emissions and sequestration of carbon and fluxes of nitrogen oxides, aerosols, and water shown here were used to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

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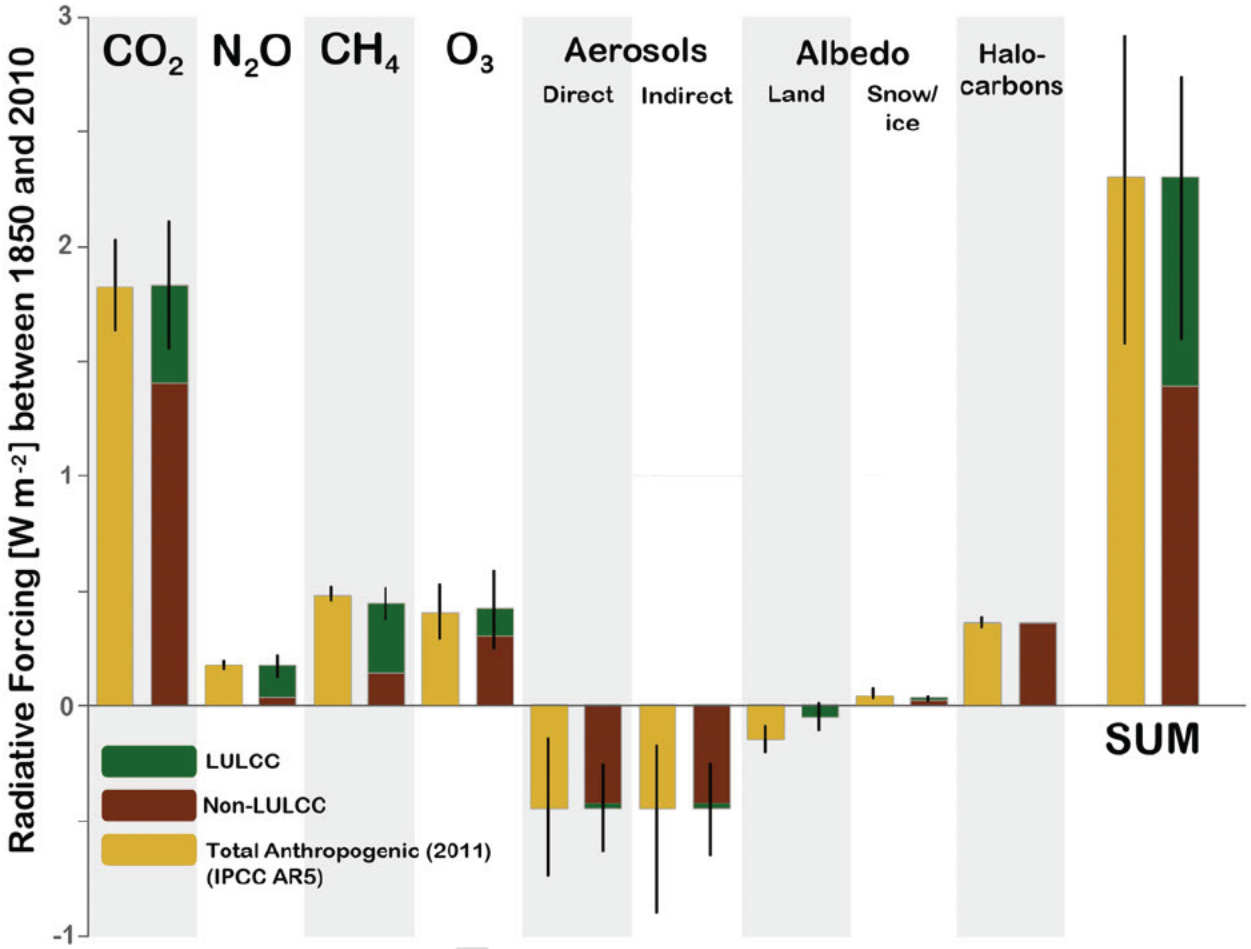

Figure 10.2. Anthropogenic radiative forcing (RF) contributions, separated by land-use and land-cover change (LULCC) and non-LULCC sources (green and maroon bars, respectively), are decomposed by atmospheric constituent to year 2010 in this diagram, using the year 1850 as the reference. Total anthropogenic 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).

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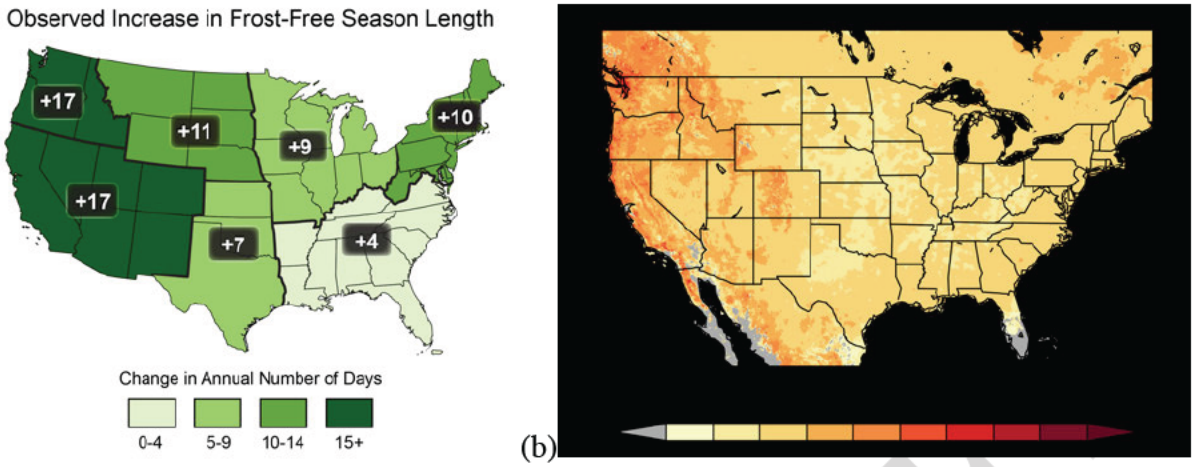

(a) (b)

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).

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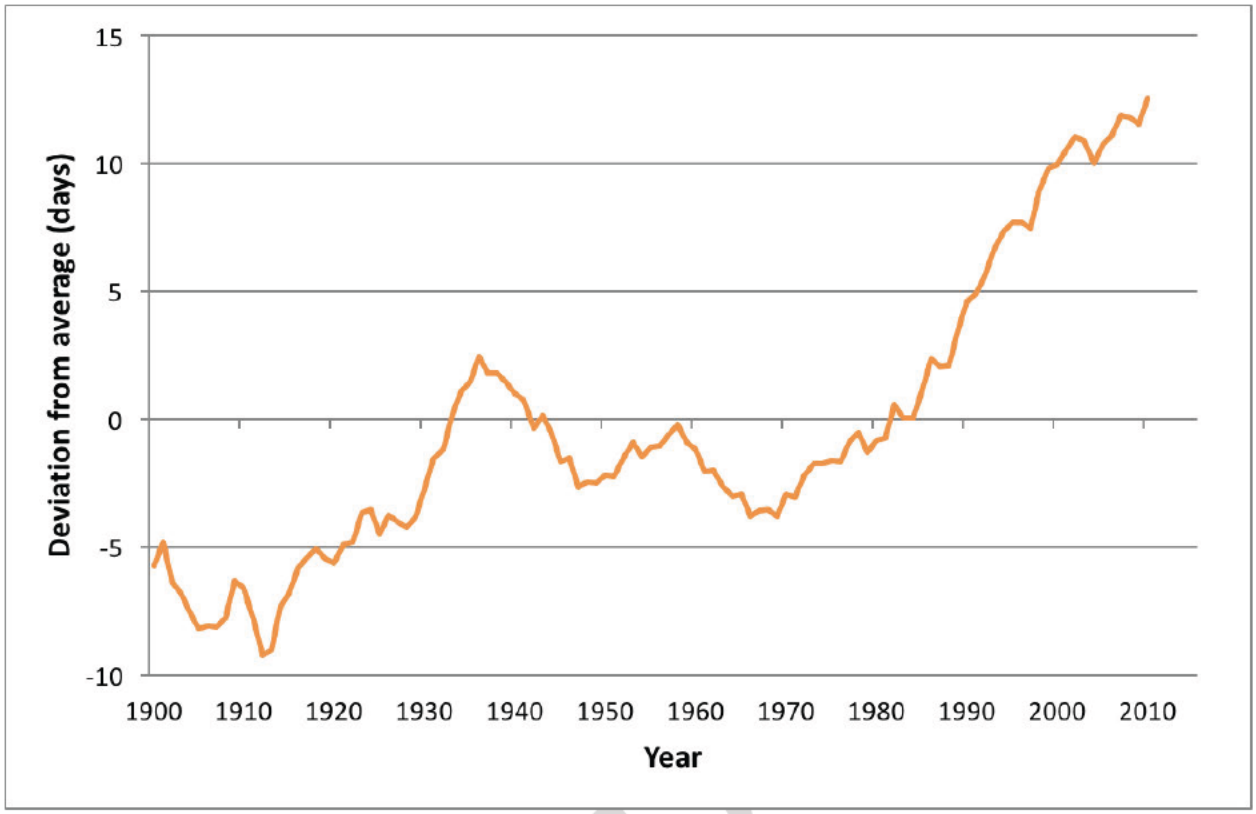

Figure 10.4. The length of the growing season in the contiguous 48 states compared with a long-term average (1895–2015), where "growing season" is defined by a daily minimum temperature threshold of 41°F. For each year, the line represents the number of days shorter or longer than the long-term average. 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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11. Arctic Changes and their Effects on Alaska and the Rest of the United States

KEY FINDINGS

1. Annual average near-surface air temperatures across Alaska and the Arctic have increased over the last 50 years at a rate more than twice as fast as the global average temperature. (*Very high confidence*)

2. Rising Alaskan permafrost temperatures are causing permafrost to thaw and become more discontinuous; this process releases additional CO_2 and methane, resulting in an amplifying feedback and additional warming (*high confidence*). The overall magnitude of the permafrost–carbon feedback is uncertain; however, it is clear that these emissions have the potential to complicate the ability to meet policy goals for the reduction of greenhouse gas concentrations.

3. Arctic land and sea ice loss observed in the last three decades continues, in some cases accelerating (*very high confidence*). It is *virtually certain* that Alaska glaciers have lost mass over the last 50 years, with each year since 1984 showing an annual average ice mass less than the previous year. Based on gravitational data from satellites, average ice mass loss from Greenland was –269 Gt per year between April 2002 and April 2016, accelerating in recent years (*high confidence*). Since the early 1980s, annual average Arctic sea ice has decreased in extent between 3.5% and 4.1% per decade, become thinner by between 4.3 and 7.5 feet, and began melting at least 15 more days each year. September sea ice extent has decreased between 10.7% and 15.9% per decade (*very high confidence*). Arctic-wide ice loss is expected to continue through the 21st century, *very likely* resulting in nearly sea ice-free late summers by the 2040s (*very high confidence*).

4. It is *virtually certain* that human activities have contributed to Arctic surface temperature warming, sea ice loss since 1979, glacier mass loss, and northern hemisphere snow extent decline observed across the Arctic (*very high confidence*). Human activities have *likely* contributed to more than half of the observed Arctic surface temperature rise and September sea ice decline since 1979 (*high confidence*).

5. Atmospheric circulation patterns connect the climates of the Arctic and the contiguous United States. Evidenced by recent record warm temperatures in the Arctic and emerging science, the midlatitude circulation has influenced observed Arctic temperatures and sea ice (*high confidence*). However, confidence is *low* regarding whether or by what mechanisms observed Arctic warming may have influenced the midlatitude circulation and weather patterns over the continental United States. The influence of Arctic changes on U.S. weather over the coming decades remains an open question with the potential for significant impact.

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11.1. Introduction

Climate changes in Alaska and across the Arctic continue to outpace changes occurring across the globe. The Arctic, defined as the area north of the Arctic Circle, is a vulnerable and complex system integral to Earth's climate. The vulnerability stems in part from the extensive cover of ice and snow, where the freezing point marks a critical threshold that when crossed has the potential to transform the region. Because of its high sensitivity to radiative forcing and its role in amplifying warming (Manabe and Wetherald 1975), the Arctic cryosphere is a key indicator of the global climate state. Accelerated melting of multiyear sea ice, mass loss from the Greenland Ice Sheet (GrIS), reduction of terrestrial snow cover, and permafrost degradation are stark examples of the rapid Arctic-wide response to global warming. These local Arctic changes influence global sea level, ocean salinity, the carbon cycle, and potentially atmospheric and oceanic circulation patterns. Arctic climate change has altered the global climate in the past (Knies et al. 2014) and will influence climate in the future.

As an Arctic nation, United States' adaptation, mitigation, and policy decisions depend on projections of future Alaskan and Arctic climate. Aside from uncertainties due to natural variability, scientific uncertainty, and greenhouse gas emissions uncertainty (see Ch. 4: Projections), additional unique uncertainties in our understanding of Arctic processes thwart projections, including mixed-phase cloud processes (Wyser et al. 2008); boundary layer processes (Bourassa et al. 2013); sea ice mechanics (Bourassa et al. 2013); and ocean currents, eddies, and tides that affect the advection of heat into and around the Arctic Ocean (Maslowski et al. 2012, 2014). The inaccessibility of the Arctic has made it difficult to sustain the high-quality observations of the atmosphere, ocean, land, and ice required to improve physically-based models. Improved data quality and increased observational coverage would help address societally relevant Arctic science questions.

Despite these challenges, our scientific knowledge is sufficiently advanced to effectively inform policy. This chapter documents significant scientific progress and knowledge about how the Alaskan and Arctic climate has changed and will continue to change.

11.2. Arctic Changes

11.2.1. Alaska and Arctic Temperature

Surface temperature—an essential component of the Arctic climate system—drives and signifies change, fundamentally controlling the melting of ice and snow. Further, the vertical profile of boundary layer temperature modulates the exchange of mass, energy, and momentum between the surface and atmosphere, influencing other components such as clouds (Kay and Gettelman 2009; Taylor et al. 2015). Arctic temperatures exhibit spatial and interannual variability due to interactions and feedbacks between sea ice, snow cover, atmospheric heat transports, vegetation, clouds, water vapor, and the surface energy budget (Overland et al. 2015b; Johannessen et al.

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2016; Overland and Wang 2016). Interannual variations in Alaskan temperatures are strongly influenced by decadal variability like the Pacific Decadal Oscillation (Hartmann and Wendler 2005; McAfee 2014; Ch. 5: Circulation and Variability). However, observed temperature trends exceed this variability.

Arctic surface and atmospheric temperatures have substantially increased in the observational record. Multiple observation sources, including land-based surface stations since at least 1950 and available meteorological reanalysis datasets, provide evidence that Arctic near-surface air temperatures have increased more than twice as fast as the global average (Serreze et al. 2009; Bekryaev et al. 2010; Screen and Simmonds 2010; Hartmann et al. 2013; Overland et al. 2014). Showing enhanced Arctic warming since 1981, satellite-observed Arctic average surface skin temperatures have increased by 1.08 ± 0.13°F (+0.60 ± 0.07°C) per decade (Comiso and Hall 2014). As analyzed in Chapter 6: Temperature Change (Figure 6.1), strong near-surface air temperature warming has occurred across Alaska exceeding 1.5°F (0.8°C) over the last 30 years. Especially strong warming has occurred over Alaska's North Slope during autumn. For example, Utqiagvik's (formally Barrow) warming since 1979 exceeds 7°F (3.8°C) in September, 12°F (6.6°C) in October, and 10°F (5.5°C) in November (Wendler et al. 2014).

Enhanced Arctic warming is a robust feature of the climate response to anthropogenic forcing (Collins et al. 2013; Taylor et al. 2013). An anthropogenic contribution to Arctic and Alaskan surface temperature warming over the past 50 years is *virtually certain* and *likely* amounting to more than 50% of observed warming (Gillett et al. 2008; Bindoff et al. 2013). One study argues that the natural forcing has not contributed to the long-term Arctic warming in a discernable way (Najafi et al. 2015). Also, other anthropogenic forcings (mostly aerosols) have *likely* offset up to 60% of the high-latitude greenhouse gas warming since 1913 (Najafi et al. 2015), suggesting that Arctic warming to date would have been larger without the offsetting aerosols influence. It is *virtually certain* that Arctic surface temperatures will continue to increase faster than the global mean through the 21st century (Christensen et al. 2013).

11.2.2. Arctic Sea Ice Change

Arctic sea ice strongly influences Alaskan, Arctic, and global climate by modulating exchanges of mass, energy, and momentum between the ocean and the atmosphere. Variations in Arctic sea ice cover also influence atmospheric temperature and humidity, wind patterns, clouds, ocean temperature, thermal stratification, and ecosystem productivity (Kay and Gettelman 2009; Kay et al. 2011a; Pavelsky et al. 2011; Taylor et al. 2011a; Boisvert et al. 2013; Vaughan et al. 2013; Solomon et al. 2014; Boisvert et al. 2015a,b; Johannessen et al. 2016). Arctic sea ice exhibits significant interannual, spatial, and seasonal variability driven by atmospheric wind patterns and cyclones, atmospheric temperature and humidity structure, clouds, radiation, sea ice dynamics, and the ocean (Ogi and Wallace 2007; Kwok and Untersteiner 2011; Taylor et al. 2011b; Stroeve et al. 2012a,b; Ogi and Rigor 2013; Carmack et al. 2015).

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Overwhelming evidence indicates that the character of Arctic sea ice is rapidly changing. Observational evidence shows Arctic-wide sea ice decline since 1979, accelerating ice loss since 2000, and some of the fastest loss along the Alaskan coast (Stroeve et al. 2014a,b; Comiso and Hall 2014; Wendler et al. 2014). Although sea ice loss is found in all months, satellite observations show the fastest loss in late summer and autumn (Stroeve et al. 2014a). Since 1979, the annual average Arctic sea ice extent has *very likely* decreased at a rate of 3.5%–4.1% per decade (Vaughan et al. 2013; Comiso and Hall 2014). Regional sea ice melt along the Alaskan coasts exceeds the Arctic average rates with declines in the Beaufort and Chukchi Seas of –4.1% and –4.7% per decade, respectively (Wendler et al. 2014). The annual minimum and maximum sea ice extent have decreased over the last 35 years by –13.3 ± 2.6% and –2.7± 0.5% per decade, respectively (Perovich et al. 2016). The ten lowest September sea ice extents over the satellite period have all occurred in the last ten years, the lowest in 2012. The 2016 September sea ice minimum tied with 2007 for the second lowest on record, but rapid refreezing resulted in the 2016 September monthly average extent being the fifth lowest. Despite the rapid initial refreezing, sea ice extent was again in record low territory during fall–winter 2016/2017 due to anomalously warm temperatures in the marginal seas around Alaska (Perovich et al. 2016), contributing to a new record low in winter ice-volume (See: http://psc.apl.uw.edu/research/projects/arctic-sea-ice-volume-anomaly, Schweiger et al. 2011).

Other important characteristics of Arctic sea ice have also changed, including thickness, age, and volume. Sea ice thickness is monitored using an array of satellite, aircraft, and vessel measurements (Vaughan et al. 2013; Stroeve et al. 2014a). The mean thickness of the Arctic sea ice during winter between 1980 and 2008 has decreased between 4.3 and 7.5 feet (1.3 and 2.3 meters) (Vaughan et al. 2013). The age distribution of sea ice has become younger since 1988. In March 2016, first-year (multi-year) sea ice accounted for 78% (22%) of the total extent, whereas in the 1980s first-year (multi-year) sea ice accounted for 55% (45%) (Perovich et al. 2016). Moreover, ice older than four years accounted for 16% of the March 1985 icepack but accounted for only 1.2% of the icepack in March 2016, indicating significant changes in sea ice volume (Perovich et al. 2016). The top two panels in Figure 11.1 show the September sea ice extent and age in 1984 and 2016, illustrating significant reductions in sea ice age (Tschudi et al. 2016). While these panels show only two years (beginning point and ending point) of the complete time series, these two years are representative of the overall trends discussed and shown in the September sea ice extent time series in the bottom panel of Fig 11.1. Younger, thinner sea ice is more susceptible to melt, therefore reductions in age and thickness imply a larger interannual variability of extent.

[INSERT FIGURE 11.1 HERE]

Sea ice melt season—defined as the number of days between spring melt onset and fall freeze-up—has lengthened Arctic-wide by at least five days per decade since 1979, with larger regional changes (Stroeve et al. 2014b; Parkinson 2014). Some of the largest observed changes in sea ice

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melt season (Figure 11.2) are found along Alaska's northern and western coasts, lengthening the melt season by 20–30 days per decade and increasing the annual number of ice-free days by more than 90 (Parkinson 2014). Summer sea ice retreat along coastal Alaska has led to longer open water seasons, making the Alaskan coastline more vulnerable to erosion (Chapin et al. 2014; Gibbs and Richmond 2015). Increased melt season length corresponds to increased absorption of solar radiation by the Arctic Ocean during summer and increases upper ocean temperature delaying fall freeze-up. Overall, this process significantly contributes to reductions in Arctic sea ice (Stroeve et al. 2012a; Stroeve et al. 2014b). Wind-driven sea ice export through the Fram Strait has not increased over the last 80 years (Vaughn et al. 2013), however one recent study suggests that it may have increased since 1979 (Smedsrud et. al. 2017).

[INSERT FIGURE 11.2 HERE]

It is *virtually certain* that there is an anthropogenic contribution to the observed Arctic sea ice decline since 1979. A range of modeling studies analyzing the September sea ice extent trends in simulations with and without anthropogenic forcing conclude that these declines cannot be explained by natural variability alone (Vinnikov et al. 1999; Stroeve et al. 2007; Min et al. 2008; Kay et al. 2011b; Day et al. 2012; Wang and Overland 2012). Further, observational-based analyses considering a range of anthropogenic and natural forcing mechanisms for September sea ice loss reach the same conclusion (Notz and Marotzke 2012). Considering the occurrence of individual September sea ice anomalies, internal climate variability alone *very likely* could not have caused recently observed record low Arctic sea ice extents, such as in September 2012 (Zhang and Knutson 2013; Kirchmeyer-Young et al. 2017). The potential contribution of natural variability to Arctic sea ice trends is significant (Kay et al. 2011b; Jahn et al. 2016; Swart et al. 2015). One recent study (Ding et al. 2017) indicates that internal variability dominates Arctic atmospheric circulation trends, accounting for 30%–50% of the sea ice reductions since 1979, and up to 60% in September. However, previous studies indicate that the contributions from internal variability are smaller than 50% (Kay et al. 2011b; Day et al. 2012). This apparent significant contribution of natural variability to sea ice decline is consistent with the statement that *likely* more than half of the observed sea ice loss since 1979 has an anthropogenic contribution.

Continued sea ice loss is expected across the Arctic, which is *very likely* to result in late summers becoming nearly ice-free (areal extent less than 10^6 km^2 or approximately 3.9×10^5 mi^2) by mid-century (Collins et al. 2013; Snape and Forster 2014). Natural variability (Wettstein and Deser 2014), future emissions, and model uncertainties (Gagné et al. 2015; Stroeve and Notz 2015; Swart et al. 2015) all influence sea ice projections. One study suggests that internal variability alone accounts for a 20-year prediction uncertainty in the timing of the first occurrence of an ice-free summer, whereas emissions scenario differences between RCP8.5 and RCP4.5 add only 5 years (Jahn et al. 2016). Projected September sea ice reductions by 2081–2100 range from 43% for RCP2.6 to 94% for RCP8.5 (Collins et al. 2013). However, September sea ice projections

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over the next few decades are similar for the different anthropogenic forcing associated with these scenarios; scenario dependent sea ice loss only becomes apparent after 2050. Another study (Notz and Stroeve 2016) indicates that the total sea ice loss scales roughly linearly with CO_2 emissions, such that an additional 1,000 GtC from present day levels corresponds to ice-free conditions in September. A key message from the Third National Climate Assessment (NCA3; Melillo et al. 2014) was that Arctic sea ice is disappearing. The fundamental conclusion of this assessment is unchanged; additional research corroborates the NCA3 statement.

11.2.3. Arctic Ocean and Marginal Seas

SEA SURFACE TEMPERATURE

Arctic Ocean sea surface temperatures (SSTs) have increased since comprehensive records became available in 1982. Satellite-observed Arctic Ocean SSTs, poleward of 60°N, exhibit a trend of 0.16 ± 0.02°F (0.09 ± 0.01°C) per decade (Comiso and Hall 2014). Arctic Ocean SST is controlled by a combination of factors, including solar radiation and energy transport from ocean currents and atmospheric winds. Summertime Arctic Ocean SST trends and patterns strongly couple with sea ice extent; however, clouds, ocean color, upper-ocean thermal structure, and atmospheric circulation also play a role (Ogi and Rigor 2013; Rhein et al. 2013). Along coastal Alaska, SSTs in the Chukchi Sea exhibit a statistically significant (95% confidence) trend of 0.9 ± 0.5°F (0.5 ± 0.3°C) per decade (Timmermans and Proshutinksy 2015).

Arctic Ocean temperatures also increased at depth (Polyakov et al. 2012; Rhein et al. 2013). Since 1970, Arctic Ocean Intermediate Atlantic Water—located between 150 and 900 meters—has warmed by 0.86 ± 0.09°F (0.48 ± 0.05°C) per decade; the most recent decade being the warmest (Polyakov et al. 2012). The observed temperature level is unprecedented in the last 1,150 years for which proxy indicators provide records (Spielhagen et al. 2011; Jungclaus et al. 2014). The influence of Intermediate Atlantic Water warming on future Alaska and Arctic sea ice loss is unclear (Döscher et al. 2014; Carmack et al. 2015).

ALASKAN SEA LEVEL RISE

The Alaskan coastline is vulnerable to sea level rise (SLR); however, strong regional variability exists in current trends and future projections. Some regions are experiencing relative sea level fall, whereas others are experiencing relative sea level rise, as measured by tide gauges that are part of NOAA's National Water Level Observation Network. These tide gauge data show sea levels rising fastest along the northern coast of Alaska but still slower than the global average, due to isostatic rebound (Church et al. 2013; Ch. 12: Sea Level Rise). However, considerable uncertainty in relative sea level rise exists to due to a lack of tide gauges; for example, no tide gauges are located between Bristol Bay and Norton Sound or between Cape Lisburne and Prudhoe Bay. Under almost all future scenarios, SLR along most of the Alaskan coastline is projected to be less than the global average (Ch. 12: Sea Level Rise).

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SALINITY

Arctic Ocean salinity influences the freezing temperature of sea ice (less salty water freezes more readily) and the density profile representing the integrated effects of freshwater transport, river runoff, evaporation, and sea ice processes. Arctic Ocean salinity exhibits multidecadal variability, hampering the assessment of long-term trends (Rawlins et al. 2010). Emerging evidence suggests that the Arctic Ocean and marginal sea salinity has decreased in recent years despite short-lived regional salinity increases between 2000 and 2005 (Rhein et al. 2013). Increased river runoff, rapid melting of sea and land ice, and changes in freshwater transport have influenced observed Arctic Ocean salinity (Rhein et al. 2013; Köhl and Serra 2014).

OCEAN ACIDIFICATION

Arctic Ocean acidification is occurring at a faster rate than the rest of the globe (Mathis et al. 2015; see also Ch. 13: Ocean Changes). Coastal Alaska and its ecosystems are especially vulnerable to ocean acidification because of the high sensitivity of Arctic Ocean water chemistry to changes in sea ice, respiration of organic matter, upwelling, and increasing river runoff (Mathis et al. 2015). Sea ice loss and a longer melt season contribute to increased vulnerability of the Arctic Ocean to acidification by lowering total alkalinity, permitting greater upwelling, and influencing the primary production characteristics in coastal Alaska (Arrigo et al. 2008; Cai et al. 2010; Hunt et al. 2011; Stabeno et al. 2012; Mathis et al. 2012; Bates et al. 2014). Global-scale modeling studies suggest that the largest and most rapid changes in pH will continue along Alaska's coast, indicating that ocean acidification may increase enough by the 2030s to significantly influence coastal ecosystems (Mathis et al. 2015).

11.2.4. Boreal Wildfires

Alaskan wildfire activity has increased in recent decades. This increase has occurred both in the boreal forest (Flannigan et al. 2009) and in the Arctic tundra (Hu et al. 2015), where fires historically were smaller and less frequent. A shortened snow cover season and higher temperatures over the last 50 years (Derksen et al. 2015) make the Arctic more vulnerable to wildfire (Flannigan et al. 2009; Hu et al. 2015; Young et al. 2016). Total area burned and the number of large fires (those with area greater than 1000 km^2 or 386 mi^2) in Alaska exhibit significant interannual and decadal variability, from influences of atmospheric circulation patterns and controlled burns, but have *likely* increased since 1959 (Kasischke and Turetsky 2006). The most recent decade has seen an unusually large number of years with anomalously large wildfires in Alaska (Sanford et al. 2015). Studies indicate that anthropogenic climate change has *likely* lengthened the wildfire season and increased the risk of severe fires (Partain et al. 2016). Further, wildfire risks are expected to increase through the end of the century due to warmer, drier conditions (French et al. 2015; Young et al. 2017). Using climate simulations to force an ecosystem model over Alaska (Alaska Frame-Based Ecosystem Code, ALFRESCO), the total area burned is projected to increase between 25% and 53% by 2100 (Joly et al. 2012). A

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transition into a regime of fire activity unprecedented in the last 10,000 years is possible (Kelly et al. 2013). We conclude that there is *medium confidence* for a human-caused climate change contribution to increased forest fire activity in Alaska in recent decades. See Chapter 8: Drought, Floods, and Wildfires for more details.

A significant amount of the total global soil carbon is found in the boreal forest and tundra ecosystems, including permafrost (McGuire et al. 2009; Mishra and Riley 2012; Mishra et al. 2013). Increased fire activity could deplete these stores, releasing them to the atmosphere to serve as an additional source of atmospheric CO_2 (McGuire et al. 2009; Kelly et al. 2016). Increased fires may also enhance the degradation of Alaska's permafrost by blackening the ground, reducing surface albedo, and removing protective vegetation (Swanson 1996; Yoshikawa et al. 2003; Myers-Smith et al. 2008; Brown et al. 2015).

11.2.5. Snow Cover

Snow cover extent has significantly decreased across the Northern Hemisphere and Alaska over the last decade (Derksen and Brown 2012; Kunkel et al. 2016; see also Ch. 7: Precipitation Change and Ch. 10: Land Cover). Northern Hemisphere June snow cover decreased by more than 65% between 1967 and 2012 (Brown and Robinson 2011; Vaughan et al. 2013), at a trend of −17.2% per decade since 1979 (Derksen et al. 2015). June snow cover dipped below 3 million square km (approximately 1.16 million square miles) for the fifth time in six years between 2010 and 2015, a threshold not crossed in the previous 43 years of record (Derksen et al. 2015). Early season snow cover in May, which affects the accumulation of solar insolation through the summer, has also declined at −7.3% per decade, due to reduced winter accumulation from warmer temperatures. Regional trends in snow cover duration vary, with some showing earlier onsets while others show later onsets (Derksen et al. 2015). In Alaska, the 2016 May statewide snow coverage of 595,000 square km (approximately 372,000 square miles) was the lowest on record dating back to 1967; the snow coverage of 2015 was the second lowest, and 2014 was the fourth lowest.

Human activities have contributed to observed snow cover declines over the last 50 years. Attribution studies indicate that observed trends in Northern Hemisphere snow cover cannot be explained by natural forcing alone, but instead require anthropogenic forcing (Rupp et al. 2013; Bindoff et al. 2013; Kunkel et al. 2016). Declining snow cover is expected to continue and will be affected by both the anthropogenic forcing and evolution of Arctic ecosystems. The observed tundra shrub expansion and greening (Myers-Smith et al. 2011; Mao et al. 2016) affects melt by influencing snow depth, melt dynamics, and the local surface energy budget. Nevertheless, model simulations show that future reductions in snow cover influence biogeochemical feedbacks and warming more strongly than changes in vegetation cover and fire in the North American Arctic (Euskirchen et al. 2016).

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11.2.6. Continental Ice Sheets and Mountain Glaciers

Mass loss from ice sheets and glaciers influences sea level rise, the oceanic thermohaline circulation, and the global energy budget. Moreover, the relative contribution of GrIS to global sea level rise continues to increase, exceeding the contribution from thermal expansion (see Ch. 12: Sea Level Rise). Observational and modeling studies indicate that GrIS and glaciers in Alaska are out of mass balance with current climate conditions and are rapidly losing mass (Vaughan et al. 2013; Zemp et al. 2015). In recent years, mass loss has accelerated and is expected to continue (Zemp et al. 2015; Harig and Simons 2016).

Dramatic changes have occurred across GrIS, particularly at its margins. GrIS average annual mass loss from January 2003 to May 2013 was –244 ± 6 Gt per year (approximately 0.26 inches per decade sea level equivalent) (Harig and Simons 2016). One study indicates that ice mass loss from Greenland was –269 Gt per year between April 2002 and April 2016 (Perovich et al. 2016). Increased surface melt, runoff, and increased outlet glacier discharge from warmer air temperatures are primary contributing factors (Howat et al. 2008; van den Broeke et al. 2009; Rignot et al. 2010; Straneo et al. 2011; Khan et al. 2014). The effects of warmer air and ocean temperatures on GrIS can be amplified by ice dynamical feedbacks, such as faster sliding, greater calving, and increased submarine melting (Joughin et al. 2008; Holland et al. 2008; Rignot et al. 2010; Bartholomew et al. 2011). Shallow ocean warming and regional ocean and atmospheric circulation changes also contribute to mass loss (Dupont and Alley 2005; Lim et al. 2016; Tedesco et al. 2016). The underlying mechanisms of the recent discharge speed-up remain unclear (Straneo et al. 2010; Johannessen et al. 2011); however, warmer subsurface ocean and atmospheric temperatures (Velicogna 2009; van den Broeke et al. 2009; Andresen et al. 2012) and meltwater penetration to the glacier bed (Johannessen et al. 2011; Mernild et al. 2012) *very likely* contribute.

Annual average ice mass from Arctic-wide glaciers has decreased every year since 1984 (AMAP 2011; Pelto 2015; Zemp et al. 2015), with significant losses in Alaska, especially over the past two decades (Figure 11.3; Vaughan et al. 2013; Sharp et al. 2015). Figure 11.4 illustrates observed changes from U.S. Geological Survey repeat photography of Alaska's Muir Glacier, retreating more than 4 miles between 1941 and 2004, and its tributary the Riggs Glacier. Total glacial ice mass in the Gulf of Alaska region has declined steadily since 2003 (Harig and Simons 2016). NASA's Gravity Recovery and Climate Experiment (GRACE) indicates mass loss from the northern and southern parts of the Gulf of Alaska region of –36 ± 4 Gt per year and –4 ± 3 Gt per year, respectively (Harig and Simons 2016). Studies show imbalances in Alaskan glaciers, indicating that melt will continue through the 21st century (Zemp et al. 2015; Mengel et al. 2016). Multiple datasets indicate that it is *extremely likely* that Alaskan glaciers have lost mass over the last 50 years and will continue to do so (Larsen et al. 2015).

[INSERT FIGURES 11.3 AND 11.4 HERE]

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11.3. Arctic Feedbacks on the Lower 48 and Globally

11.3.1. Linkages between Arctic Warming and Lower Latitudes

Midlatitude circulation influences Arctic climate and climate change (Rigor et al. 2002; Graversen 2006; Perlwitz et al. 2015; Francis et al. 2017; Screen et al. 2012; Park et al. 2015; Lee 2014; Lee et al. 2011; Ding et al. 2014; Screen and Francis 2016; Overland and Wang 2016). Record warm Arctic temperatures in winter 2016 resulted primarily from the transport of midlatitude air into the Arctic, demonstrating the significant midlatitude influence (Overland et al. 2016). Emerging science demonstrates that warm, moist air intrusions from midlatitudes results in increased downwelling longwave radiation, warming the Arctic surface and hindering wintertime sea ice growth (Liu and Key 2014; Lee 2014; Park et al. 2015; Woods and Caballero 2016).

The extent to which enhanced Arctic surface warming and sea ice loss influence the large-scale atmospheric circulation and midlatitude weather and climate extremes has become an active research area (Overland et al. 2016; Francis et al. 2017). Several pathways have been proposed (see reference in Cohen et al. 2014 and Barnes and Screen 2015): reduced meridional temperature gradient, a more sinuous jet-stream, trapped atmospheric waves, modified storm tracks, weakened stratospheric polar vortex. While modeling studies link a reduced meridional temperature gradient to fewer cold temperature extremes in the continental United States (Ayarzagüena and Screen 2016; Sun et al. 2016; Screen et al. 2015a,b), other studies hypothesize that a slower jet stream may amplify Rossby waves and increase the frequency of atmospheric blocking, causing more persistent and extreme weather in midlatitudes (Francis and Vavrus 2012).

Multiple observational studies suggest that the concurrent changes in the Arctic and Northern Hemisphere large-scale circulation since the 1990s did not occur by chance, but were caused by arctic amplification (Cohen et al. 2014; Vihma 2014; Barnes and Screen 2015). Reanalysis data suggest a relationship between arctic amplification and observed changes in persistent circulation phenomena like blocking and planetary wave amplitude (Francis and Skific 2015; Francis and Vavrus 2012, 2015). The recent multi-year California drought serves as an example of an event caused by persistent circulation phenomena (Swain et al. 2014; Seager et al. 2015; Teng and Branstator 2017; see Ch. 5: Circulation and Variability and Ch. 8: Drought, Floods, and Wildfires). Robust empirical evidence is lacking because the Arctic sea ice observational record is too short (Overland et al. 2015a) or because the atmospheric response to arctic amplification depends on the prior state of the atmospheric circulation, reducing detectability (Overland et al. 2016). Furthermore, it is not possible to draw conclusions regarding the direction of the relationship between Arctic warming and midlatitude circulation based on empirical correlation and covariance analyses alone. Observational analyses have been combined with modeling studies to test causality statements.

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Studies with simple models and Atmospheric General Circulation Models (AGCMs) provide evidence that Arctic warming can affect midlatitude jet streams and location of storm tracks (Barnes and Screen 2015; Overland et al. 2016; Francis et al. 2017). In addition, analysis of CMIP5 models forced with increasing greenhouse gases suggests that the magnitude of arctic amplification affects the future midlatitude jet position, specifically during boreal winter (Barnes and Polvani 2015). However, the effect of arctic amplification on blocking is not clear (Hoskins and Woollings 2015; Ch. 5: Circulation and Variability).

Regarding attribution, AGCM simulations forced with observed changes in Arctic sea ice suggest that the sea ice loss effect on observed recent midlatitude circulation changes and winter climate in the continental United States is small compared to natural large-scale atmospheric variability (Screen et al. 2012; Perlwitz et al. 2015; Sigmond and Fyfe 2016; Sun et al. 2016). It is argued, however, that climate models do not properly reproduce the linkages between arctic amplification and lower latitude climate due to model errors, including incorrect sea ice–atmosphere coupling and poor representation of stratospheric processes (Cohen et al. 2013; Francis et al. 2017)

In summary, emerging science demonstrates a strong influence of the midlatitude circulation on the Arctic, affecting temperatures and sea ice (*high confidence*). The influence of Arctic changes on the midlatitude circulation and weather patterns are an area of active research. Currently, confidence is *low* regarding whether or by what mechanisms observed Arctic warming may have influenced midlatitude circulation and weather patterns over the continental United States. The nature and magnitude of arctic amplification's influence on U.S. weather over the coming decades remains an open question.

11.3.2. Freshwater Effects on Ocean Circulation

The addition of freshwater to the Arctic Ocean from melting sea ice and land ice can influence important Arctic climate system characteristics, including ocean salinity, altering ocean circulation, density stratification, and sea ice characteristics. Observations indicate that river runoff is increasing, driven by land ice melt, adding freshwater to the Arctic Ocean (Nummelin et al. 2016). Melting Arctic sea and land ice combined with time-varying atmospheric forcing (Giles et al. 2012; Köhl and Serra 2014) control Arctic Ocean freshwater export to the North Atlantic. Large-scale circulation variability in the central Arctic not only controls the redistribution and storage of freshwater in the Arctic (Köhl and Serra 2014) but also the export volume (Morison et al. 2012). Increased freshwater fluxes can weaken open ocean convection and deep water formation in the Labrador and Irminger seas, weakening the Atlantic meridional overturning circulation (AMOC; Rahmstorf et al. 2015; Yang et al. 2016). AMOC-associated poleward heat transport substantially contributes to North American and continental European climate; any AMOC slow-down could have implications for global climate change as well (Smeed et al. 2014; Liu et al. 2017; see Ch. 15: Potential Surprises). Connections to subarctic

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ocean variations and the Atlantic Meridional Overturning Circulation have not been conclusively established and require further investigation (see Ch. 13: Ocean Changes).

11.3.3. Permafrost–Carbon Feedback

Alaska and Arctic permafrost characteristics have responded to increased temperatures and reduced snow cover in most regions since the 1980s (AMAP 2011). The permafrost warming rate varies regionally; however, colder permafrost is warming faster than warmer permafrost (Vaughan et al. 2013; Romanovsky et al. 2015). This feature is most evident across Alaska, where permafrost on the North Slope is warming more rapidly than in the interior. Permafrost temperatures across the North Slope at various depths ranging from 39 to 65 feet (12 to 20 meters) have warmed between 0.3° and 1.3°F (0.2° and 0.7°C) per decade over the observational period (Figure 11.5; Romanovsky et al. 2016). Permafrost active layer thickness increased across much of the Arctic while showing strong regional variations (AMAP 2011; Shiklomanov et al. 2012; Vaughan et al. 2013). Further, recent geologic survey data indicate significant permafrost thaw slumping in northwestern Canada and across the circumpolar Arctic that indicate significant ongoing permafrost thaw, potentially priming the region for more rapid thaw in the future (Kokelj et al. 2017). Continued degradation of permafrost and a transition from continuous to discontinuous permafrost is expected over the 21st century (Vaughan et al. 2013; Schuur et al. 2015; Grosse et al. 2016).

[INSERT FIGURE 11.5 HERE]

Permafrost contains large stores of carbon. Though the total contribution of these carbon stores to global methane emission is uncertain, Alaska's permafrost contains rich and vulnerable organic carbon soils (Tarnocai et al. 2009; Mishra and Riley 2012; Schuur et al. 2015). Thus, warming Alaska permafrost is a concern for the global carbon cycle as it provides a possibility for a significant and potentially uncontrollable release of carbon, complicating the ability to meet global policy goals. Current methane emissions from Alaskan Arctic tundra and boreal forests contribute a small fraction of the global methane (CH_4) budget (Chang et al. 2014). However, gas flux measurements have directly measured the release of CO_2 and CH_4 from Arctic permafrost (Schuur et al. 2009). Recent measurement indicate that cold season methane emissions (after snowfall) are greater than summer emissions in Alaska, and methane emissions in upland tundra are greater than in wetland tundra (Zona et al. 2016).

The permafrost–carbon feedback represents the additional release of CO_2 and CH_4 from thawing permafrost soils providing additional radiative forcing, a source of a potential surprise (Treat et al. 2015; Ch. 15: Potential Surprises). Thawing permafrost makes previously frozen organic matter available for microbial decomposition, producing CO_2 and CH_4. The specific condition under which microbial decomposition occurs, aerobic or anaerobic, determines the proportion of CO_2 and CH_4 released. This distinction has potentially significant implications, as CH_4 has a 100-year global warming potential 35 times that of CO_2 (Myhre et al. 2013). Emerging science

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indicates that 3.4 times more carbon is released under aerobic conditions than anaerobic conditions, and 2.3 times more carbon after accounting for the stronger greenhouse effect of CH_4 (Schädel et al. 2016). Additionally, CO_2 and CH_4 production strongly depends on vegetation and soil properties (Treat et al. 2015).

Combined data and modeling studies indicate a positive permafrost–carbon feedback with a global sensitivity between –14 and –19 GtC per °C (approximately –25 to –34 GtC per °F) soil carbon loss (Koven et al. 2015a,b) resulting in a total 120 ± 85 GtC release from permafrost by 2100 and an additional global temperature increase of 0.52 ± 0.38°F (0.29 ± 0.21°C) by the permafrost–carbon feedback (Schaefer et al. 2014). More recently, Chadburn et al. (2017) infer a –4 million km^2 per °C (or approximately 858,000 mi^2 per °F) reduction in permafrost area to globally averaged warming at stabilization by constraining climate models with the observed spatial distribution of permafrost; this sensitivity is 20% higher than previous studies. In the coming decades, enhanced high-latitude plant growth and its associated CO_2 sink should partially offset the increased emissions from permafrost thaw (Friedlingstein et al. 2006; Schaefer et al. 2014; Schuur et al. 2015); thereafter, decomposition is expected to dominate uptake. Permafrost thaw is occurring faster than models predict due to poorly understood deep soil, ice wedge, and thermokarst processes (Fisher et al. 2014; Koven et al. 2015a; Liljedahl et al. 2016). Additional, uncertainty stems from the surprising uptake of methane from mineral soils (Oh et al. 2016). There is *high confidence* in the positive sign of the permafrost–carbon feedback, but *low confidence* in the feedback magnitude.

11.3.4. Methane Hydrate Instability

Significant stores of CH_4, in the form of methane hydrates (also called clathrates), lie below permafrost and under the global ocean. The estimated total global inventory of methane hydrates ranges from 500 to 3,000 GtC (Archer 2007; Ruppel 2011; Piñero et al. 2013) with a central estimate of 1800 GtC (Ruppel and Kessler 2017). Methane hydrates are solid compounds formed at high pressures and cold temperatures trapping methane gas within the crystalline structure of water. In the Arctic Ocean and along the shallow coastal Alaskan seas, methane hydrates form on shallow but cold continental shelves and may be vulnerable to small increases in ocean temperature (Bollman et al. 2010; Ruppel 2011; Ruppel and Kessler 2017).

Rising sea levels and warming oceans have a competing influence on methane hydrate stability (Bollman et al. 2010; Hunter et al. 2013). Studies indicate that the temperature effect dominates and that the overall influence is *likely* a destabilizing effect. Projected warming rates for the 21st century Arctic Ocean are not expected to lead to sudden or catastrophic destabilization of sea floor methane hydrates (Kretschmer et al. 2015). Recent observations indicate increased CH_4 emission from the Arctic sea floor near Svalbard; however, these emissions are not reaching the atmosphere (Graves et al. 2015; Ruppel and Kessler 2017). It is likely that most of the methane hydrate deposits will remain stable for the foreseeable future (the next few thousand years).

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However, deposits off of coastal Alaska are among the most vulnerable and are expected to continue to release CH_4 during the 21st century (Archer 2007; Kretschmer et al. 2015).

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TRACEABLE ACCOUNTS

Key Finding 1

Annual average near-surface air temperatures across Alaska and the Arctic have increased over the last 50 years at a rate more than twice as fast as the global average temperature. (*Very high confidence*)

Description of evidence base

The Key Finding is supported by observational evidence from ground-based observing stations, satellites, and data-model temperature analyses from multiple sources and independent analysis techniques (Serrreze et al. 2009; Bekryaev et al. 2010; Screen and Simmonds 2010; Hartmann et al. 2013; Overland et al. 2014; Comiso and Hall 2014; Wendler et al. 2014). For more than 40 years, climate models have predicted enhanced Arctic warming, indicating a solid grasp on the underlying physics and positive feedbacks driving the accelerated Arctic warming (Manabe and Wetherald 1975; Collins et al. 2013; Taylor et al. 2013). Lastly, similar statements have been made in NCA3 (Melillo et al. 2014), IPCC AR5 (Hartmann et al. 2013), and in other Arctic-specific assessments such as the Arctic Climate Impacts Assessment (ACIA 2005) and Snow, Water, Ice and Permafrost in the Arctic (AMAP 2011).

Major Uncertainties

The lack of high quality and restricted spatial resolution of surface and ground temperature data over many Arctic land regions and essentially no measurements over the Central Arctic Ocean hampers the ability to better refine the rate of Arctic warming and completely restricts our ability to quantify and detect regional trends, especially over the sea ice. Climate models generally produce an Arctic warming between two to three times the global mean warming. A key uncertainty is our quantitative knowledge of the contributions from individual feedback processes in driving the accelerated Arctic warming. Reducing this uncertainty will help constrain projections of future Arctic warming.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

V*ery high confidence* that the Arctic surface and air temperatures have warmed across Alaska and the Arctic at a much faster rate than the global average is provided by the multiple datasets analyzed by multiple independent groups indicating the same conclusion. Additionally, climate models capture the enhanced warming in the Arctic indicating a solid understanding of the underlying physical mechanisms.

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If appropriate, estimate likelihood of impact or consequence, including short description of basis of estimate

It is *very likely* that the accelerated rate of Arctic warming will have a significant consequence for the United States due to accelerated land and sea ice melt driving changes in the ocean including sea level rise threatening our coastal communities and freshening of sea water that is influencing marine ecology.

Summary sentence or paragraph that integrates the above information

Annual average near-surface air temperatures across Alaska and the Arctic have increased over the last 50 years at a rate more than twice the global average. Observational studies using ground-based observing stations and satellites analyzed by multiple independent groups support this finding. The enhanced sensitivity of the Arctic climate system to anthropogenic forcing is also supported by climate modeling evidence, indicating a solid grasp on the underlying physics. These multiple lines of evidence provide *very high confidence* of enhanced Arctic warming with potentially significant impacts on coastal communities and marine ecosystems.

Key Finding 2

Rising Alaskan permafrost temperatures are causing permafrost to thaw and become more discontinuous; this process releases additional CO_2 and methane, resulting in an amplifying feedback and additional warming (*high confidence*). The overall magnitude of the permafrost–carbon feedback is uncertain; however, it is clear that these emissions have the potential to complicate the ability to meet policy goals for the reduction of greenhouse gas concentrations.

Description of evidence base

The Key Finding is supported by observational evidence of warming permafrost temperatures and a deepening active layer, in situ gas measurements and laboratory incubation experiments of CO_2 and CH_4 release, and model studies (Vaughan et al. 2013; Fisher et al. 2014; Schuur et al. 2015; Koven et al. 2015a,b; Schädel et al. 2016; Liljedahl et al. 2016). Alaska and Arctic permafrost characteristics have responded to increased temperatures and reduced snow cover in most regions since the 1980s, with colder permafrost warming faster than warmer permafrost (AMAP 2011; Vaughan et al. 2013; Romanovsky et al. 2016). Large carbon soil pools (more than 50% of the global below-ground organic carbon pool) are locked up in the permafrost soils (Tarnocai et al. 2009), with the potential to be released. Thawing permafrost makes previously frozen organic matter available for microbial decomposition. In situ gas flux measurements have directly measured the release of CO_2 and CH_4 from Arctic permafrost (Schuur et al. 2009; Zona et al. 2016). The specific conditions of microbial decomposition, aerobic or anaerobic, determines the relative production of CO_2 and CH_4. This distinction is significant as CH_4 is a

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much more powerful greenhouse gas than CO_2 (Myhre et al. 2013). However, incubation studies indicate that 3.4 times more carbon is released under aerobic conditions than anaerobic conditions, leading to a 2.3 times the stronger radiative forcing under aerobic conditions (Schädel et al. 2016). Combined data and modeling studies suggest a global sensitivity of the permafrost–carbon feedback warming global temperatures in 2100 by 0.52 ± 0.38°F (0.29 ± 0.21°C) alone (Schaefer et al. 2014). Chadburn et al. (2017) infer the sensitivity of permafrost area to globally averaged warming to be 4 million km^2 by constraining a group of climate models with the observed spatial distribution of permafrost; this sensitivity is 20% higher than previous studies. Permafrost thaw is occurring faster than models predict due to poorly understood deep soil, ice wedge, and thermokarst processes (Fisher et al. 2014; Koven et al. 2015a; Hollesen et al. 2015; Liljedahl et al. 2016). Additional uncertainty stems from the surprising uptake of methane from mineral soils (Oh et al. 2016) and dependence of emissions on vegetation and soil properties (Treat et al. 2015). The observational and modeling evidence supports the Key Finding that the permafrost–carbon cycle is positive.

Major uncertainties

A major limiting factor is the sparse observations of permafrost in Alaska and remote areas across the Arctic. Major uncertainties are related to deep soil, ice wedging, and thermokarst processes and the dependence of CO_2 and CH_4 uptake and production on vegetation and soil properties. Uncertainties also exist in relevant soil processes during and after permafrost thaw, especially those that control unfrozen soil carbon storage and plant carbon uptake and net ecosystem exchange. Many processes with the potential to drive rapid permafrost thaw (such as thermokarst) are not included in current earth system models.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

There is *high confidence* that permafrost is thawing, becoming discontinuous, and releasing CO_2 and CH_4. Physically-based arguments and observed of increases in CO_2 and CH_4 emissions as permafrost thaws indicate that the feedback is positive. This confidence level is justified based on observations of rapidly changing permafrost characteristics.

If appropriate, estimate likelihood of impact or consequence, including short description of basis of estimate

Thawing permafrost *very likely* has significant impacts to the global carbon cycle and serves as a source of CO_2 and CH_4 emission that complicates the ability to meet policy goals

Summary sentence or paragraph that integrates the above information

Permafrost is thawing, becoming more discontinuous, and releasing CO_2 and CH_4. Observational and modeling evidence indicates that permafrost has thawed and released additional CO_2 and

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CH_4 indicating that the permafrost–carbon cycle feedback is positive accounting for additional warming of approximately 0.08ºC to 0.50ºC on top of climate model projections. Although the magnitude of the permafrost–carbon feedback is uncertain due to a range of poorly understood processes (deep soil and ice wedge processes, plant carbon uptake, dependence of uptake and emissions on vegetation and soil type, and the role of rapid permafrost thaw processes, such as thermokarst), emerging science and the newest estimates continue to indicate that this feedback is more likely on the larger side of the range. Impacts of permafrost thaw and the permafrost carbon feedback complicates our ability to meet policy goals by adding a currently unconstrained radiative forcing to the climate system.

Key Finding 3

Arctic land and sea ice loss observed in the last three decades continues, in some cases accelerating (*very high confidence*). It is *virtually certain* that Alaska glaciers have lost mass over the last 50 years, with each year since 1984 showing an annual average ice mass less than the previous year. Based on gravitational data from satellites, average ice mass loss from Greenland was –269 Gt per year between April 2002 and April 2016, accelerating in recent years (*high confidence*). Since the early 1980s, annual average Arctic sea ice has decreased in extent between 3.5% and 4.1% per decade, become thinner by between 4.3 and 7.5 feet, and began melting at least 15 more days each year. September sea ice extent has decreased between 10.7% and 15.9% per decade (*very high confidence*). Arctic-wide ice loss is expected to continue through the 21st century, *very likely* resulting in nearly sea ice-free late summers by the 2040s (*very high confidence*).

Description of evidence base

The Key Finding is supported by observational evidence from multiple ground-based and satellite-based observational techniques (including passive microwave, laser and radar altimetry, and gravimetry) analyzed by independent groups using different techniques reaching similar conclusions (Vaughan et al. 2013; Comiso and Hall 2014; Stroeve et al. 2014a; Larsen et al. 2015; Zemp et al. 2015; Larsen et al. 2015; Harig and Simons 2016; Mengel et al. 2016; Perovich et al. 2016). Additionally, the U.S. Geological Survey repeat photography database shows the glacier retreat for many Alaskan glaciers (Figure 11.4: Muir Glacier). Several independent model analysis studies using a wide array of climate models and different analysis techniques indicate that sea ice loss will continue across the Arctic, *very likely* resulting in late summers becoming nearly ice-free by mid-century (Wang and Overland 2012; Collins et al. 2013; Snape and Forster 2014).

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Major uncertainties

Key uncertainties remain in the quantification and modeling of key physical processes that contribute to the acceleration of land and sea ice melting. Climate models are unable to capture the rapid pace of observed sea and land ice melt over the last 15 years; a major factor is our inability to quantify and accurately model the physical processes driving the accelerated melting. The interactions between atmospheric circulation, ice dynamics and thermodynamics, clouds, and specifically the influence on the surface energy budget are key uncertainties. Mechanisms controlling marine-terminating glacier dynamics—specifically the roles of atmospheric warming, seawater intrusions under floating ice shelves, and the penetration of surface meltwater to the glacier bed—are key uncertainties in projecting Greenland Ice Sheet melt.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

There is *very high confidence* that Arctic sea and land ice melt is accelerating and mountain glacier ice mass is declining given the multiple observational sources and analysis technique documented in the peer reviewed climate science literature.

If appropriate, estimate likelihood of impact or consequence, including short description of basis of estimate

It is *very likely* that accelerating Arctic land and sea ice melt impacts the United States. Accelerating Arctic Ocean sea ice melt increases coastal erosion in Alaska and makes Alaskan fisheries more susceptible to ocean acidification by changing Arctic Ocean chemistry. Greenland Ice Sheet and Alaska mountain glacier melt drives sea level rise threatening coastal communities in the United State and worldwide, influencing marine ecology, and potentially altering the thermohaline circulation.

Summary sentence or paragraph that integrates the above information

Arctic land and sea ice loss observed in the last three decades continues, in some cases accelerating. A diverse range of observational evidence from multiple data sources and independent analysis techniques provide consistent evidence of substantial declines in Arctic sea ice extent, thickness, and volume since at least 1979, mountain glacier melt over the last 50 years, and accelerating mass loss from Greenland. An array of different models and independent analyses indicate that future declines in ice across the Arctic are expected resulting in late summers in the Arctic becoming ice free by midcentury.

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Key Finding 4

It is *virtually certain* that human activities have contributed to Arctic surface temperature warming, sea ice loss since 1979, glacier mass loss, and northern hemisphere snow extent decline observed across the Arctic (*very high confidence*). Human activities have *likely* contributed to more than half of the observed Arctic surface temperature rise and September sea ice decline since 1979 (*high confidence*).

Description of evidence base

The Key Finding is supported by many attribution studies including a wide array of climate models documenting the anthropogenic influence on Arctic temperature, sea ice, mountain glaciers and snow extent (Vinnikov et al. 1999; Stroeve et al. 2007; Gillett et al. 2008; Min et al. 2008; Kay et al. 2011b; Day et al. 2012; Wang and Overland 2012; Bindoff et al. 2013; Christensen et al. 2013; Najafi et al. 2015). Observation-based analyses also support an anthropogenic influence (Notz and Marotzke 2012; Notz and Stroeve 2015). Emerging science indicates it is *very likely* that natural variability alone could not have caused the recently observed record low Arctic sea ice extents, such as in September 2012 (Zhang and Knutson 2013; Kirchmeyer-Young et al. 2017). Natural variability in the Arctic is significant (Swart et al. 2015; Jahn et al. 2016), however the majority of studies indicate that the contribution from internal variability to observed trends in Arctic temperature and sea ice are less than 50% (Kay et al. 2011b; Day et al. 2012; Ding et al. 2017), therefore human activities have *likely* contributed to more than half of the observed sea ice loss since 1979. Multiple lines evidence, independent analysis techniques, models, and studies support the Key Finding.

Major uncertainties

A major limiting factor in our ability to attribute Arctic sea ice and glacier melt to human activities is the significant natural climate variability in the Arctic. Longer data records and a better understanding of the physical mechanisms that drive natural climate variability in the Arctic are required to reduce this uncertainty. Another major uncertainty is the ability of climate models to capture the relevant physical processes and climate changes at a fine spatial scale, especially those at the land and ocean surface in the Arctic.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

There is *very high confidence* that human activities have contributed to Arctic surface temperature warming, sea ice loss since 1979, glacier mass loss, and northern hemisphere snow extent given multiple independent analysis techniques from independent groups using many different climate models indicate the same conclusion.

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If appropriate, estimate likelihood of impact or consequence, including short description of basis of estimate

Arctic sea ice and glacier mass loss impacts the United States by affecting coastal erosion in Alaska and key Alaskan fisheries through an increased vulnerability to ocean acidification. Glacier mass loss is a significant driver of sea level rise threatening coastal communities in the United States and worldwide, influencing marine ecology, and potentially altering the Atlantic Meridional Overturning Circulation (Liu et al. 2017).

Summary sentence or paragraph that integrates the above information

Evidenced by the multiple independent studies, analysis techniques, and the array of different climate models used over the last 20 years, it is *virtually certain* that human activities have contributed to Arctic surface temperature warming, sea ice loss since 1979, glacier mass loss, and Northern Hemisphere snow extent decline observed across the Arctic. Key uncertainties remain in the understanding and modeling of Arctic climate variability; however, the majority of studies indicate that contribution from internal variability to observed trends in Arctic temperature and sea ice are less than 50%. This suggests that it is also *likely* that human activities have contributed to more than half of the observed September sea ice decline since 1979.

Key Finding 5

Atmospheric circulation patterns connect the climates of the Arctic and the contiguous United States. Evidenced by recent record warm temperatures in the Arctic and emerging science, the midlatitude circulation has influenced observed Arctic temperatures and sea ice (*high confidence*). However, confidence is *low* regarding whether or by what mechanisms observed Arctic warming may have influenced the midlatitude circulation and weather patterns over the continental United States. The influence of Arctic changes on U.S. weather over the coming decades remains an open question with the potential for significant impact.

Description of evidence base

The midlatitude circulation influences the Arctic through the transport of warm, moist air, altering the Arctic surface energy budget (Rigor et al. 2002; Graverson et al. 2006; Screen et al. 2012; Perlwitz et al. 2015). The intrusion of warm, moist air from midlatitudes increases downwelling longwave radiation, warming the Arctic surface and hindering wintertime sea ice growth (Lee 2014; Liu and Key 2014). Emerging research provides a new understanding of the importance of synoptic time scales and the episodic nature of midlatitude air intrusions (Lee 2014; Park et al. 2015; Woods and Caballero 2016). The combination of recent observational and model-based evidence as well as the physical understanding of the mechanisms of midlatitude circulation effects on Arctic climate supports this Key Finding.

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In addition, research on the impact of Arctic climate on midlatitude circulation is rapidly evolving, including observational analysis and modeling studies. Multiple observational studies provide evidence for concurrent changes in the Arctic and Northern Hemisphere large-scale circulation changes (Cohen et al. 2014; Vihma 2014; Barnes and Screen 2015). Further, modeling studies demonstrate that Arctic warming can influence the midlatitude jet stream and storm track (Barnes and Screen 2015; Barnes and Polvani 2015; Overland et al. 2016; Francis et al. 2017). However, attribution studies indicate that the observed midlatitude circulation changes over the continental United States are smaller than natural variability and are therefore not detectable in the observational record (Screen et al. 2012; Perlwitz et al. 2015; Sigmond and Fyfe 2016; Sun et al. 2016). This disagreement between independent studies using different analysis techniques and the lack of understanding of the physical mechanism(s) support this Key Finding.

Major uncertainties

A major limiting factor is our understanding and modeling of natural climate variability in the Arctic. Longer data records and a better understanding of the physical mechanisms that drive natural climate variability in the Arctic are required to reduce this uncertainty. The inability of climate models to accurately capture interactions between sea ice and the atmospheric circulation and polar stratospheric processes limits our current understanding.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

High confidence in the impact of midlatitude circulation on Arctic changes from the consistency between observations and models as well as a solid physical understanding.

Low confidence on the detection of an impact of Arctic warming on midlatitude climate is based on short observational data record, model uncertainty, and lack of physical understanding.

Summary sentence or paragraph that integrates the above information

The midlatitude circulation has influenced observed Arctic temperatures, supported by recent observational and model-based evidence as well as the physical understanding from emerging science. In turn, confidence is low regarding the mechanisms by which observed Arctic warming has influenced the midlatitude circulation and weather patterns over the continental United States, due to the disagreement between numerous studies and a lack of understanding of the physical mechanism(s). Resolving the remaining questions requires longer data records and improved understanding and modeling of physics in the Arctic. The influence of Arctic changes on U.S. weather over the coming decades remains an open question with the potential for significant impact.

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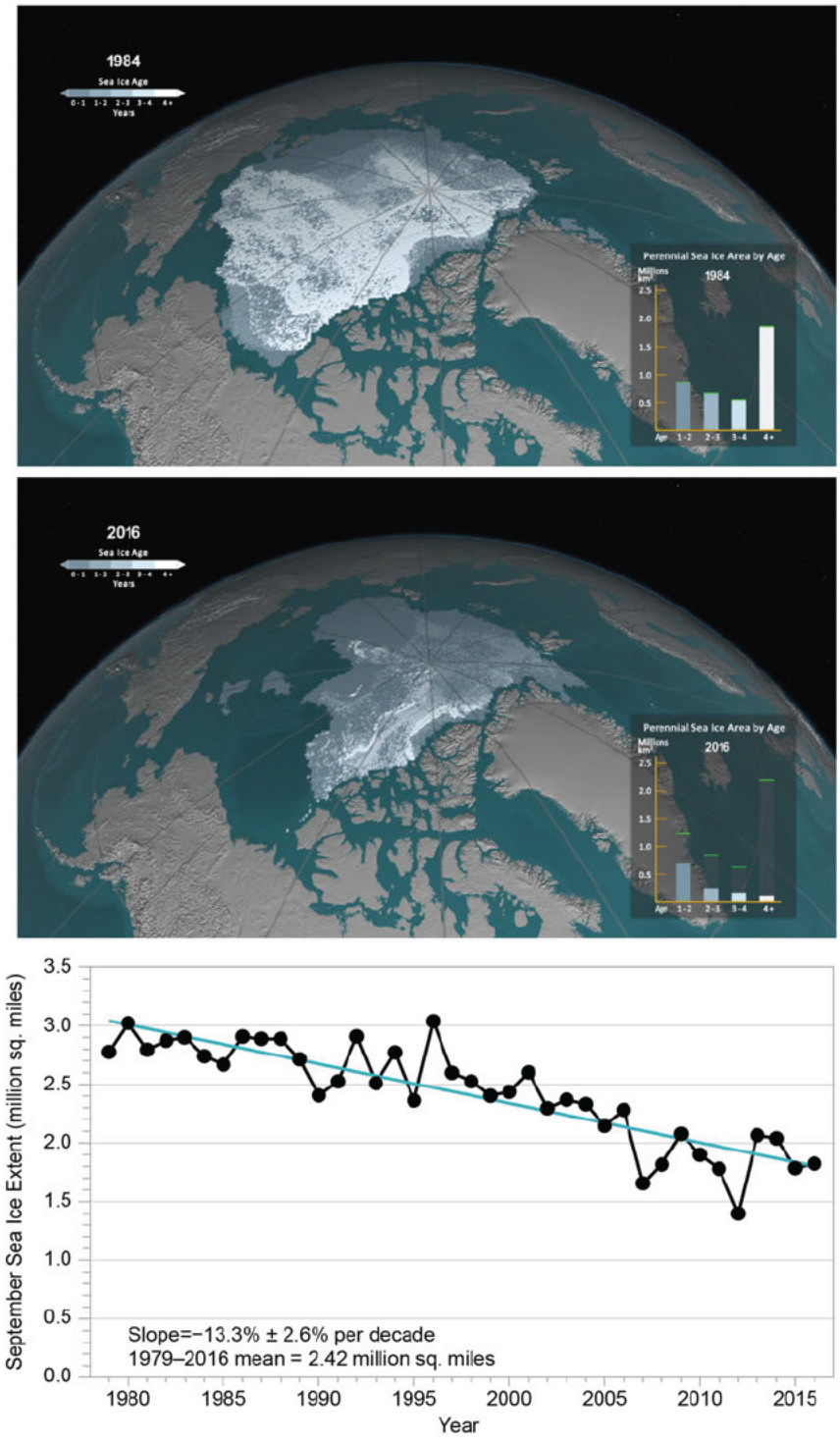
1 **FIGURES**



2

3 **Figure 11.1**: September sea ice extent and age shown for (a) 1984 and (b) 2016, illustrating
4 significant reductions in sea ice extent and age (thickness). Bar graph in the lower right of each
5 panel illustrates the sea ice area (unit: million km^2) covered within each age category (> 1 year),

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and the green bars represent the maximum value for each age range during the record. The year 1984 is representative of September sea ice characteristics during the 1980s. The years 1984 and 2016 are selected as endpoints in the time series; a movie of the complete time series is available at http://svs.gsfc.nasa.gov/cgi-bin/details.cgi?aid=4489. (c) Shows the satellite-era Arctic sea ice areal extent trend from 1979 to 2016 for September (unit: million mi^2). (Figure source: Panel (a,b): NASA Science Visualization Studio; data: Tschudi et al. 2016; Panel (c) data: Fetterer et al. 2016).

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