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13. Ocean Acidification and Other Ocean Changes

KEY FINDINGS

1. The world's oceans have absorbed about 93% of the excess heat caused by greenhouse gas warming since the mid-20th century, making them warmer and altering global and regional climate feedbacks. Ocean heat content has increased at all depths since the 1960s and surface waters have warmed by about 1.3° ± 0.1°F (0.7° ± 0.08°C) per century globally since 1900 to 2016. Under a high emissions scenario, a global increase in average sea surface temperature of 4.9° ± 1.3°F (2.7° ± 0.7°C) by 2100 is projected, with even higher changes in some U.S. coastal regions. (*Very high confidence*)

2. The potential slowing of the Atlantic Meridional Overturning Circulation (AMOC) (of which the Gulf Stream is one component)—as a result of increasing ocean heat content and freshwater driven buoyancy changes—could have dramatic climate feedbacks as the ocean absorbs less heat and CO_2 from the atmosphere. This slowing would also affect the climates of North America and Europe. Any slowing documented to date cannot be directly tied to anthropogenic forcing primarily due to lack of adequate observational data and to challenges in modeling ocean circulation changes. Under a high emissions scenario (RCP8.5) in CMIP5 simulations, it is likely that the AMOC will weaken over the 21st century by 12% to 54%. (*Low confidence*)

3. The world's oceans are currently absorbing more than a quarter of the CO_2 emitted to the atmosphere annually from human activities, making them more acidic (*very high confidence*), with potential detrimental impacts to marine ecosystems. In particular, higher-latitude systems typically have a lower buffering capacity against pH change, exhibiting seasonally corrosive conditions sooner than low-latitude systems. Acidification is regionally increasing along U.S. coastal systems as a result of upwelling (for example, in the Pacific Northwest) (*high confidence*), changes in freshwater inputs (for example, in the Gulf of Maine) (*medium confidence*), and nutrient input (for example, in urbanized estuaries) (*high confidence*). The rate of acidification is unparalleled in at least the past 66 million years (*medium confidence*). Under RCP8.5, the global average surface ocean acidity is projected to increase by 100% to 150% (*high confidence*).

4. Increasing sea surface temperatures, rising sea levels, and changing patterns of precipitation, winds, nutrients, and ocean circulation are contributing to overall declining oxygen concentrations at intermediate depths in various ocean locations and in many coastal areas. Over the last half century, major oxygen losses have occurred in inland seas, estuaries, and in the coastal and open ocean (*high confidence*). Ocean oxygen levels are projected to decrease by as much as 3.5% under the RCP8.5 scenario by 2100 relative to preindustrial values (*high confidence*).

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13.0 A Changing Ocean

Anthropogenic perturbations to the global Earth system have included important alterations in the nutrient composition, temperature, and circulation of the oceans. Some of these changes will be distinguishable from the background natural variability in nearly half of the global open ocean within a decade, with important consequences for marine ecosystems and their services (Gattuso et al. 2015). However, the timeframe for detection will vary depending on the parameter featured (Henson et al 2010; Henson et al 2016).

13.1 Ocean Warming

13.1.1 General Background

Approximately 93% of excess heat energy trapped since the 1970s has been absorbed into the oceans, lessening atmospheric warming and leading to a variety of changes in ocean conditions, including sea level rise and ocean circulation (see Ch. 2: Physical Drivers of Climate Change, Ch. 6: Temperature Change, and Ch. 12: Sea Level Rise in this report; Rhein et al. 2013; Gattuso et al. 2015). This is the result of the high heat capacity of seawater relative to the atmosphere, the relative area of the ocean compared to the land, and the ocean circulation that enables the transport of heat into deep waters. This large heat absorption by the oceans moderates the effects of increased anthropogenic greenhouse emissions on terrestrial climates while altering the fundamental physical properties of the ocean and indirectly impacting chemical properties such as the biological pump through increased stratification (Gattuso et al. 2015; Rossby 1959). Although upper ocean temperature varies over short- and medium timescales (for example, seasonal and regional patterns), there are clear long-term increases in surface temperature and ocean heat content over the past 65 years (Cheng et al. 2017; Rhein et al. 2013; Levitus et al. 2012).

13.1.2 Ocean Heat Content

Ocean heat content (OHC) is an ideal variable to monitor changing climate as it is calculated using the entire water column, so ocean warming can be documented and compared between particular regions, ocean basins, and depths. However, for years prior to the 1970s, estimates of ocean uptake are confined to the upper ocean due to sparse spatial and temporal coverage and limited vertical capabilities of many of the instruments in use. Ocean heat content estimates are improved for time periods after 1970 with increased sampling coverage and depth (Abraham et al. 2013; Rhein et al. 2013). Estimates of OHC have been calculated going back to the 1950s using averages over longer time intervals (i.e., decadal or 5-year intervals) to compensate for sparse data distributions, allowing for clear long-term trends to emerge (e.g., Levitus et al. 2012).

From 1960 to 2015, ocean heat content (OHC) significantly increased for both 0–700 and 700–2,000 m depths, for a total ocean warming of $33.5 \pm 7.0 \times 10^{22}$ J (a net heating of 0.37 ± 0.08 W/m^2), although there is some uncertainty with global ocean heat estimates (Figure 13.1; Cheng

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et al. 2017). During this period, there is evidence of an acceleration of ocean warming beginning in 1998 (Lee et al. 2015), with a total heat increase of about 15.2×10^{22} J (Cheng et al. 2017). Robust ocean warming occurs in the upper 700 m and is slow to penetrate into the deep ocean. However, the 700–2,000 m depths constitute an increasing portion of the total ocean energy budget as compared to the surface ocean (Figure 13.1; Cheng et al. 2017). The role of the deep ocean (below 2,000 m [6,600 ft]) in ocean heat uptake remains uncertain, both in the magnitude but also the sign of the uptake (Purkey and Johnson 2010; Llovel et al. 2014). Penetration of surface waters to the deep ocean is a slow process, which means that while it takes only about a decade for near-surface temperatures to respond to increased heat energy, the deep ocean will continue to warm, and as a result sea levels will rise for centuries to millennia even if all further emissions cease (Rhein et al. 2013).

[INSERT FIGURE 13.1 HERE]

Several sources have documented warming in all ocean basins from 0–2,000 m depths over the past 50 years (Figure 13.2; Boyer et al. 2016; Cheng et al. 2017; Levitus et al. 2012). Annual fluctuations in surface temperatures and OHC are attributed to the combination of a long-term secular trend and decadal and smaller time scale variations, such as the Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO) (Ch. 5: Large-Scale Circulation and Climate Variability; Ch. 12: Sea Level Rise; Trenberth et al. 2014; Steinman et al. 2015). The transport of heat to the deep ocean is likely linked to the strength of the Atlantic Meridional Overturning Circulation (see Section 13.2.1), where the Atlantic and Southern Ocean accounts for the dominant portion of total OHC at the 700–2,000 m depth (Figure 13.2; Cheng et al. 2017; Lee et al. 2015; Roemmich et al. 2015; Abraham et al. 2013). Decadal variability in ocean heat uptake is mostly attributed to ENSO phases (with El Niños warming and La Niñas cooling). For instance, La Niña conditions over the past decade have led to colder ocean temperatures in the eastern tropical Pacific (Abraham et al. 2013; Cheng et al. 2017; Lee et al. 2015; Kosaka and Xie 2013). For the Pacific and Indian Oceans, the decadal shifts are primarily observed in the upper 350 m depth, likely due to shallow subtropical circulation, leading to an abrupt increase of OHC in the Indian Ocean carried by the Indonesian throughflow from the Pacific Ocean over the last decade (Lee et al. 2015). Although there is natural variability in ocean temperature, there remain clear increasing trends due to anthropogenic influences.

[INSERT FIGURE 13.2 HERE]

13.1.3 Sea Surface Temperature and U.S. Regional Warming

In addition to OHC, sea surface temperature (SST) measurements are widely available. SST measurements are useful because 1) the measurements have been taken over 150 years (albeit using different platforms, instruments, and depths through time); 2) SST reflects the lower boundary condition of the atmosphere; and 3) SST can be used to predict specific regional impacts of global warming on terrestrial and coastal systems (Roemmich et al. 2015; Yan et al.

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2016; Matthews 2013). Globally, surface ocean temperatures have increased by 1.3° ± 0.1°F (0.70° ± 0.08°C) per century from 1900 to 2016 for the Extended Reconstructed Sea Surface Temperature version 4 (ERSST v4) record (Huang et al. 2015). All U.S. coastal waters have warmed by more than 0.7°F (0.4°C) over this period as shown in both Table 13.1 and Chapter 6: Temperature Change, Figure 6.6. During the past century, the rates of increase of SSTs for the coastal waters of three U.S. regions were above the global average rate. These included the waters around Alaska, the Caribbean, and the Southwest (Table 13.1). Over the last decade, some regions have experienced increased high ocean temperature anomalies. For instance, due to a resilient ridge of high pressure over the North American west coast, storm activity and mixing was suppressed and heat in the upper ocean intensified in 2013 in a phenomenon known as "The Blob" (Bond et al. 2015). These anomalously warm waters persisted in the coastal waters of the Alaskan and Pacific Northwest until 2015. Under a higher emissions pathway (RCP8.5), ocean SST are projected to increase by an additional 4.9°F (2.7°C) by 2100 (Figure 13.3), whereas for a lower emissions scenario (RCP4.5) the SST increase would be 2.3°F (1.3°C; Bopp et al. 2013). In all U.S. coastal regions, the warming since 1901 is detectable compared to natural variability and attributable to anthropogenic forcing, according to an analysis of the CMIP5 models (Ch. 6: Temperature Change, Figure 6.5).

[INSERT TABLE 13.1 AND FIGURE 13.3 HERE]

13.1.4 Ocean Heat Feedback

The residual heat not taken up by the oceans increases land surface temperatures (approximately 3%) and atmospheric temperatures (approximately 1%), and melts both land and sea ice (approximately 3%), leading to sea level rise (see Ch. 12: Sea Level Rise; Cheng et al. 2017; Rhein et al. 2013; Nieves et al. 2015). The meltwater from land and sea ice amplifies further subsurface ocean warming and ice shelf melting, primarily due to increased thermal stratification, which reduces the ocean's efficiency in transporting heat to deep waters (Rhein et al. 2013). Surface ocean stratification has increased by about 4% during the period 1971–2010 (Ciais et al. 2013) due to thermal heating and freshening from increased freshwater inputs (precipitation and evaporation changes and land and sea ice melting). The increase of ocean stratification will contribute to further feedback of ocean warming and, indirectly, mean sea level. In addition, increases in stratification are associated with suppression of tropical cyclone intensification (Mei et al. 2015), retreat of the polar ice sheets (Straneo and Heimbach 2013), and reductions of the convective mixing at higher latitudes that transports heat to the deep ocean through the Atlantic Meridional Overturning Circulation (AMOC) (Rahmstorf et al. 2015). Ocean heat uptake therefore represents an important feedback that will have a significant influence on future shifts in climate (see Ch. 2: Physical Drivers of Climate Change).

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13.2 Ocean Circulation

13.2.1 Atlantic Meridional Overturning Circulation

The Atlantic Meridional Overturning Circulation (AMOC) refers to the three-dimensional, time-dependent circulation of the Atlantic Ocean, which has been a high priority topic of study in recent decades. The AMOC plays an important role in climate through its transport of heat, freshwater, and carbon (e.g., Johns et al. 2011; McDonagh et al. 2015; Talley et al. 2016). AMOC-associated poleward heat transport substantially contributes to North American and continental European climate (see Ch. 5: Circulation and Variability). The Gulf Stream, in contrast to other western boundary currents, is expected to slow down because of the weakening of the AMOC, which would impact the European climate (Yang et al. 2016). Variability in the AMOC has been attributed to wind forcing on intra-annual time scales and to geostrophic forces on interannual to decadal timescales (Buckley and Marshal 2016). Increased freshwater fluxes from melting arctic sea and land ice can weaken open ocean convection and deep-water formation in the Labrador and Irminger Seas, which could weaken the AMOC (Ch. 11: Arctic Changes; Rahmstorf et al. 2015; Yang et al. 2016; Also see Chapter 5, Section 5.2.3: North Atlantic Oscillation and Northern Annular Mode).

While one recent study has suggested that the AMOC has slowed since preindustrial times (Rahmstorf et al. 2015) and another suggested slowing on faster time scales (Smeed et al. 2014), there is at present insufficient observational evidence to support a finding of long term slowdown of AMOC strength over the 20th century (Rhein et al. 2013) or within the last 50 years (Buckley and Marshall 2016) as decadal ocean variability can obscure long-term trends. Some studies show long-term trends (Longworth et al. 2011; Bryden et al. 2005), but the combination of sparse data and large seasonal variability may also lead to incorrect interpretations (e.g., Kanzow et al. 2010). Several recent high resolution modeling studies constrained with the limited existing observational data (Jackson et al. 2016) and/or with reconstructed freshwater fluxes (Böning et al. 2016) suggest that the recently observed AMOC slowdown at 26°N (off the Florida coast) since 2004 (e.g., as described in Smeed et al. 2014) is mainly due to natural variability, and that anthropogenic forcing has not yet caused a significant AMOC slowdown. In addition, direct observations of the AMOC in the South Atlantic fail to unambiguously demonstrate anthropogenic trends (e.g., Dong et al. 2015; Garzoli et al. 2013).

Under a high emissions future scenario (RCP8.5) in CMIP5 simulations, it is very likely that the AMOC will weaken over the 21st century. The projected decline ranges from 12% to 54% (Collins et al. 2013), with the range width reflecting substantial uncertainty in quantitative projections of AMOC behavior. In RCP4.5 scenarios, CMIP5 models predict a 20% weakening of the AMOC during the first half of the 21st century and a stabilization and slight recovery after that (Cheng et al. 2013). The projected slowdown of the AMOC will be counteracted by the warming of the deep ocean (below 700 m [2,300 ft]), which will tend to strengthen the AMOC (Patara and Böning 2014). The situation is further complicated due to the known bias in coupled

climate models related to the direction of the salinity transport in models versus observations, which is an indicator of AMOC stability (e.g., Drijhout et al. 2011; Bryden et al. 2011; Garzoli et al. 2013). Some argue that coupled climate models should be corrected for this known bias and that AMOC variations could be even larger than the gradual decrease most models predict if the AMOC were to shut down completely and "flip states" (Liu et al. 2017). Any AMOC slowdown will result in less heat and CO_2 absorbed by the ocean from the atmosphere, which is a positive feedback to climate change (also see Ch. 2: Physical Drivers of Climate Change).

13.2.2 Changes in Salinity Structure

As a response to warming, increased atmospheric moisture leads to stronger evaporation or precipitation in terrestrial and oceanic environments and melting of land and sea ice. Approximately 80% of precipitation/evaporation events occur over the ocean, leading to patterns of higher salt content or freshwater anomalies and changes in ocean circulation (see Ch. 2: Physical Drivers of Climate Change and Ch. 6: Temperature Change; Durack and Wijffels 2010). Over 1950–2010, average global amplification of the surface salinity pattern amounted to 5.3%; where fresh regions in the ocean became fresher and salty regions became saltier (Skliris et al. 2014). However, the long-term trends of these physical and chemical changes to the ocean are difficult to isolate from natural large-scale variability. In particular, ENSO displays particular salinity and precipitation/evaporation patterns that skew the trends. More research and data are necessary to better model changes to ocean salinity. Several models have shown a similar spatial structure of surface salinity changes, including general salinity increases in the subtropical gyres, a strong basin-wide salinity increase in the Atlantic Ocean, and reduced salinity in the western Pacific warm pools and the North Pacific subpolar regions (Durack and Wijffels 2010; Skliris et al. 2014). There is also a stronger distinction between the upper salty thermocline and fresh intermediate depth through the century. The regional changes in salinity to ocean basins will have an overall impact on ocean circulation and net primary production, leading to corresponding carbon export (see Ch. 2: Physical Drivers of Climate Change). In particular, the freshening of the Arctic Ocean due to melting of land and sea ice can lead to buoyancy changes which could slow down the AMOC (see Section 13.2.1).

13.2.3 Changes in Upwelling

Significant changes to ocean stratification and circulation can also be observed regionally, along the eastern ocean boundaries and at the equator. In these areas, wind-driven upwelling brings colder, nutrient- and carbon-rich water to the surface; this upwelled water is more efficient in heat and CO_2 uptake. There is some evidence that coastal upwelling in mid- to high-latitude eastern boundary regions has increased in intensity and/or frequency (García-Reyes et al. 2015), but in more tropical areas of the western Atlantic, such as in the Caribbean Sea, it has decreased between 1990 and 2010 (Taylor et al. 2012; Astor et al. 2013). This has led to a decrease in primary productivity in the southern Caribbean Sea (Taylor et al. 2012). Within the continental United States, the California Current is experiencing fewer (by about 23%–40%) but stronger

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upwelling events (Hoegh-Guldberg et al. 2014; Sydeman et al. 2014; Jacox et al. 2014). Stronger offshore upwelling combined with cross-shelf advection brings nutrients from the deeper ocean but also increased offshore transport (Bakun et al. 2015). The net nutrient load in the coastal regions is responsible for increased productivity and ecosystem function.

IPCC 2013 concluded that there is low confidence in the current understanding of how eastern upwelling systems will be altered under future climate change because of the obscuring role of multidecadal climate variability (Ciais et al. 2013). However, subsequent studies show that by 2100, upwelling is predicted to start earlier, end later, and intensify in three of the four major eastern boundary upwelling systems (not in the California Current; Wang et al. 2015). Southern Ocean upwelling will intensify while the Atlantic equatorial upwelling systems will weaken (Hoegh-Guldberg et al. 2014; Wang et al. 2015). The intensification is attributed to the strengthening of regional coastal winds as observations already show (Sydeman et al. 2014), and RCP8.5 model projection scenarios estimate wind intensifying near poleward boundaries (including northern California current system [CCS]) and weakening near equatorward boundaries (including southern CCS) for the 21st century (Rykaczewski et al. 2015; Wang et al. 2015).

13.3 Ocean Acidification

13.3.1 General Background

In addition to causing changes in climate, increasing atmospheric levels of carbon dioxide (CO_2) from the burning of fossil fuels and other human activities, including changes in land use, have a direct effect on ocean chemistry (Orr et al. 2005; Feely et al. 2009). Ocean acidification refers to a change in ocean chemistry in response to the uptake of increasing CO_2 in the atmosphere. Ocean acidification causes a variety of chemical changes in seawater: an increase in the partial pressure of CO_2 (pCO_2,sw), dissolved inorganic carbon (DIC), and an increase in the concentration of hydrogen and bicarbonate ions and a decrease in the concentration of carbonate ions (Figure 13.4). In brief, CO_2 is an acid gas that combines with water to form carbonic acid, which then dissociates to hydrogen and bicarbonate ions. Increasing concentrations of seawater hydrogen ions result in a decrease of carbonate ions through their conversion to bicarbonate ions. Ocean acidity refers to the concentration of hydrogen ions in ocean seawater regardless of ocean pH, which is fundamentally basic (e.g., pH > 7). Ocean surface waters have become 30% more acidic over the last 150 years as they have absorbed large amounts of CO_2 from the atmosphere (Feely et al. 2004), and anthropogenically sourced CO_2 is gradually invading into oceanic deep waters. Since the preindustrial period, the oceans have absorbed approximately 27% of all CO_2 emitted to the atmosphere. Oceans currently absorb about 26% of the human-caused CO_2 anthropogenically emitted into the atmosphere (Le Quéré et al. 2016).

[INSERT FIGURE 13.4 HERE]

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13.3.2 Open Ocean Acidification

Surface waters in the open ocean experience changes in carbonate chemistry reflective of large-scale physical oceanic processes (see Ch. 2: Physical Drivers of Climate Change). These processes include both the global uptake of atmospheric CO_2 and the shoaling of naturally acidified subsurface waters due to vertical mixing and upwelling. In general, the rate of ocean acidification in open ocean surface waters at a decadal time-scale closely approximates the rate of atmospheric CO_2 increase (Bates et al. 2014). Large, multidecadal phenomena such as the Atlantic Multidecadal Oscillation and Pacific Decadal Oscillation can add variability to the observed rate of change (Bates et al. 2014).

13.3.3 Coastal Acidification

Coastal shelf and nearshore waters are influenced by the same processes as open ocean surface waters such as absorption of atmospheric CO_2 and upwelling, as well as a number of additional, local-level processes, including freshwater and nutrient input (Duarte et al. 2013). Coastal acidification generally exhibits higher-frequency variability and short-term episodic events relative to open-ocean acidification (Borges and Gypens 2010; Waldbusser and Salisbury 2014; Hendriks et al. 2015; Sutton et al. 2016). Upwelling is of particular importance in coastal waters, especially along the Pacific Coast. Deep waters that shoal with upwelling are enriched in CO_2 due to uptake of anthropogenic atmospheric CO_2 when last in contact with the atmosphere, coupled with deep water respiration processes and lack of gas exchange with the atmosphere (Feely et al. 2009; Harris et al. 2013). Freshwater inputs to coastal waters change seawater chemistry in ways that make it more susceptible to acidification, largely by freshening ocean waters and contributing varying amounts of dissolved inorganic carbon (DIC), total alkalinity (TA), dissolved and particulate organic carbon, and nutrients from riverine and estuarine sources. Coastal waters of the East Coast and mid-Atlantic are far more influenced by freshwater inputs than are Pacific Coast waters (Gledhill et al. 2015). Coastal waters can episodically experience riverine and glacial melt plumes that create conditions in which seawater can dissolve calcium carbonate structures (Evans et al. 2014; Salisbury et al. 2008). While these processes have persisted historically, climate-induced increases in glacial melt and high-intensity precipitation events can yield larger freshwater plumes than have occurred in the past. Nutrient runoff can increase coastal acidification by creating conditions that enhance biological respiration. In brief, nutrient loading typically promotes phytoplankton blooms, which, when they die, are consumed by bacteria. Bacteria respire CO_2 and thus bacterial blooms can result in acidification events whose intensity depends on local hydrographic conditions, including water column stratification and residence time (Waldbusser and Salisbury 2014). Long-term changes in nutrient loading, precipitation, and/or ice melt may also impart long-term, secular changes in the magnitude of coastal acidification.

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13.3.4 Latitudinal Variation

Ocean carbon chemistry is highly influenced by water temperature, largely because the solubility of CO_2 in seawater increases as water temperature declines. Thus, cold, high-latitude waters can absorb more CO_2 than warm, lower-latitude waters (Gledhill et al. 2015; Bates and Mathis 2009). Because carbonate minerals also more readily dissolve in colder waters, these waters can more regularly become undersaturated with respect to calcium carbonate whereby mineral dissolution is energetically favored. This chemical state, often referred to as seawater being "corrosive" to calcium carbonate, is important when considering the ecological implications of ocean acidification as many species make structures such as shells and skeletons from calcium carbonate. Some high-latitude waters already experience such corrosive conditions, which are rarely documented in low-latitude systems. For example, corrosive conditions have been documented in the Arctic and northeastern Pacific Oceans (Bates and Mathis 2009; Feely et al. 2008; Qi et al. 2017; Sutton et al. 2016). It is important to note that low-latitude waters are experiencing a greater absolute rate of change in calcium carbonate saturation state than higher latitudes, though these low-latitude waters are not approaching the undersaturated state except within near-shore or some benthic habitats (Friedrich et al. 2012).

13.3.5 Paleo Evidence

Evidence suggests that the current rate of ocean acidification is the fastest in the last 66 million years (the K-Pg boundary) and possibly even the last 300 million years (when the first pelagic calcifiers evolved providing proxy information and also a strong carbonate buffer, characteristic of the modern ocean) (Hönisch et al. 2012; Zeebe et al. 2016). The Paleo-Eocene Thermal Maximum (PETM; around 56 million years ago) is often referenced as the closest analogue to the present, although the overall rate of change in CO_2 conditions during that event (estimated between 0.6 and 1.1 GtC/year) was much lower than the current increase in atmospheric CO_2 of 10 GtC/year (Wright and Schaller 2013; Zeebe et al. 2016). The relatively slower rate of atmospheric CO_2 increase at the PETM likely led to relatively small changes in carbonate ion concentration in seawater compared with the contemporary acidification rate, due to the ability of rock weathering to buffer the change over the longer time period (Zeebe et al. 2016). Some of the presumed acidification events in Earth's history have been linked to selective extinction events suggestive of how guilds of species may respond to the current acidification event (Hönisch et al. 2012).

13.3.6 Projected Changes

Projections indicate that by the end of the century under higher emissions pathways, such as SRES A1fI or RCP8.5, open-ocean surface pH will decline from the current average level of 8.1 to a possible average of 7.8 (Figure 13.5; Gattuso et al. 2015). When the entire ocean volume is considered under the same scenario, the volume of waters undersaturated with respect to calcium carbonate could expand from 76% in the 1990s to 91% in 2100. As discussed above, for a

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variety of reasons, not all ocean and coastal regions will experience acidification in the same way depending on other compounding factors. For instance, recent observational data from the Arctic Basin show that the Beaufort Sea became undersaturated, for part of the year, with respect to aragonite in 2001, while other continental shelf seas in the Arctic Basin are projected to do so closer to the middle of the century (e.g., the Chukchi Sea in about 2033 and Bering Sea in about 2062; Mathis et al. 2015). Deviation from the global average rate of acidification will be especially true in coastal and estuarine areas where the rate of acidification is influenced by other drivers than atmospheric CO_2, some of which are under the control of local management decisions (for example, nutrient pollution loads).

[INSERT FIGURE 13.5 HERE]

13.4 Ocean Deoxygenation

13.4.1 General Background

Oxygen is essential to most life in the ocean, governing a host of biogeochemical and biological processes. Oxygen influences metabolic, physiological, reproductive, behavioral, and ecological processes, ultimately shaping the composition, diversity, abundance, and distribution of organisms from microbes to whales. Increasingly, climate-induced oxygen loss (deoxygenation) associated with ocean warming and reduced ventilation to deep waters has become evident locally, regionally, and globally. Deoxygenation can also be attributed to anthropogenic nutrient input, especially in the coastal regions, where the nutrients can lead to the proliferation of primary production and, consequently, enhanced drawdown of dissolved oxygen by microbes (Altieri and Gedan 2015). In addition, acidification (Section 13.2) can co-occur with deoxygenation as a result of warming-enhanced biological respiration (Breitburg et al. 2015). As aerobic organisms respire, O_2 is consumed and CO_2 is produced. Understanding the combined effect of both low O_2 and low pH on marine ecosystems is an area of active research (Gobler et al. 2014). Warming also raises biological metabolic rates which, in combination with intensified coastal and estuarine stratification, exacerbates eutrophication-induced hypoxia. We now see earlier onset and longer periods of seasonal hypoxia in many eutrophic sites, most of which occur in areas that are also warming (Altieri and Gedan 2015).

13.4.2 Climate Drivers of Ocean Deoxygenation

Global ocean deoxygenation is a direct effect of warming. Ocean warming reduces the solubility of oxygen (that is, warmer water can hold less oxygen) and changes physical mixing (for example, upwelling and circulation) of oxygen in the oceans. The increased temperature of global oceans accounts for about 15% of current global oxygen loss (Helm et al. 2011), although changes in temperature and oxygen are not uniform throughout the ocean (Roemmich et al. 2015). Warming also exerts direct influence on thermal stratification and enhances salinity stratification through ice melt and climate change-associated precipitation effects. Intensified

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stratification leads to reduced ventilation (mixing of oxygen into the ocean interior) and accounts for up to 85% of global ocean oxygen loss (Helm et al. 2011). Effects of ocean temperature change and stratification on oxygen loss are strongest in intermediate or mode waters at bathyal depths (in general, 200–3,000 m) and also nearshore and in the open ocean; these changes are especially evident in tropical and subtropical waters globally, in the Eastern Pacific (Stramma et al. 2010), and in the Southern Ocean (Helm et al. 2011).

There are also other, less direct effects of global temperature increase. Warming on land reduces terrestrial plant water efficiency (through effects on stomata; see Ch. 8: Drought, Floods, and Wildfires, Key Message 3), leading to greater runoff, on average, into coastal zones (see Ch. 8: Drought, Floods, and Wildfires for other hydrological effects of warming) and further enhancing hypoxia potential because greater runoff means more nutrient transport (See Ch. 2: Physical Drivers of Climate Change; Reay et al. 2008; Rabalais et al. 2009). Estuaries, especially ones with minimal tidal mixing, are particularly vulnerable to oxygen-depleted dead zones from the enhanced runoff and stratification. Warming can induce dissociation of frozen methane in gas hydrates buried on continental margins, leading to further drawdown of oxygen through aerobic methane oxidation in the water column (Boetius and Wenzhöfer 2013). On eastern ocean boundaries, warming can enhance the land–sea temperature differential, causing increased upwelling due to higher winds with (a) greater nutrient input leading to production, sinking, decay, and biochemical drawdown of oxygen and (b) upwelling of naturally low-oxygen, high-CO_2 waters onto the upper slope and shelf environments (Sydeman et al. 2014; Feely et al. 2009). However, in the California Current Ecosystem, upwelling intensification has occurred only in the poleward regions (north of San Francisco), and the drivers may not be associated with land–sea temperature differences (Rykaczewski et al. 2015). Taken together, the effects of warming are manifested as low-oxygen water in open oceans are transported to and upwelled along coastal regions. These low-oxygen upwelled waters are then coupled with eutrophication-induced hypoxia, further reducing oxygen content in coastal areas.

Changes in precipitation, winds, circulation, airborne nutrients, and sea level can also contribute to ocean deoxygenation. Projected increases in precipitation in some regions will intensify stratification, reducing vertical mixing and ventilation, and intensify nutrient input to coastal waters through excess runoff, which leads to increased algal biomass and concurrent dissolved oxygen consumption via community respiration (Lee et al. 2016). Coastal wetlands that might remove these nutrients before they reach the ocean may be lost through rising sea level, further exacerbating hypoxia (Rabalais et al. 2009). Some observations of oxygen decline are linked to regional changes in circulation involving low-oxygen water masses. Enhanced fluxes of airborne iron and nitrogen are interacting with natural climate variability and contributing to fertilization, enhanced respiration, and oxygen loss in the tropical Pacific (Ito et al. 2016). In contrast to the many sources of climate-induced oxygen loss, the projected increase in incidence and intensity of cyclones and hurricanes will induce mixing, which can ameliorate hypoxia locally (Rabalais et al. 2009).

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13.4.3 Biogeochemical Feedbacks of Deoxygenation to Climate and Elemental Cycles

Climate patterns and ocean circulation have a large effect on global nitrogen and oxygen cycles, which in turn affect phosphorus and trace metal availability and generate feedbacks to the atmosphere and oceanic production. Global ocean productivity may be affected by climate-driven changes below the tropical and subtropical thermocline which control the volume of suboxic waters (< 5 micromolar O_2), and consequently the loss of fixed nitrogen through denitrification (Codispoti et al. 2001; Deutsch et al. 2011). The extent of suboxia in the open ocean also regulates the production of the greenhouse gas nitrous oxide (N_2O); as oxygen declines, greater N_2O production may intensify global warming, as N_2O is about 310 times more effective at trapping heat than CO_2 (see Ch. 2: Physical Drivers of Climate Change, Section 2.3.2; Gruber 2008; EPA 2017). Production of hydrogen sulfide (H_2S, which is highly toxic) and intensified phosphorus recycling can occur at low oxygen levels (Wallmann 2003). Other feedbacks may emerge as oxygen minimum zone (OMZ) shoaling diminishes the depths of diurnal vertical migrations by fish and invertebrates, and as their huge biomass and associated oxygen consumption deplete oxygen (Bianchi et al. 2013).

13.4.4 Past Trends

Over hundreds of millions of years, oxygen has varied dramatically in the atmosphere and ocean and has been linked to biodiversity gains and losses (Knoll and Carroll 1999; McFall-Ngai et al. 2013). Variation in oxygenation in the paleo record is very sensitive to climate—with clear links to temperature and often CO_2 variation (Falkowski et al. 2011). OMZs expand and contract in synchrony with warming and cooling events, respectively (Robinson et al. 2007). Episodic climate events that involve rapid temperature increases over decades, followed by a cool period lasting a few hundred years, lead to major fluctuations in the intensity of Pacific and Indian Ocean OMZs (i.e., DO of < 20 μM). These events are associated with rapid variations in North Atlantic deep water formation (Schmittner et al. 2007). Ocean oxygen fluctuates on glacial-interglacial timescales of thousands of years in the Eastern Pacific (Galbraith et al. 2004; Moffitt et al. 2015).

13.4.5 Modern Observations (last 50+ years)

Long-term oxygen records made over the last 50 years reflect oxygen declines in inland seas (Justić et al. 1987; Zaitzev 1992; Conley et al. 2011), in estuaries (Brush 2009; Gilbert et al. 2005), and in coastal waters (Rabalais et al. 2007, 2010; Booth et al. 2012; Baden et al. 1990). The number of coastal, eutrophication-induced hypoxic sites in the United States has grown dramatically over the past 40 years (Diaz and Rosenberg 2008). Over larger scales, global syntheses show hypoxic waters have expanded by 4.5 million km^2 at a depth of 200 m (Stramma et al. 2010), with widespread loss of oxygen in the Southern Ocean (Helm et al. 2011), Western Pacific (Takatani et al. 2012), and North Atlantic (Stendardo and Gruber 2012). Overall oxygen declines have been greater in coastal ocean than in the open ocean (Gilbert et al. 2010) and often

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greater inshore than offshore (Bograd et al. 2015). The emergence of a deoxygenation signal in regions with naturally high oxygen variability will unfold over longer time periods (20–50 years from now) (Long et al. 2016).

13.4.6 Projected Changes

GLOBAL MODELS

Global models generally agree that ocean deoxygenation is occurring; this finding is also reflected in in situ observations from past 50 years. Compilations of 10 Earth System models predict a global average loss of oxygen of –3.5% (RCP8.5) to –2.4% (RCP4.5) by 2100, but much stronger losses regionally, and in intermediate and mode waters (Bopp et al. 2013) (Figure 13.6). The North Pacific, North Atlantic, Southern Ocean, subtropical South Pacific, and South Indian Oceans all are expected to experience deoxygenation, with O_2 decreases of as much as 17% in the North Pacific by 2100 for the RCP8.5 pathway. However, the tropical Atlantic and tropical Indian Oceans show increasing O_2 concentrations. In the many areas where oxygen is declining, high natural variability makes it difficult to identify anthropogenically forced trends (Long et al. 2016).

[INSERT FIGURE 13.6 HERE]

REGIONAL MODELS

Regional models are critical because many oxygen drivers are local, influenced by bathymetry, winds, circulation, and fresh water and nutrient inputs. Most eastern boundary upwelling areas are predicted to experience intensified upwelling to 2100 (Wang et al. 2015), although on the West Coast projections for increasing upwelling for the northern California Current occur only north of San Francisco (see Section 13.2.3).

Particularly notable for the western United States, variation in trade winds in the eastern Pacific Ocean can affect nutrient inputs, leading to centennial periods of oxygen decline or oxygen increase distinct from global oxygen decline (Deutsch et al. 2014). Oxygen dynamics in the Eastern Tropical Pacific are highly sensitive to equatorial circulation changes (Montes et al. 2014).

Regional modeling also shows that year-to-year variability in precipitation in the central United States affects the nitrate–N flux by the Mississippi River and the extent of hypoxia in the Gulf of Mexico (Donner and Scavia 2007). A host of climate influences linked to warming and increased precipitation are predicted to lower dissolved oxygen in Chesapeake Bay (Najjar et al. 2010).

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13.5 Other Coastal Changes

13.5.1 Sea Level Rise

Sea level is an important variable that affects coastal ecosystems. Global sea level rose very rapidly at the end of the last glaciation, as glaciers and the polar ice sheets thinned and melted at their fringes. On average around the globe, sea level is estimated to have risen at rates exceeding 2.5 mm/year between about 8,000 and 6,000 years before present. These rates steadily decreased to less than 2.0 mm/year through about 4,000 years ago and stabilized at less than 0.4 mm/year through the late 1800s. Global sea level rise has accelerated again within the last 100 years, and now averages about 1 to 2 mm/year (Thompson et al. 2016). See Chapter 12: Sea Level Rise for more thorough analysis of how sea level rise has already and will affect the U.S. coasts.

13.5.2 Wet and Dry Deposition

Dust transported from continental desert regions to the marine environment deposits nutrients such as iron, nitrogen, phosphorus, and trace metals that stimulate growth of phytoplankton and increase marine productivity (Jickells and Moore 2015). U.S. continental and coastal regions experience large dust deposition fluxes originating from the Saharan desert to the East and from Central Asia and China to the Northwest (Chiapello 2014). Changes in drought frequency or intensity resulting from anthropogenically forced climate change, as well as other anthropogenic activities such as agricultural practices and land-use changes may play an important role in the future viability and strength of these dust sources (e.g., Mulitza et al. 2010).

Additionally, oxidized nitrogen, released during high-temperature combustion over land, and reduced nitrogen, released from intensive agriculture, are emitted in high population areas in North America and are carried away and deposited through wet or dry deposition over coastal and open ocean ecosystems via local wind circulation. Wet deposition of pollutants produced in urban areas is known to play an important role in changes of ecosystem structure in coastal and open ocean systems through intermediate changes in the biogeochemistry, for instance in dissolved oxygen or various forms of carbon (Paerl et al. 2002).

13.5.3 Primary Productivity

Marine phytoplankton represent about half of the global net primary production (NPP) (approximately 50 ± 28 GtC /year), fixing atmospheric CO_2 into a bioavailable form for utilization by higher trophic levels (see also Ch. 2: Physical Drivers of Climate Change; Carr et al. 2006; Franz et al. 2016). As such, NPP represents a critical component in the role of the oceans in climate feedback. The effect of climate change on primary productivity varies across the coasts depending on local conditions. For instance, nutrients that stimulate phytoplankton growth are impacted by various climate conditions, such as increased stratification which limits the transport of nutrient-rich deep water to the surface, changes in circulation leading to variability in dry and wet deposition of nutrients to coasts, and altered precipitation/evaporation

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which changes runoff of nutrients from coastal communities. The effect of the multiple physical factors on NPP is complex and leads to model uncertainties (Chavez et al. 2011). There is considerable variation in model projections for NPP, from estimated decreases or no changes, to the potential increase by 2100 (Frölicher et al. 2016; Fu et al. 2016; Laufkötter et al. 2015). Simulations from nine Earth system models projected total NPP in 2090 to decrease by 2%–16% and export production (that is, particulate flux to the deep ocean) to drop by 7%–18% as compared to 1990 (RCP8.5; Fu et al. 2016). More information on phytoplankton species response and associated ecosystem dynamics is needed as any reduction of NPP would have a strong impact on atmospheric CO_2 levels and marine ecosystems in general.

13.5.4 Estuaries

Estuaries are critical ecosystems of biological, economic, and social importance in the United States. They are highly dynamic, influenced by the interactions of atmospheric, freshwater, terrestrial, oceanic, and benthic components. Of the 28 national estuarine research reserves in the United States and Puerto Rico, all are being impacted by climate change to varying levels (Robinson et al. 2013). In particular, sea level rise, saltwater intrusion, and the degree of freshwater discharge influence the forces and processes within these estuaries (Monbaliu et al. 2014). Sea level rise and subsidence are leading to drowning of existing salt marshes and/or subsequent changes in the relative area of the marsh plain, if adaptive upslope movement is impeded due to urbanization along shorelines. Several model scenarios indicate a decline in salt marsh habitat quality and an accelerated degradation as the rate of sea level rise increases in the latter half of the century (Schile et al. 2014; Swanson et al. 2015). The increase in sea level as well as alterations to oceanic and atmospheric circulation can result in extreme wave conditions and storm surges, impacting coastal communities (Robinson et al. 2013). Additional climate change impacts to the physical and chemical estuarine processes include more extreme sea surface temperatures (higher highs and lower lows compared to the open ocean due to shallower depths and influence from land temperatures), changes in flow rates due to changes in precipitation, and potentially greater extents of salinity intrusion.

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TRACEABLE ACCOUNTS

Key Finding 1

The world's oceans have absorbed about 93% of the excess heat caused by greenhouse gas warming since the mid-20th century, making them warmer and altering global and regional climate feedbacks. Ocean heat content has increased at all depths since the 1960s and surface waters have warmed by about 1.3° ± 0.1°F (0.7° ± 0.08°C) per century globally since 1900 to 2016. Under a high emissions scenario, a global increase in average sea surface temperature of 4.9° ± 1.3°F (2.7° ± 0.7°C) by 2100 is projected, with even higher changes in some U.S. coastal regions. (*Very high confidence*)

Description of evidence base

The key finding and supporting text summarizes the evidence documented in climate science literature, including Rhein et al. 2013 and thereafter. Oceanic warming has been documented in a variety of data sources, most notably the WOCE (http://www.nodc.noaa.gov/woce/wdiu/), ARGO database (https://www.nodc.noaa.gov/argo/), and ERSSTv4 (https://www.ncdc.noaa.gov/data-access/marineocean-data/extended-reconstructed-sea-surface-temperature-ersst-v4). There is particular confidence in calculated warming for the time period since 1971 due to increased spatial and depth coverage and the level of agreement among independent SST observations from satellites, surface drifters and ships, and independent studies using differing analyses, bias corrections, and data sources (Cheng et al. 2017; Levitus et al. 2012; Llovel et al. 2014). Other observations such as the increase in mean sea level rise (see Ch. 12: Sea Level Rise) and reduced Arctic/Antarctic ice sheets (see Ch. 11: Arctic Changes) further confirm the increase in thermal expansion. For the purpose of extending the selected time periods back from 1900 to 2016 and analyzing U.S. regional SSTs, the Extended Reconstructed Sea Surface Temperature version 4 (ERSSTv4; Huang et al. 2015) is used. For the centennial time scale changes over 1900–2016, warming trends in all regions are statistically significant with the 95% confidence level. U.S. regional SST warming is similar between calculations using ERSSTv4 in this report and those published by Belkin (2016), suggesting confidence in these findings. The projected increase in SST is based on evidence from the latest generation of Earth System Models (CMIP5).

Major uncertainties

Uncertainties in the magnitude of ocean warming stem from the disparate measurements of ocean temperature over the last century. There is low uncertainty in warming trends of the upper ocean temperature from 0–700 m depth, whereas there is more uncertainty for deeper ocean depths of 700–2,000 m due to the short record of measurements from those areas. Data on warming trends at depths greater than 2,000 m are even more sparse. There are also uncertainties in the timing and reasons for particular decadal and interannual variations in ocean heat content and the contributions that different ocean basins play in the overall ocean heat uptake.

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Summary sentence or paragraph that integrates the above information

There is *very high confidence* in measurements that show increases in the ocean heat content and warming of the ocean, based on the agreement of different methods. However, long-term data in total ocean heat uptake in the deep ocean are sparse leading to limited knowledge of the transport of heat between and within ocean basins.

Key Finding 2

The potential slowing of the Atlantic Meridional Overturning Circulation (AMOC) (of which the Gulf Stream is one component)—as a result of increasing ocean heat content and freshwater driven buoyancy changes—could have dramatic climate feedbacks as the ocean absorbs less heat and CO_2 from the atmosphere. This slowing would also affect the climates of North America and Europe. Any slowing documented to date cannot be directly tied to anthropogenic forcing primarily due to lack of adequate observational data and to challenges in modeling ocean circulation changes. Under a high emissions scenario (RCP8.5) in CMIP5 simulations, it is likely that the AMOC will weaken over the 21st century by 12% to 54%. (*Low confidence*)

Description of evidence base

Investigations both through direct observations and models since 2013 (Rhein et al. 2013) have raised significant concerns about whether there is enough evidence to determine the existence of an overall slowdown in the AMOC. As a result, more robust international observational campaigns are underway currently to measure AMOC circulation. Direct observations have determined a statistically significant slowdown at the 95% confidence level at 26°N (off Florida; see Baringer et al. 2016), but modeling studies constrained with observations cannot attribute this to anthropogenic forcing (Jackson et al. 2016). The study (Rahmstorf et al. 2015) which seemed to indicate broad-scale slowing has since been discounted due to its heavy reliance on sea surface temperature cooling as proxy for slowdown rather than actual direct observations. Since Rhein et al. 2013, more observations have led to increased statistical confidence in the measurement of the AMOC. Current observation trends indicate the AMOC slowing down at the 95% confidence level at 26°N and 41°N but a more limited in situ estimate at 35°S shows an increase in the AMOC (Smeed et al. 2014; Baringer et al. 2016). There is no one collection spot for AMOC-related data, but the U.S. Climate Variability and Predictability Program (US CLIVAR) has a U.S. AMOC priority focus area and a webpage with relevant data sites (https://usclivar.org/amoc/amoc-time-series).

The IPCC 2013 WG1 projections indicate a high likelihood of AMOC slowdown in the next 100 years, however overall understanding is limited by both a lack of direct observations (which is being remedied) and a lack of model skill to resolve deep ocean dynamics. As a result, this key finding was given an overall assessment of *low confidence*.

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Major uncertainties

As noted, uncertainty about the overall trend of the AMOC is high given opposing trends in northern and southern ocean time series observations. Although earth system models do indicate a high likelihood of AMOC slowdown as a result of a warming, climate projections are subject to high uncertainty. This uncertainty stems from intermodel differences, internal variability that is different in each model, uncertainty in stratification changes, and most importantly uncertainty in both future freshwater input at high latitudes as well as the strength of the subpolar gyre circulation.

Summary sentence or paragraph that integrates the above information

The increased focus on direct measurements of the AMOC should lead to a better understanding of 1) how it is changing and its variability by region, and 2) whether those changes are attributable to climate drivers through both model improvements and incorporation of those expanded observations into the models.

Key Finding 3

The world's oceans are currently absorbing more than a quarter of the CO_2 emitted to the atmosphere annually from human activities, making them more acidic (*very high confidence*), with potential detrimental impacts to marine ecosystems. In particular, higher-latitude systems typically have a lower buffering capacity against pH change, exhibiting seasonally corrosive conditions sooner than low-latitude systems. Acidification is regionally increasing along U.S. coastal systems as a result of upwelling (for example, in the Pacific Northwest) (*high confidence*), changes in freshwater inputs (for example, in the Gulf of Maine) (*medium confidence*), and nutrient input (for example, in urbanized estuaries) (*high confidence*). The rate of acidification is unparalleled in at least the past 66 million years (*medium confidence*). Under RCP8.5, the global average surface ocean acidity is projected to increase by 100% to 150% (*high confidence*).

Description of evidence base

Evidence on the magnitude of the ocean sink is obtained from multiple biogeochemical and transport ocean models and two observation-based estimates from the 1990s for the uptake of the anthropogenic CO_2. Estimates of the carbonate system (DIC and alkalinity) were based on multiple survey cruises in the global ocean in the 1990s (WOCE, JGOFS). Coastal carbon and acidification surveys have been executed along the U.S. coastal large marine ecosystem since at least 2007, documenting significantly elevated pCO_2 and low pH conditions relative to oceanic waters. The data is available from the National Centers for Environmental Information (https://www.ncei.noaa.gov/). Other sources of biogeochemical bottle data can be found from

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HOT-DOGS ALOHA (http://hahana.soest.hawaii.edu/hot/hot-dogs) or ERSL/GFM Data Finder (https://www.esrl.noaa.gov/gmd/dv/data). Rates of change associated with the Palaeocene-Eocene Thermal Maximum (PETM, 56 million years ago) were derived using stable carbon and oxygen isotope records preserved in the sedimentary record from the New Jersey shelf using time series analysis and carbon cycle–climate modelling. This evidence supports a carbon release during the onset of the PETM over no less than 4,000 years, yielding a maximum sustained carbon release rate of less than 1.1 GtC per year (Zeebe et al. 2016). The projected increase in global surface ocean acidity is based on evidence from ten of the latest generation earth system models which include six distinct biogeochemical models that were included in the latest IPCC AR5 2013.

Major uncertainties

In 2014 the ocean sink was 2.6 ± 0.5 GtC (9.5 $GtCO_2$), equivalent to 26% of the total emissions attributed to fossil fuel use and land use changes (Le Quéré et al. 2016). Estimates of the PETM ocean acidification event evidenced in the geological record remains a matter of some debate within the community. Evidence for the 1.1 GtC per year cited by Zeebe et al. (2016), could be biased as a result of brief pulses of carbon input above average rates of emissions were they to transpire over timescales ≲ 40 years.

Summary sentence or paragraph that integrates the above information

There is *very high confidence* in evidence that the oceans absorb about a quarter of the carbon dioxide emitted in the atmosphere and hence become more acidic. The magnitude of the ocean carbon sink is known at a *high confidence* level because it is estimated using a series of disparate data sources and analysis methods, while the magnitude of the interannual variability is based only on model studies. There is medium confidence that the current rate of climate acidification is unprecedented in the past 66 million years. There is also *high confidence* that oceanic pH will continue to decrease.

Key Finding 4

Increasing sea surface temperatures, rising sea levels, and changing patterns of precipitation, winds, nutrients, and ocean circulation are contributing to overall declining oxygen concentrations at intermediate depths in various ocean locations and in many coastal areas. Over the last half century, major oxygen losses have occurred in inland seas, estuaries, and in the coastal and open ocean (*high confidence*). Ocean oxygen levels are projected to decrease by as much as 3.5% under the RCP8.5 scenario by 2100 relative to preindustrial values (*high confidence*).

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Description of evidence base

The key finding and supporting text summarizes the evidence documented in climate science literature including Rhein et al. 2013, Bopp et al. 2013, and Schmidtko et al. 2017. Evidence arises from extensive global measurements of the World Ocean Circulation Experiment (WOCE) after 1989 and individual profiles before that (Helm et al. 2011). The first basin-wide dissolved oxygen surveys were performed in the 1920s (Schmidtko et al. 2017). The confidence level is based on globally integrated O_2 distributions in a variety of ocean models. Although the global mean exhibits low interannual variability, regional contrasts are large.

Major uncertainties

Uncertainties (as estimated from the intermodel spread) in the global mean are moderate mainly because ocean oxygen content exhibits low interannual variability when globally averaged. Uncertainties in long-term decreases of the global averaged oxygen concentration amount to 25% in the upper 1,000 m for the 1970–1992 period and 28% for the 1993–2003 period. Remaining uncertainties relate to regional variability driven by mesoscale eddies and intrinsic climate variability such as ENSO.

Summary sentence or paragraph that integrates the above information

Major ocean deoxygenation is taking place in bodies of water inland, at estuaries, and in the coastal and the open ocean (*high confidence*). Regionally, the phenomenon is exacerbated by local changes in weather, ocean circulation, and continental inputs to the oceans.

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TABLE

Table 13.1. Historical sea surface temperature trends (°C per century) and projected trends by 2080 (°C) for eight U.S. coastal regions and globally. Historical temperature trends are presented for the 1900–2016 and 1950–2016 periods with 95% confidence level, observed using the Extended Reconstructed Sea Surface Temperature version 4 (ERSSTv4; Huang et al. 2015). Global and regional predictions are calculated for RCP4.5 and RCP8.5 emission levels with 80% spread of all the CMIP5 members compared to the 1976–2005 period (Scott et al. 2016). The historical trends were analyzed for the latitude and longitude in the table, while the projected trends were analyzed for the California current instead of the Northwest and Southwest separately and for the Bering Sea in Alaska (NOAA).

Region	latitude and longitude	Historical Trend (°C/100 years)		Projected Trend by 2080 (relative to 1976-2005 climate) (°C)	
		1900–2016	1950–2016	RCP4.5	RCP8.5
Global		0.70 ± 0.08	1.00 ± 0.11	1.3 ± 0.6	2.7 ± 0.7
Alaska	50°–66°N, 150°–170°W	0.82 ± 0.26	1.22 ± 0.59	2.5 ± 0.6	3.7 ± 1.0
Northwest (NW)	40°–50°N, 120°–132°W	0.64 ± 0.30	0.68 ± 0.70	1.7 ± 0.4	2.8 ± 0.6
Southwest (SW)	30°–40°N, 116°–126°W	0.73 ± 0.33	1.02 ± 0.79		
Hawaii (HI)	18°–24°N, 152°–162°W	0.58 ± 0.19	0.46 ± 0.39	1.6 ± 0.4	2.8 ± 0.6
Northeast (NE)	36°–46°N, 64°–76°W	0.63 ± 0.31	1.10 ± 0.71	2.0 ± 0.3	3.2 ± 0.6

Southeast (SE)	24°–34°N, 64°–80°W	0.40 ± 0.18	0.13 ± 0.34	1.6 ± 0.3	2.7 ± 0.4
Gulf of Mexico (GOM)	20°–30°N, 80°–96°W	0.52 ± 0.14	0.37 ± 0.27	1.6 ± 0.3	2.8 ± 0.3
Caribbean	10°–20°N, 66°–86°W	0.76 ± 0.15	0.77 ± 0.32	1.5 ± 0.4	2.6 ± 0.3

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FIGURES

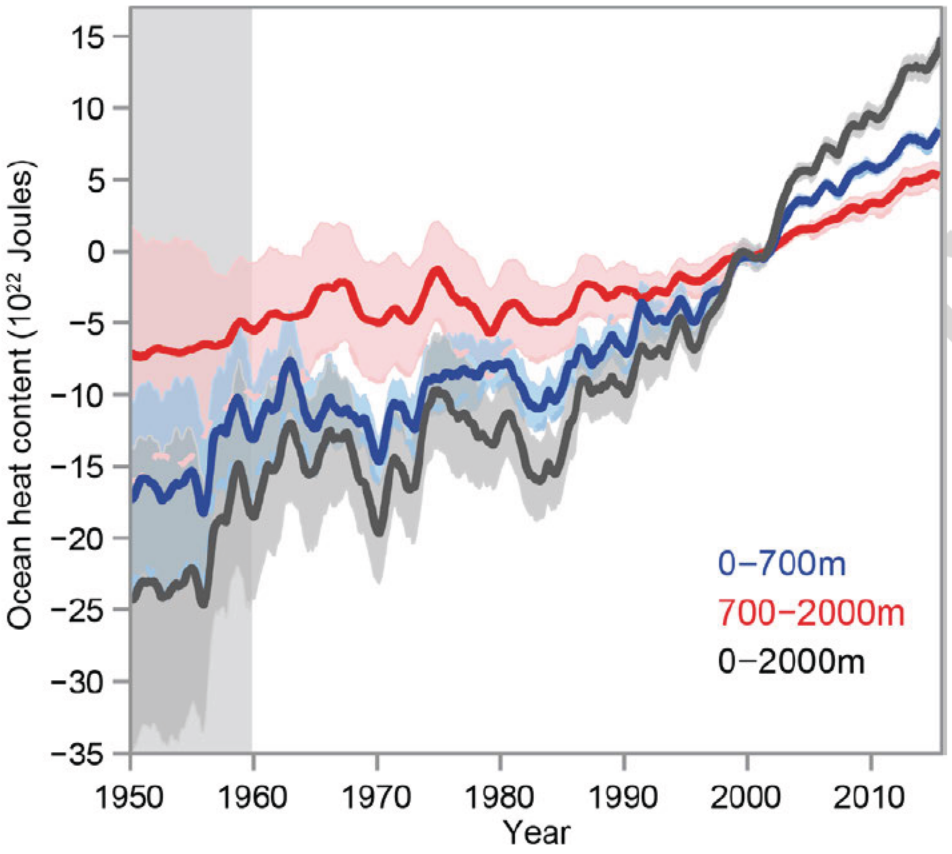

Figure 13.1: Global Ocean heat content change time series. Ocean heat content from 0 to 700 m (blue), 700 to 2,000 m (red), and 0 to 2,000 m (dark gray) from 1955 to 2015 with an uncertainty interval of ±2 standard deviations shown in shading. All time series of the analysis performed by Cheng et al. (2017) are smoothed by a 12-month running mean filter, relative to the 1997–2005 base period. (Figure source: Cheng et al. 2017).

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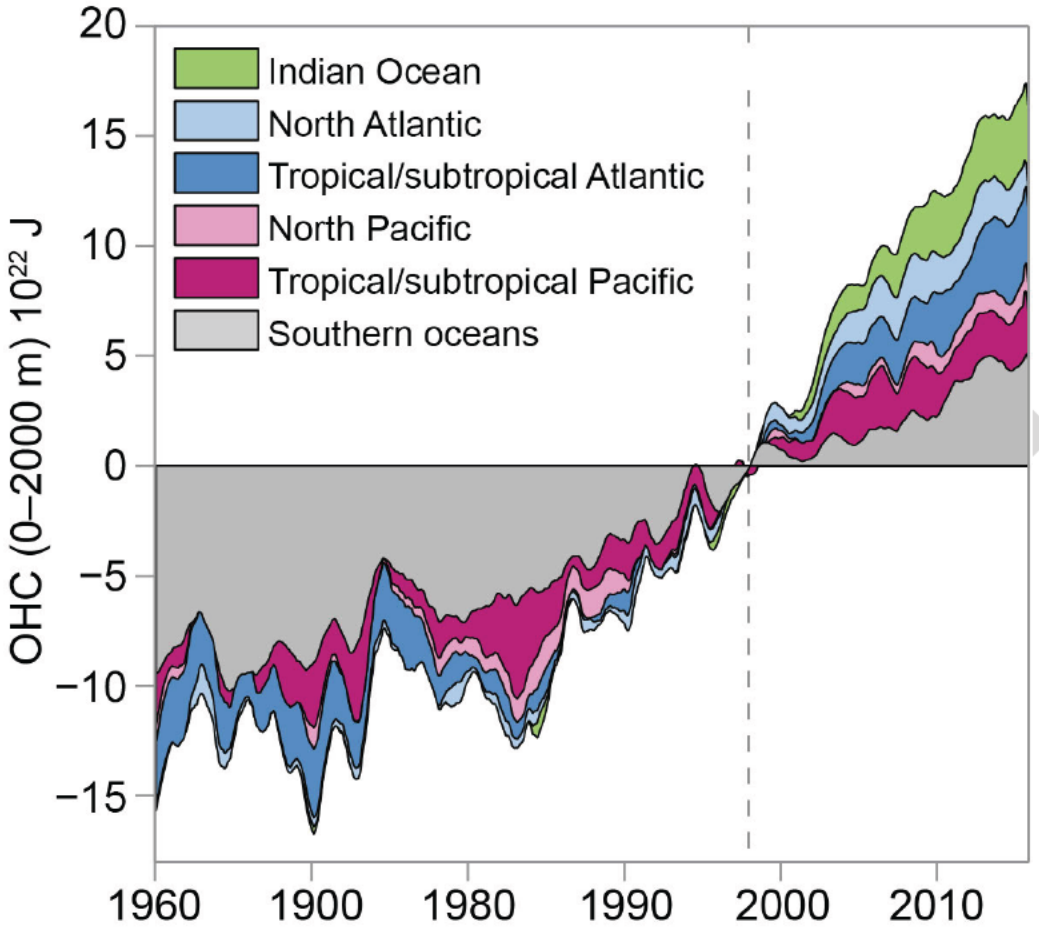

Figure 13.2: Ocean heat content changes from 1960 to 2015 for different ocean basins for 0 to 2,000 m depths. Time series is relative to the 1997–1999 base period and smoothed by a 12-month running filter by Cheng et al. (2017). The curves are additive, and the ocean heat content changes in different ocean basins are shaded in different colors (Figure source: Cheng et al. 2017).

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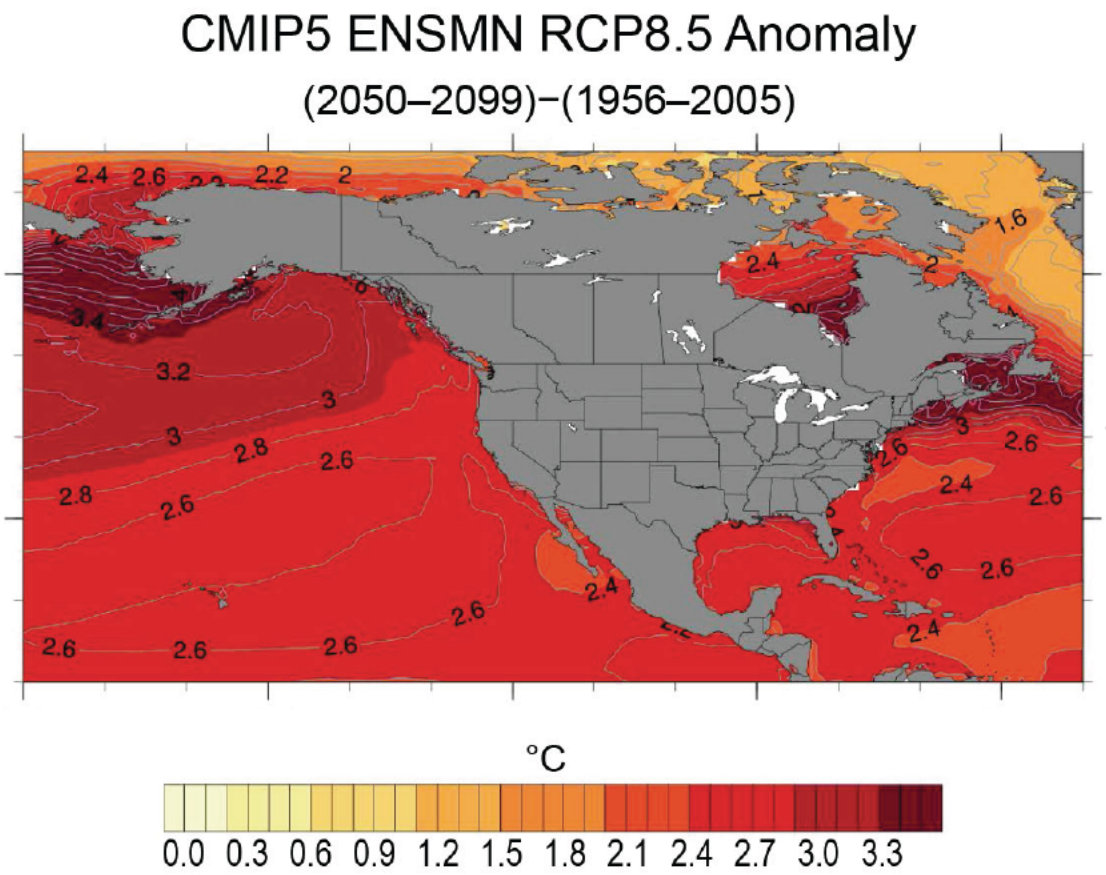

Figure 13.3: Projected changes in sea surface temperature (°C) for the coastal United States under emission scenario RCP8.5. Projected anomalies for the 2050–2099 period are calculated using a comparison from the average sea surface temperatures over 1956–2005. Projected changes are examined using the Coupled Model Intercomparison Project Phase 5 (CMIP5) suite of model simulations. (Figure source: NOAA).

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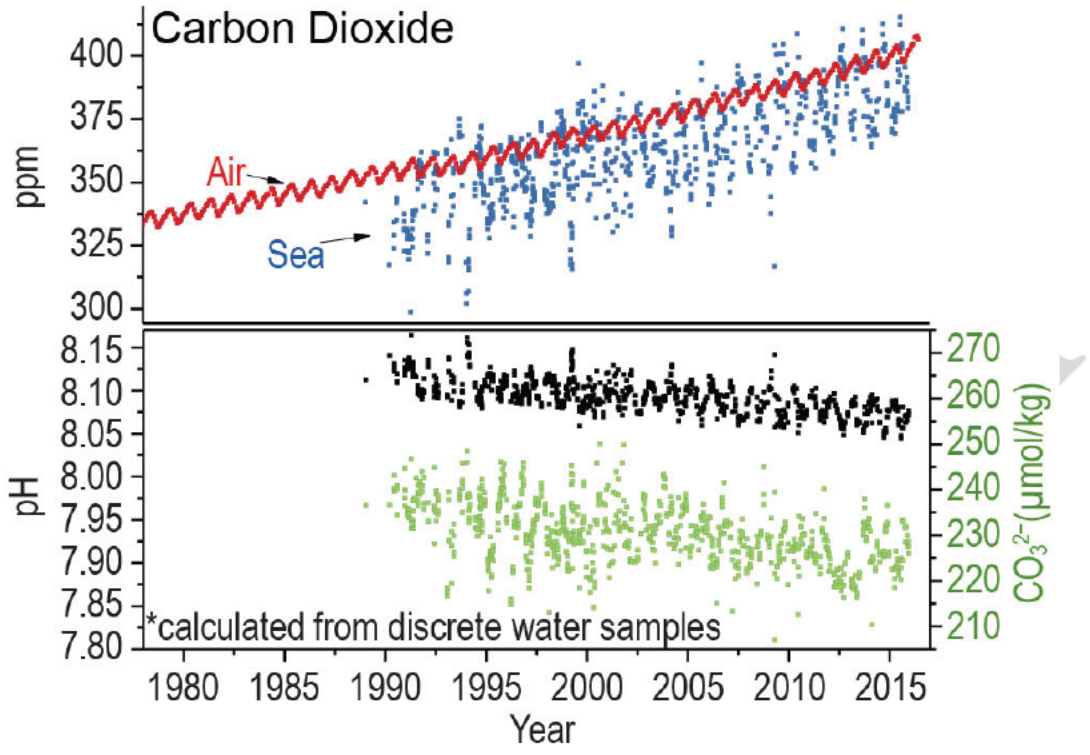

Figure 13.4: Trends in surface (< 50 m) ocean carbonate chemistry calculated from observations obtained at the Hawai'i Ocean Time-series (HOT) Program in the North Pacific over 1988–2015. The upper panel shows the linked increase in atmospheric (red points) and seawater (blue points) CO_2 concentrations. The bottom panels shows a decline in seawater pH (black points, primary y-axis) and carbonate ion concentration (green points, secondary y-axis). Ocean chemistry data were obtained from the Hawai'i Ocean Time-series Data Organization & Graphical System (HOT-DOGS, http://hahana.soest.hawaii.edu/hot/hot-dogs/index.html). (Figure source: NOAA).

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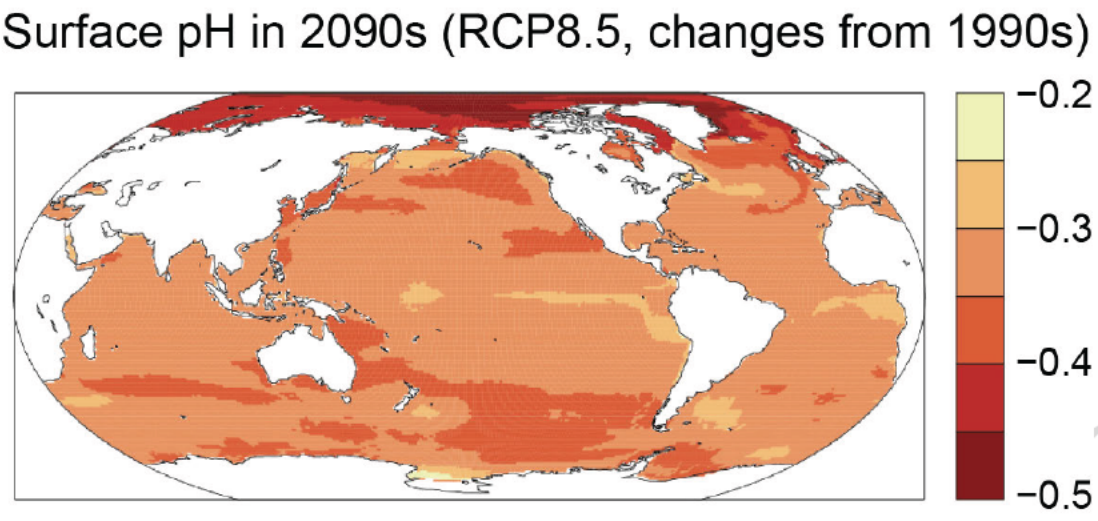

Figure 13.5 Predicted change in sea surface pH in 2090–2099 relative to 1990–1999 under RCP8.5, based on the Community Earth System Models–Large Ensemble Experiments CMIP5 (Figure source: adapted from Bopp et al. 2013).

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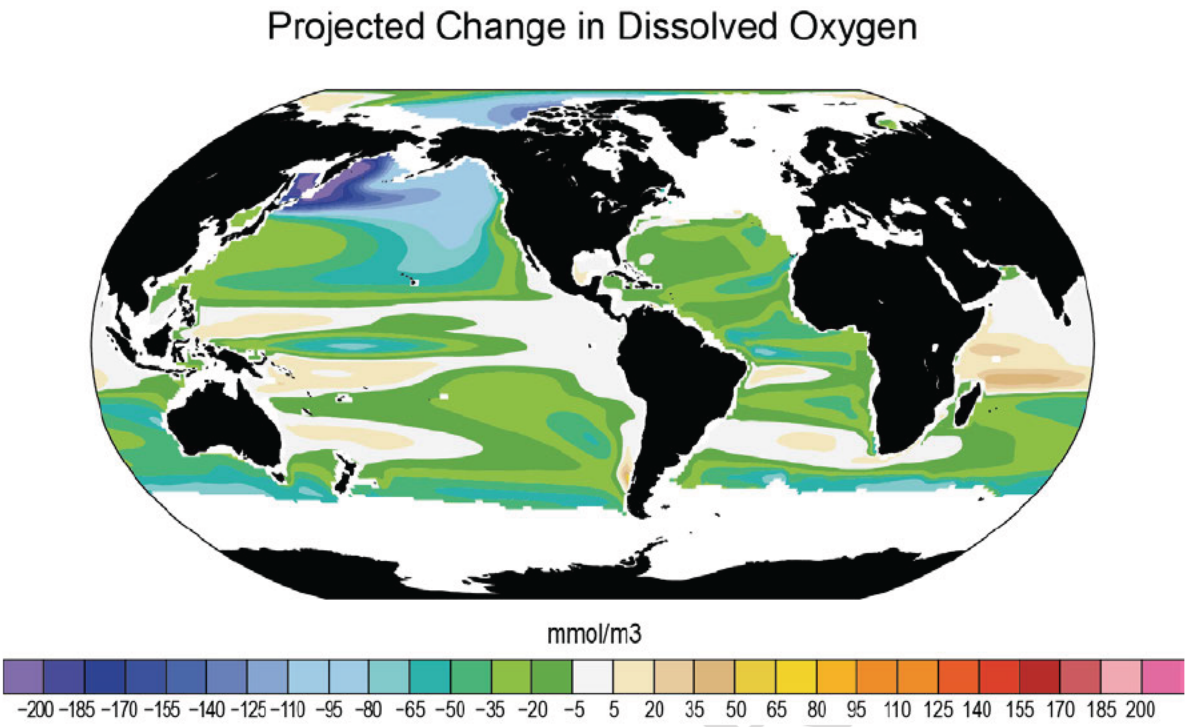

Figure 13.6: Predicted change in dissolved oxygen on the $\sigma_\theta = 26.5$ (average depth of approximately 290 m) potential density surface, between the 1981–2000 and 2081–2100, based on the Community Earth System Models–Large Ensemble Experiments (Figure source: redrawn from Long et al. 2016).

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14. Perspectives on Climate Change Mitigation

KEY FINDINGS

1. Warming and associated climate effects from CO_2 emissions persist for decades to millennia. In the near-term, changes in climate are determined by past and present greenhouse gas emissions modified by natural variability. Reducing the total concentration of atmospheric CO_2 is necessary to limit near-term climate change and stay below long-term warming targets (such as the oft-cited 3.6°F [2°C] goal). Other greenhouse gases (for example, methane) and black carbon aerosols exert stronger warming effects than CO_2 on a per ton basis, but they do not persist as long in the atmosphere; therefore, mitigation of non-CO_2 species contributes substantially to near-term cooling benefits but cannot be relied upon for ultimate stabilization goals. (*Very high confidence*)

2. Stabilizing global mean temperature below long-term warming targets requires an upper limit on the accumulation of CO_2 in the atmosphere. The relationship between cumulative CO_2 emissions and global temperature response is estimated to be nearly linear. Nevertheless, in evaluating specific temperature targets, there are uncertainties about the exact amount of compatible anthropogenic CO_2 emissions due to uncertainties in climate sensitivity, the response of the carbon cycle including feedbacks, the amount of past CO_2 emissions, and the influence of past and future non-CO_2 species. (*Very high confidence*)

3. Stabilizing global mean temperature below 3.6°F (2°C) or lower relative to preindustrial levels requires significant reductions in net global CO_2 emissions relative to present-day values before 2040 and likely requires net emissions to become zero or possibly negative later in the century. Accounting for the temperature effects of non-CO_2 species, cumulative CO_2 emissions are required to stay below about 800 GtC in order to provide a two-thirds likelihood of preventing 3.6°F (2°C) of warming, meaning approximately 230 GtC more could be emitted globally. Assuming global emissions follow the range between the RCP8.5 and RCP4.5 scenarios, emissions could continue for approximately two decades before this cumulative carbon threshold is exceeded. (*High confidence*)

4. Successful implementation of the first round of Nationally Determined Contributions associated with the Paris Agreement will provide some likelihood of meeting the long-term temperature goal of limiting global warming to "well below" 3.6°F (2°C) above preindustrial levels; the likelihood depends strongly on the magnitude of global emission reductions after 2030. (*High confidence*)

5. Climate intervention or geoengineering strategies such as solar radiation management are measures that attempt to limit or reduce global temperature increases. If interest in geoengineering increases with observed impacts and/or projected risks of climate change,

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interest will also increase in assessments of the technical feasibilities, costs, risks, co-benefits, and governance challenges of these additional measures, which are as yet unproven at scale. These assessments are a necessary step before judgments about the benefits and risks of these approaches can be made with high confidence. (*High confidence*)

Introduction

This chapter provides scientific context for key issues regarding the long-term mitigation of climate change. As such, this chapter first addresses the science underlying the timing of when and how CO_2 and other greenhouse gas (GHG) mitigation activities that occur in the present affect the climate of the future. When do we see the benefits of a GHG emission reduction activity? Chapter 4: Projections provides further context for this topic. Relatedly, the present chapter discusses the significance of the relationship between cumulative CO_2 emissions and eventual global warming levels. The chapter reviews studies of the climate effects of the first round of national contributions associated with the Paris Agreement if fully implemented. Looking beyond the first round of national contributions (which do not set emission reduction targets past 2030), what global-scale emissions pathways are estimated to be necessary by mid-century and beyond in order to have a high likelihood of preventing 3.6°F (2°C) or 2.7°F (1.5°C) of warming relative to preindustrial times? In response to this question, this chapter briefly reviews the status of climate intervention proposals and how these types of mitigation actions could possibly play a role in avoiding future climate change.

14.1 The Timing of Benefits from Mitigation Actions

14.1.1 Lifetime of Greenhouse Gases and Inherent Delays in the Climate System

Carbon dioxide (CO_2) concentrations in the atmosphere are directly affected by human activities in the form of CO_2 emissions. Atmospheric CO_2 concentrations adjust to human emissions of CO_2 over long time scales, spanning from decades to millennia (Ciais et al. 2013; Joos et al. 2013). The IPCC estimated that 15% to 40% of CO_2 emitted until 2100 will remain in the atmosphere longer than 1,000 years (Ciais et al. 2013). The persistence of warming is longer than the atmospheric lifetime of CO_2 and other GHGs, owing in large part to the thermal inertia of the ocean (Collins et al. 2013). Climate change resulting from anthropogenic CO_2 emissions, and any associated risks to the environment, human health and society, are thus essentially irreversible on human timescales (Solomon et al. 2009). The world is committed to some degree of irreversible warming and associated climate change resulting from emissions to date.

The long lifetime in the atmosphere of CO_2 (Joos et al. 2013) and some other key GHGs, coupled with the time lag in the response of the climate system to atmospheric forcing

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(Tebaldi and Friedlingstein 2013), has timing implications for the benefits (i.e., avoided warming or risk) of mitigation actions. Large reductions in emissions of the long-lived GHGs are estimated to have modest temperature effects in the *near term* (e.g., over one to two decades), because total atmospheric concentration levels require long periods to adjust (Prather et al. 2009), but are necessary in the *long term* to achieve any objective of preventing warming of any desired magnitude. Near-term projections of global mean surface air temperature are therefore not strongly influenced by changes in emissions but rather dominated by natural variability, the Earth system response to past and current GHG emissions, and by model spread (i.e., the different climate outcomes associated with different models using the same emissions scenario) (Kirtman et al. 2013). Long-term projections of global surface temperature (after mid-century), on the other hand, show that emissions scenario choice, and thus the mitigation pathway, is the dominant source of future uncertainty in climate outcomes (Paltsev et al. 2015; Collins et al. 2013).

Some studies have nevertheless shown the potential for some near-term benefits of mitigation. For example, one study found that, even at the regional scale, heat waves would already be significantly more severe by the 2030s in a non-mitigation scenario compared to a moderate mitigation scenario (Tebaldi and Wehner 2016). The mitigation of non-CO_2 GHGs with short atmospheric lifetimes (such as methane, some hydrofluorocarbons [HFCs], and ozone) and black carbon (an aerosol that absorbs solar radiation; see Ch. 2: Physical Drivers of Climate Change), collectively referred to as short-lived climate pollutants (SLCPs), has been highlighted as a particular way to achieve more rapid climate benefits (e.g., Zaelke and Borgford-Parnell 2015). SLCPs are substances that not only have an atmospheric lifetime shorter (for example, weeks to a decade) than CO_2 but also exert a stronger radiative forcing (and hence temperature effect) compared to CO_2 on a per ton basis (Myhre et al. 2013). For these reasons, mitigation of SLCP emissions produces more rapid radiative responses. In the case of black carbon, with an atmospheric lifetime of a few days to weeks (Bond et al. 2013), emissions (and therefore reductions of those emissions) produce strong regional effects. Mitigation of black carbon and methane also generate direct health co-benefits (Anenberg et al. 2012; Rao et al. 2016). Reductions and/or avoidances of SLCP emissions could be a significant contribution to staying at or below a 3.6°F (2°C) or any other chosen global mean temperature increase (Hayhoe et al. 1998; Shah et al. 2015; Shindell et al. 2012; Rogelj et al. 2015). The recent Kigali Amendment to the Montreal Protocol seeks to phase down global HFC production and consumption in order to avoid substantial GHG emissions in coming decades. Stringent near-term SLCP mitigation could potentially increase allowable CO_2 budgets for avoiding warming beyond any desired future level, by up to 25% under certain scenarios (Rogelj et al. 2015). However, given that economic and technological factors tend to couple CO_2 and many SLCP emissions to varying degrees, significant SLCP emissions reduction would be a co-benefit of CO_2 mitigation.

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14.1.2 Stock and Stabilization: Cumulative CO_2 and the Role of Other Greenhouse Gases

Cumulative CO_2 emissions in the industrial era will largely determine long-term, global mean temperature change. A robust feature of model climate change simulations is a nearly linear relationship between cumulative CO_2 emissions and global mean temperature increases, irrespective of the details and exact timing of the emissions pathway (see Figure 14.1; see also Ch. 4: Projections). Limiting and stabilizing warming to any level implies that there is an upper limit to the cumulative amount of CO_2 that can be added to the atmosphere (Collins et al. 2013). Eventually stabilizing the global temperature requires CO_2 emissions to approach zero (NRC 2011). Thus, for a 3.6°F (2°C) or any desired global mean temperature target, an estimated range of allowable cumulative CO_2 emissions from the current period onward can be calculated. The key sources of uncertainty for any compatible, forward looking CO_2 budget associated with a given future warming objective include the climate sensitivity, the response of the carbon cycle including feedbacks (for example, the release of GHGs from permafrost thaw), the amount of past CO_2 emissions, and the influence of past and future non-CO_2 species (Collins et al. 2013; NRC 2011). Increasing the probability that any given temperature target be reached therefore implies tighter constraints on cumulative CO_2 emissions. Relatedly, for any given cumulative CO_2 budget, higher emissions in the near term imply the need for steeper reductions in the long term.

[INSERT FIGURE 14.1 HERE]

Between 1870 and 2015, human activities, primarily the burning of fossil fuels and deforestation, emitted about 560 GtC in the form of CO_2 into the atmosphere (Le Quéré et al. 2016). According to best estimates in the literature, 1,000 GtC is the total cumulative amount of CO_2 that could be emitted yet still provide a two-thirds likelihood of preventing 3.6°F (2°C) of global mean warming since pre-industrial times (Collins et al. 2013; Allen et al. 2009). That estimate, however, ignores the additional radiative forcing effects of non-CO_2 species (that is, the net positive forcing resulting from the forcing of other well-mixed GHGs, including halocarbons, plus the other ozone precursor gases and aerosols). Considering both historic and projected non-CO_2 effects reduces the estimated cumulative CO_2 budget compatible with any future warming target (Rogelj et al. 2015), and in the case of 3.6°F (2°C) it reduces the aforementioned estimate to 790 GtC (Collins et al. 2013). Given this more comprehensive estimate, meeting the 3.6°F (2°C) target means approximately 230 GtC more CO_2 could be emitted globally. To illustrate, if one assumes future global emissions follow the RCP4.5 scenario, this cumulative carbon threshold is exceeded by around 2037, while under the RCP8.5 scenario this occurs by around 2033. To meet a 2.7°F (1.5°C) target, the estimated cumulative CO_2 budget is about 590 GtC (assuming linear scaling with the compatible 3.6°F (2°C) budget that also considers non-CO_2 effects), meaning only about 30

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GtC more of CO_2 could be emitted. Further emissions of 30 GtC (in the form of CO_2) are projected to occur in the next few years (Table 14.1).

[INSERT TABLE 14.1 HERE]

14.2 Pathways Centered Around 3.6°F (2°C)

In December of 2015 in Paris, the Parties to the United Nations Framework Convention on Climate Change (UNFCCC) adopted the Paris Agreement, under which all Parties committed to prepare and communicate successive Nationally Determined Contributions (NDCs) to mitigate climate change. The first NDCs extend to 2025 or 2030 and take a wide range of forms. The Agreement contains the long-term goal of "holding the increase in the global average temperature to well below 2°C above pre-industrial levels and pursuing efforts to limit the temperature increase to 1.5°C above pre-industrial levels."

Estimates of global emissions and temperature implications from a successful implementation of the first round of NDCs (Rogelj et al. 2016; Sanderson et al. 2016; Climate Action Tracker 2016; Fawcett et al. 2015; UNFCCC 2015) generally find that: 1) the first round of NDCs reduces GHG emissions growth by 2030 relative to a situation where these goals did not exist, though emissions are still not expected to be lower in 2030 than in 2015; and 2) the NDCs are a step towards meeting a 3.6°F (2°C) target, but the NDCs are, by themselves, insufficient to achieve this ambitious target. According to one study, the NDCs imply a median warming of 4.7°–5.6°F (2.6°–3.1°C) by 2100, though year 2100 temperature estimates depend on assumed emissions between 2030 and 2100 (Rogelj et al. 2016). For example, Climate Action Tracker, using alternative post-2030 assumptions, put the range at 5.9°–7.0°F (3.3°–3.9°C).

Emissions pathways consistent with the NDCs have been evaluated in the context of the likelihood of global mean surface temperature change (Figure 14.2). It was found that the likelihood of meeting the 3.6°F (2°C) or less target was enhanced by the NDCs, but depended strongly on subsequent policies and measures. The chief finding was that even without additional emission reductions after 2030, if implemented successfully, the NDCs provide some likelihood (less than 10%) of preventing a global mean surface temperature change of 3.6°F (2°C) relative to preindustrial levels (Fawcett et al. 2015). Greater emissions reductions beyond 2030 (here, based on assumed higher decarbonization rates past 2030) increase the likelihood of achieving the 3.6°F (2°C) or lower target to about 30%, and almost eliminate the likelihood of a global mean temperature increase greater than 7°F (4°C). Scenarios that assume even greater emissions reductions past 2030 would be necessary to have at least a 50% probability of limiting warming to 3.6°F (2°C) (Fawcett et al. 2015), as discussed and illustrated further below.

[INSERT FIGURE 14.2 HERE]

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There are only a limited number of pathways which enable the world to remain below 3.6°F (2°C) of warming (see Figure 14.3), and almost all but the most rapid near-term mitigation pathways are heavily reliant on the implementation of CO_2 removal from the atmosphere later in the century or other climate intervention, discussed below. If global emissions are in line with the first round of NDCs by 2030, then the world likely needs to reduce effective GHG emissions to zero by 2080, and be significantly net negative by the end of the century (relying on as yet unproven technologies to remove GHGs from the atmosphere) in order to stay below 3.6°F (2°C) of warming. Avoiding 2.7°F (1.5°C) of warming requires more aggressive action still, with net zero emissions achieved by 2050 and net negative emissions thereafter. In either case, faster near-term action significantly decreases the requirements for negative emissions in the future.

[INSERT FIGURE 14.3 HERE]

14.3 The Role of Climate Intervention in Meeting Ambitious Climate Targets

Achieving a 3.6°F (2°C) target through emissions reductions or adapting to the impacts of a greater-than-3.6°F (2°C) world have been acknowledged as severely challenging tasks by the international science and policy communities. Consequently, there is increased interest by some scientists and policy makers in exploring additional measures designed to reduce net radiative forcing through other, as yet untested actions, which are often referred to as geoengineering or climate intervention (CI) actions. CI approaches are generally divided into two categories: CO_2 removal (CDR) (NAS 2015a) and solar radiation management (SRM) (NAS 2015b). CDR and SRM methods may have future roles in helping meet global temperature targets. Both methods would reduce global average temperature by reducing net global radiative forcing: CDR through reducing atmospheric CO_2 concentrations and SRM through increasing Earth's albedo.

The evaluation of the suitability and advisability of potential CI actions requires a decision framework that includes important dimensions beyond scientific and technical considerations. Among these dimensions to be considered are the potential development of global and national governance and oversight procedures, geopolitical relations, legal considerations, environmental, economic and societal impacts, ethical considerations, and the relationships to global climate policy and current GHG mitigation and adaptation actions. It is clear that these social science and other non-physical science dimensions are likely to be the major part of the decision framework and ultimately control the adoption and effectiveness of CI actions. This report only acknowledges these mostly non-physical scientific dimensions and must forego a detailed discussion.

By removing CO_2 from the atmosphere, CDR directly addresses the principal cause of climate change. Potential CDR approaches include point-source CO_2 capture, direct air

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capture, currently well-understood biological methods on land (for example, afforestation), less well-understood and potentially risky methods in the ocean (for example, ocean fertilization), and accelerated weathering (for example, forming calcium carbonate on land or in the oceans). While CDR is technically possible, the primary challenge is achieving the required scale of removal in a cost-effective manner, which in part presumes a comparison to the costs of other, more traditional GHG mitigation options. In principle, at large scale, CDR could measurably reduce CO_2 concentrations (that is, cause negative emissions). Point-source capture (as opposed to CO_2 capture from ambient air) and removal of CO_2 is a particularly effective CDR method. The climate value of avoided CO_2 emissions is essentially equivalent to that of the atmospheric removal of the same amount. To realize sustained climate benefits from CDR, however, the removal of CO_2 from the atmosphere must be essentially permanent—at least several centuries to millennia. In addition to high costs, CDR has the additional limitation of long implementation times.

By contrast, SRM approaches offer the only known CI methods of cooling Earth within a few years after inception. An important limitation of SRM is that it would not address damage to ocean ecosystems from increasing ocean acidification due to continued CO_2 uptake. SRM could theoretically have a significant global impact even if implemented by a small number of nations, and by nations that are not also the major emitters of GHGs; this could be viewed either as a benefit or risk of SRM.

Proposed SRM concepts increase Earth's albedo through injection of sulfur gases or aerosols into the stratosphere (thereby simulating the effects of explosive volcanic eruptions) or marine cloud brightening through aerosol injection near the ocean surface. Injection of solid particles is an alternative to sulfur and yet other SRM methods could be deployed in space. Studies have evaluated the expected effort and effectiveness of various SRM methods (NAS 2015b; Keith et al. 2014). For example, model runs were performed in the GeoMIP project using the full CMIP5 model suite to illustrate the effect of reducing top-of-the-atmosphere insolation to offset climate warming from CO_2 (Kravitz et al. 2013). The idealized runs, which assumed an abrupt, globally-uniform insolation reduction in a $4 \times CO_2$ atmosphere, show that temperature increases are largely offset, most sea-ice loss is avoided, average precipitation changes are small, and net primary productivity increases. However, important regional changes in climate variables are likely in SRM scenarios as discussed below.

As global ambitions increase to avoid or remove CO_2 emissions, probabilities of large increases in global temperatures by 2100 are proportionately reduced (Fawcett et al. 2015). Scenarios in which large-scale CDR is used to meet a 3.6°F (2°C) limit while allowing business-as-usual consumption of fossil fuels are likely not feasible with present technologies. Model SRM scenarios have been developed that show reductions in

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radiative forcing up to 1 W/m^2 with annual stratospheric injections of 1 Mt of sulfur from aircraft or other platforms (Pierce et al. 2010; Tilmes et al. 2016). Preliminary studies suggest that this could be accomplished at a cost as low as a few billion dollars per year using current technology, enabling an individual country or subnational entity to conduct activities having significant global climate impacts.

SRM scenarios could in principle be designed to follow a particular radiative forcing trajectory, with adjustments made in response to monitoring of the climate effects (Keith and MacMartin 2015). SRM could be used as an interim measure to avoid peaks in global average temperature and other climate parameters. The assumption is often made that SRM measures, once implemented, must continue indefinitely in order to avoid the rapid climate change that would occur if the measures were abruptly stopped. SRM could be used, however, as an interim measure to buy time for the implementation of emissions reductions and/or CDR, and SRM could be phased out as emission reductions and CDR are phased in, to avoid abrupt changes in radiative forcing (Keith and MacMartin 2015).

SRM via marine cloud brightening derives from changes in cloud albedo from injection of aerosol into low-level clouds, primarily over the oceans. Clouds with smaller and more numerous droplets reflect more sunlight than clouds with fewer and larger droplets. Current models provide more confidence in the effects of stratospheric injection than in marine cloud brightening and in achieving scales large enough to reduce global forcing (NAS 2015b).

CDR and SRM have substantial uncertainties regarding their effectiveness and unintended consequences. For example, CDR on a large scale may disturb natural systems and have important implications for land use changes. For SRM actions, even if the reduction in global average radiative forcing from SRM was exactly equal to the radiative forcing from GHGs, the regional and temporal patterns of these forcings would have important differences. While SRM could rapidly lower global mean temperatures, the effects on precipitation patterns, light availability, crop yields, acid rain, pollution levels, temperature gradients, and atmospheric circulation in response to such actions are less well understood. Also, the reduction in sunlight from SRM may have effects on agriculture and ecosystems. In general, restoring regional preindustrial temperature and precipitation conditions through SRM actions is not expected to be possible based on ensemble modeling studies (Ricke et al. 2010). As a consequence, optimizing the climate and geopolitical value of SRM actions would likely involve tradeoffs between regional temperature and precipitation changes (MacMartin et al. 2013). Alternatively, intervention options have been proposed to address particular regional impacts (MacCracken 2016).

GHG forcing has the potential to push the climate farther into unprecedented states for human civilization and increase the likelihood of “surprises” (see Ch. 15: Potential

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Surprises). CI could prevent climate change from reaching a state with more unpredictable consequences. The potential for rapid changes upon initiation (or ceasing) of a CI action would require adaptation on timescales significantly more rapid than what would otherwise be necessary. The NAS (2015a, b) and the Royal Society (Shepherd et al. 2009) recognized that research on the feasibilities and consequences of CI actions is incomplete and call for continued research to improve knowledge of the feasibility, risks, and benefits of CI techniques.

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TRACEABLE ACCOUNTS

Key Finding 1

Warming and associated climate effects from CO2 emissions persist for decades to millennia. In the near-term, changes in climate are determined by past and present greenhouse gas emissions modified by natural variability. Reducing the total concentration of atmospheric CO2 is necessary to limit near-term climate change and stay below long-term warming targets (such as the oft-cited 3.6°F [2°C] goal). Other greenhouse gases (for example, methane) and black carbon aerosols exert stronger warming effects than CO2 on a per ton basis, but they do not persist as long in the atmosphere; therefore, mitigation of non-CO2 species contributes substantially to near-term cooling benefits but cannot be relied upon for ultimate stabilization goals. (*Very high confidence*)

Description of evidence base

The first statement is supported in the literature, including by Joos et al. (2013) and Ciais et al. (2013) (see Box 6.1 in particular), describing the climate response of CO_2 pulse emissions, and further by Solomon et al. (2009), NRC (2011), and Collins et al. (2013), describing the long-term warming and other climate effects associated with CO_2 emissions. Paltsev et al. (2015) and Collins et al. (2013) describe the near-term vs. long-term nature of climate outcomes resulting from GHG mitigation. Myhre et al. (2013) synthesize numerous studies detailing information about the radiative forcing effects and atmospheric lifetimes of all GHGs and aerosols (see in particular Appendix 8A therein). A recent body of literature has emerged highlighting the particular role that non-CO_2 mitigation can play in providing near-term cooling benefits (e.g., Shindell et al. 2012; Zaelke and Borgford-Parnell 2015; Rogelj et al. 2015). For each of the individual statements made in Key Finding 1, there are numerous literature sources that provide consistent grounds on which to make these statements with very high confidence.

Major uncertainties

The Key Finding is comprised of qualitative statements that are traceable to the literature described above and in this chapter. Uncertainties affecting estimates of the exact timing and magnitude of the climate response following emissions (or avoidance of those emissions) of CO_2 and other GHGs involve the quantity of emissions, climate sensitivity, some uncertainty about the removal time or atmospheric lifetime of CO_2 and other GHGs, and the choice of model carrying out future simulations. The role of black carbon in climate change is more uncertain compared to the role of the well-mixed GHGs (see Bond et al. 2013).

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

Key Finding 1 is comprised of qualitative statements based on a body of literature for which there is a high level of agreement. There is a well-established understanding, based in the literature, of the atmospheric lifetime and warming effects of CO_2 vs. other GHGs after emission, and in turn how atmospheric concentration levels respond following the emission of CO_2 and other GHGs.

Summary sentence or paragraph that integrates the above information

The qualitative statements contained in Key Finding 1 reflect aspects of fundamental scientific understanding, well grounded in the literature, that provide a relevant framework for considering the role of CO_2 and non-CO_2 species in mitigating climate change.

Key Finding 2

Stabilizing global mean temperature below long-term warming targets requires an upper limit on the accumulation of CO_2 in the atmosphere. The relationship between cumulative CO_2 emissions and global temperature response is estimated to be nearly linear. Nevertheless, in evaluating specific temperature targets, there are uncertainties about the exact amount of compatible anthropogenic CO_2 emissions due to uncertainties in climate sensitivity, the response of carbon cycle including feedbacks, the amount of past CO_2 emissions, and the influence of past and future non-CO_2 species. (*Very high confidence*)

Description of evidence base

The qualitative statements made in Key Finding 2 are based on evidence synthesized, most notably, by both the National Academy of Sciences (NRC 2011) and by the IPCC (Collins et al. 2013).

Major uncertainties

The NRC (2011) and IPCC (Collins et al. 2013) discuss the uncertainties associated with the Key Finding 2 statement, "The relationship between cumulative CO_2 emissions and global temperature response is estimated to be nearly linear." The ratio of global mean temperature response to cumulative emissions is relatively constant over time and independent of scenario, but the exact magnitude still depends on key assumptions in the future such as climate sensitivity. The IPCC also points out that a constant ratio of cumulative CO_2 emissions to global mean temperature does not hold for stabilization scenarios on millennial time scales and that it is unknown if this constant ratio would hold for scenarios exceeding 2,000 GtC of cumulative CO_2. The other major uncertainties are identified in Key Finding 2.

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

Key Finding 2 is made with *very high* confidence because it consists of qualitative statements that represent fundamental elements of scientific understanding, supported by different literature sources for which there is high agreement.

Summary sentence or paragraph that integrates the above information

The qualitative statements contained in Key Finding 2 reflect aspects of fundamental scientific understanding, grounded in the literature, that provide a relevant framework for considering the role of CO_2 in mitigating climate change.

Key Finding 3

Stabilizing global mean temperature below 3.6°F (2°C) or lower relative to pre-industrial levels requires significant reductions in net global CO_2 emissions relative to present-day values before 2040, and likely requires net emissions to become zero or possibly negative later in the century. Accounting for the temperature effects of non-CO_2 species, cumulative CO_2 emissions are required to stay below about 800 GtC in order to provide a two-thirds likelihood of preventing 3.6°F (2°C) of warming, meaning approximately 230 GtC more could be emitted globally. Assuming global emissions follow the range between the RCP8.5 and RCP4.5 scenarios, emissions could continue for approximately two decades before this cumulative carbon threshold is exceeded. (*High confidence*)

Description of evidence base

Key Finding 3 is a case study, focused on a pathway associated with 3.6°F (2°C) of warming, based on the more general concepts described in Key Finding 2. As such, the evidence for the relationship between cumulative CO_2 emissions and global mean temperature response (NRC 2011; Collins et al. 2013; Allen et al. 2009) also supports key finding 3.

Numerous studies have provided best estimates of cumulative CO_2 compatible with 3.6°F (2°C) of warming above preindustrial levels, including a synthesis by the IPCC (Collins et al. 2013). Sanderson et al. (2016) provide further recent evidence to support the statement that net CO_2 emissions would need to approach zero or become negative later in the century in order to avoid this level of warming. Rogelj et al. 2015 and the IPCC (Collins et al. 2013) demonstrate that the consideration of non-CO_2 species has the effect of further constraining the amount of cumulative CO_2 emissions compatible with 3.6°F (2°C).

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Table 14.1 shows the IPCC estimates associated with different probabilities (>66% [the one highlighted in Key Finding 3], >50%, and >33%) of cumulative CO_2 emissions compatible with warming of 3.6°F (2°C) above preindustrial levels, and the cumulative CO_2 emissions compatible with 2.7°F (1.5°C) are in turn linearly derived from those, based on the understanding that cumulative emissions scale linearly with global mean temperature response (as stated in Key Finding 2). The IPCC estimates take into account the additional radiative forcing effects—past and future—of non-CO_2 species based on the RCP emission scenarios (available here: https://tntcat.iiasa.ac.at/RcpDb/dsd?Action=htmlpage&page=about#descript).

The authors calculated the dates shown in Table 14.1, which supports the last statement in Key Finding 3, based on Le Quéré et al. (2016) and the publicly available RCP database. Le Quéré et al. (2016) provide the widely used reference for historical global, annual CO_2 emissions from 1870 to 2015 (land-use change emissions were estimated up to year 2010 so are assumed to be constant between 2010 and 2015). Future CO_2 emissions are based on the RCP4.5 and RCP8.5 scenarios; annual numbers between model-projected years (e.g., 2020, 2030, 2040, etc.) are linearly interpolated.

Major uncertainties

There are large uncertainties about the course of future CO_2 and non-CO_2 emissions, but the fundamental point that CO_2 emissions need to eventually approach zero or possibly become net negative to stabilize warming below 3.6°F (2°C) holds regardless of future emissions scenario. There are also large uncertainties about the magnitude of past (since 1870 in this case) CO_2 and non-CO_2 emissions, which in turn influence the uncertainty about compatible cumulative emissions from the present day forward. Further uncertainties regarding non-CO_2 species, including aerosols, include their radiative forcing effects. The uncertainty in achieving the temperature targets for a given emissions pathway is in large part reflected by the range of probabilities shown in Table 14.1.

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

There is *very high* confidence in the first statement of Key Finding 3 because it is based on a number of sources with a high level of agreement. The role of non-CO_2 species in particular introduces uncertainty in the second statement of Key Finding 3 regarding compatible cumulative CO_2 emissions that take into account past and future radiative forcing effects of non-CO_2 species; though this estimate is based on a synthesis of numerous studies by the IPCC. The last statement of Key Finding 3 is straightforward based on the best available estimates of historic emissions in combination with the widely used future projections of the RCP scenarios.

FINAL DRAFT