Climate Change and Biogeochemical Impacts

Advanced article

Article Contents

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Based in part on the previous versions of this eLS article 'Climate Change and Biogeochemical Impacts' (2002, 2007).

Human activities are causing a significant build-up of heat-trapping greenhouse gases (e.g. carbon dioxide, methane and nitrous oxide) and aerosols in the atmosphere driven by emissions from fossil fuel combustion, industry, agriculture and deforestation. Atmospheric carbon dioxide is more than 40% above preindustrial levels and growing, and levels would be even higher without modulation by land biosphere and ocean uptake. Model projections for the coming century suggest that these changes in atmospheric composition will result in substantial global warming and strengthening of the hydrological cycle and there is growing observational evidence of a substantial alteration in climate patterns. Carbon cycle (CO_2 and CH_4) feedbacks to climate from the land and ocean can significantly affect future climate, but current projections of these effects are highly uncertain. Climate change and other human-driven processes such as land-use changes and ocean acidification will have profound impacts on global biogeochemistry and terrestrial and marine ecosystems.

Introduction

Over the past two centuries, the composition of Earth's atmosphere, the land surface and ocean have been altered substantially by human activities, including fossil fuel burning, agriculture, deforestation and industrial emissions (Sarmiento and Gruber, 2002). In addition to regional problems, such as higher levels of tropospheric ozone and other forms of air pollution, there is compelling evidence that these changes have already had significant impact on the planet's climate and biota (Doney *et al*., 2012). These human-driven trends are expected to accelerate over the coming century. Considerable research effort is now underway to improve predictions of future climate and ecosystem conditions and to unravel the often complex interactions and feedbacks among climate, atmospheric chemistry, biogeochemistry and ecosystem dynamics under past, present and potential future conditions.

Accurate projections of future climate change and its impact on society and the natural environment hinge on several key questions: how much radiatively active trace gases and aerosols will human activities emit in the future? Given these human emissions, how will atmospheric composition evolve in the future? How sensitive is the climate system to these changes in atmospheric composition? Will carbon cycle feedbacks significantly amplify climate change, by releasing stored carbon in the future, or will marine and terrestrial ecosystems continue to mitigate climate change through uptake of fossil CO_2? The first question involves making predictions of economic and population growth, technology development and political decisions on whether or not and how to regulate emissions (van Vuuren *et al*., 2011). We focus our review on the second, third and fourth questions related to biogeochemistry and climate, and which are, of course, not independent because many of the chemical species of interest have natural biological sources and sinks that respond to climate variations. We will touch on a related and crucial issue, the effect of the expected climate perturbations on society and on natural and managed ecosystems, but note that many aspects extend well beyond the scope of this review. **See also**: Biotic Response to Climatic Change

In recent years, researchers also have become much more aware of how variable the 'natural' climate can be from studies of historical phenomena such as El Niño and longer-term palaeorecords. Dramatic oscillations in climate and atmospheric composition have been observed in the palaeorecord with timescales as short as a few decades to centuries (Petit *et al*., 1999). Thus, future, human-induced changes must be understood against this background of an inherently variable natural system.

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Greenhouse Gases and Aerosols

At the most basic level, the mean surface climate of the Earth is determined by the radiative balance between incoming sunlight and outgoing infrared radiation (Kiehl and Trenberth, 1997).

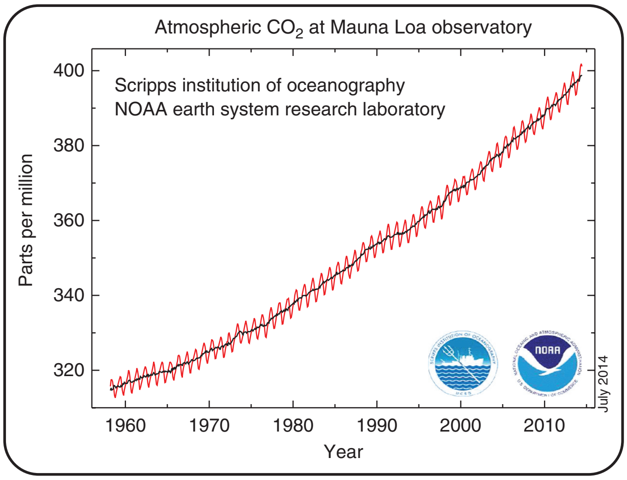

Figure 1 Time series of monthly atmospheric carbon dioxide concentration (ppm) at Mauna Loa. Figure courtesy of Dr. Pieter Tans, NOAA/ESRL (www.esrl.noaa.gov/gmd/ccgg/trends/) and Dr. Ralph Keeling, Scripps Institution of Oceanography (scrippsco2.ucsd.edu/).

Atmospheric trace gases such as carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O) and water vapour (H_2O) play an important role by absorbing infrared radiation and thus trapping heat near Earth's surface much like a blanket. The resulting 'greenhouse effect' is a natural part of the climate system and keeps the planet's average temperature significantly warmer than if there were no atmosphere.

The climate effect of airborne aerosol particles is more complicated, either warming or cooling the surface depending on their albedo, a measure of the particles' brightness. Dark aerosols such as black carbon or soot from biomass burning tend to absorb heat and warm the planet, whereas light sulphate particles produce net cooling because they reflect more solar radiation back to space. Because aerosol particles can also act as nuclei for cloud droplets, aerosol levels have indirect climate effects as well, although these are not as well characterised as the direct impacts.

From a climate perspective, perhaps the best-known human perturbation to the atmosphere is the growing level of carbon dioxide **Figure 1**. Carbon dioxide concentrations in the air have been measured directly since the late 1950s but for time periods before that it can be estimated quite accurately from air bubbles trapped in glacial ice cores (MacFarling-Meure *et al.*, 2006) **Figure 2**. Over the past two centuries, fossil fuel emissions, cement production, deforestation and agriculture **Figure 3** have contributed to an increase in atmospheric carbon dioxide from a preindustrial level of about 280 ppm to over 390 ppm by 2013 (equivalent to about 240 Pg of carbon; 1 Pg = 10^{15} g) (Ciais *et al.*, 2013). A definitive anthropogenic origin for the excess carbon dioxide in the atmosphere can be assigned based on contemporaneous changes in carbon isotopes and by the fact that the atmospheric carbon dioxide levels for the preceding several millennia of the Holocene had hovered within ±5 ppm of the preindustrial value. 'Business as usual' economic and climate scenarios project values as high as 700–1000 ppm by the end of the twenty-first century, levels not experienced on Earth for the past several million years (Pearson and Palmer, 2000). **See also**: **Global Carbon Cycle**

The atmospheric build up of excess carbon dioxide accounts for less than half of the cumulative human carbon emissions over the industrial era (about 550 PgC). For the past decade (2003–2012), for example, a recent analysis (Le Quéré *et al.*, 2014) suggests that only about 45% of the carbon released by fossil fuel emissions (8.6 ± 0.4 PgC year^{-1}) and tropical deforestation (0.9 ± 0.5 PgC year^{-1}) remains in the atmosphere. The difference (4.3 ± 0.1 PgC year^{-1}) is removed in about equal parts by ocean (2.5 ± 0.5 PgC year^{-1}) and terrestrial (2.8 ± 0.8 PgC year^{-1}) drawdown. In fact, without the ocean and land biosphere that serve as important sinks for sequestering such a substantial fraction of present-day anthropogenic carbon emissions, the atmospheric carbon dioxide levels would be even higher. Understanding the processes that control the airborne fraction is critical as it is the amount of excess carbon dioxide left in the atmosphere that directly influences climate. Future projections of atmospheric carbon dioxide levels are relatively sensitive to assumptions about the behaviour of these land and ocean carbon sinks, which are expected to change due to saturation effects and responses to the modified physical climate (Fung *et al.*, 2005; Arora *et al.*, 2013).

The atmospheric levels of other radiatively active species are growing as well **Figure 2**. Changes in land use, agricultural practices and industrial practices have resulted in atmospheric increases in methane (CH_4), nitrous oxide (N_2O) and chlorofluorocarbons (CFCs) (Ciais *et al.*, 2013; Kirschke *et al.*, 2013). Fossil fuel combustion releases nitrogen and sulphur compounds, leading to sharp increases in atmospheric sulphate aerosol levels and deposition of nitrogen to the land and ocean. Tropical biomass burning emits additional black carbon aerosols into the atmosphere, and when deposited in the cryosphere darken snow and ice surfaces, accelerating melting. The greenhouse gases carbon dioxide, nitrous oxide, methane and the CFCs have atmospheric lifetimes of years to centuries and are, therefore, relatively well mixed in the lower atmosphere or troposphere. These gases take time to accumulate in the atmosphere. But, conversely, they can only be removed slowly, and the climatic impact of current greenhouse gas emissions will be with us for a long time to come. By contrast, tropospheric aerosols are removed by precipitation and surface deposition on timescales of a few days to weeks; sulphate aerosols, for example, are thus concentrated near and downwind of a few main industrial source regions in eastern North America, western Europe and east Asia.

In terms of net radiative heating, excess anthropogenic carbon dioxide is presently the most significant, but not overwhelmingly dominant, greenhouse gas (IPCC, 2013). Excess methane, nitrous oxide, CFCs and tropospheric ozone, in aggregate, currently contribute an additional heating equivalent to about 55% of anthropogenic carbon dioxide. Aerosol cooling effects are poorly known and span a large range from minor to significant; mid-range estimates suggest that they may counter globally about a third of the total greenhouse gas warming. But because of carbon dioxide's long atmospheric lifetime and the integral nature of fossil fuel energy sources in the current (and the envisioned, near-term future) global economy, carbon dioxide will almost certainly be the major driving factor for climate change over the next 50–100 years. As global climate models are improved (see

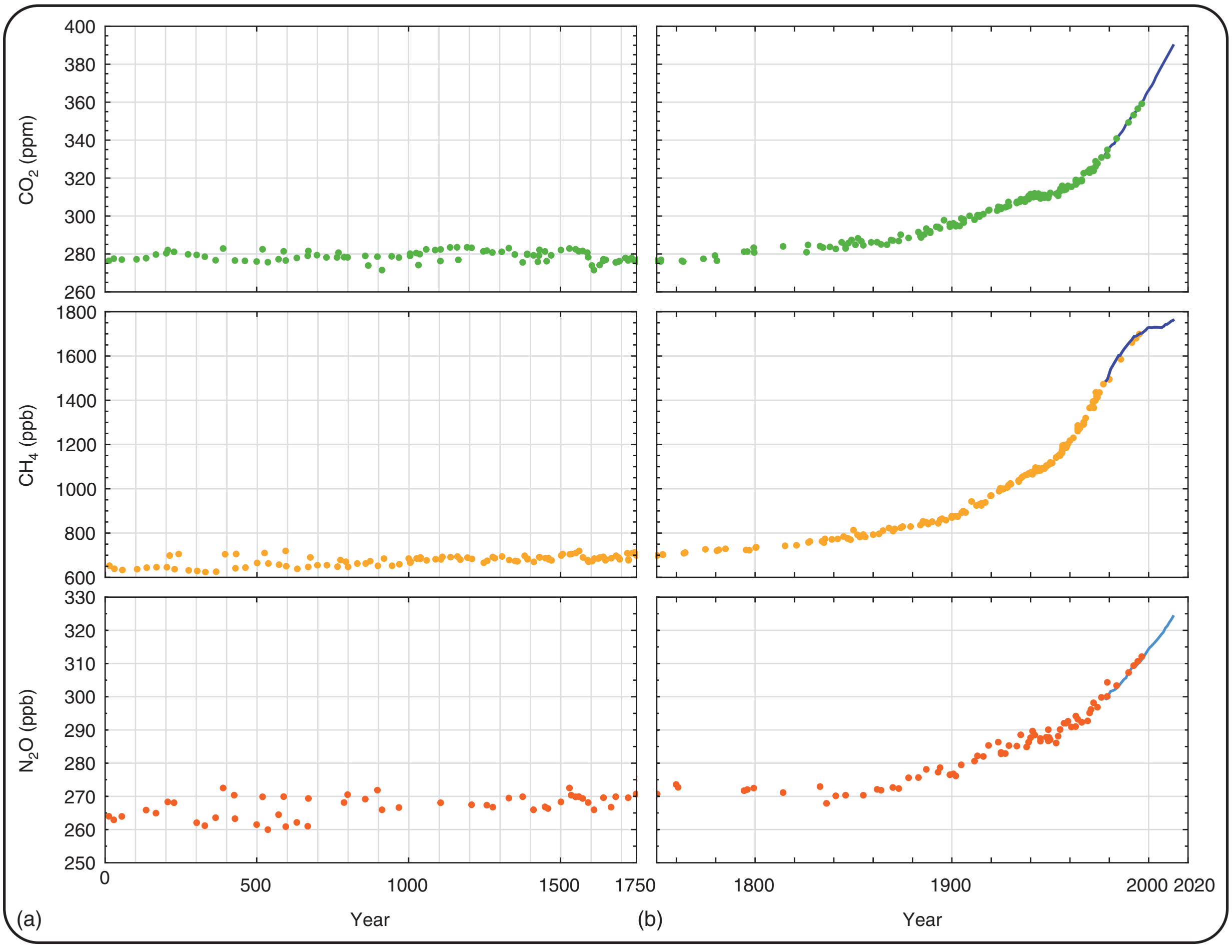

Figure 2 Atmospheric CO_2, CH_4 and N_2O concentration history over the industrial era (b) and from year 0 to the year 1750 (a), determined from air enclosed in ice cores and firn air (colour symbols) and from direct atmospheric measurements (blue lines, measurements from the Cape Grim observatory). Reproduced from Ciais *et al.*, 2013.

later), atmospheric forcing because of trace gases remains one of the main uncertainties associated with future climate projections (Hansen *et al*., 2000; Bodman *et al*., 2013).

The future levels of atmospheric carbon dioxide (and the other greenhouse gases) are poorly constrained in part because carbon emissions depend on changing social and economic factors that are not well known, such as: global population, per capita energy use, technological development and deliberate climate mitigation. Because projecting future emissions from current economic and technological change is so uncertain and controversial, recent climate projections make use of extrapolation from current atmospheric concentrations with higher and lower assumed future concentrations. These scenarios, known as representative concentration pathways (RCPs), are then used to explore both the range of socio-technological futures consistent with those concentrations and the carbon cycle and climate implications (Moss *et al*., 2010). In addition, the human perturbations to carbon dioxide (as well as methane and nitrous oxide) occur on top of a large, natural carbon cycle, a complex system involving ocean, atmosphere and land domains as well as the fluxes between them, which is sensitive to climate variations. **See also**: **Biogeochemical Cycles**

The Climate System

The changes in atmospheric composition induced by humans produce perturbations to the radiative balance of the planet and thus its climate. The radiative warming impacts for the relatively well-mixed greenhouse gases can be calculated using sophisticated one-dimensional radiative transfer models. The direct net cooling effects of aerosols are less well known because of uncertainties in their composition, optical properties and distribution; the indirect aerosol effects on cloud amount and brightness are even more problematic at present. Human and natural modifications in land cover alter climate by changing surface reflectance (albedo), water balance and greenhouse gases. Globally, the impacts are relatively small compared to carbon dioxide emissions, but locally the effects can be large and can lead to

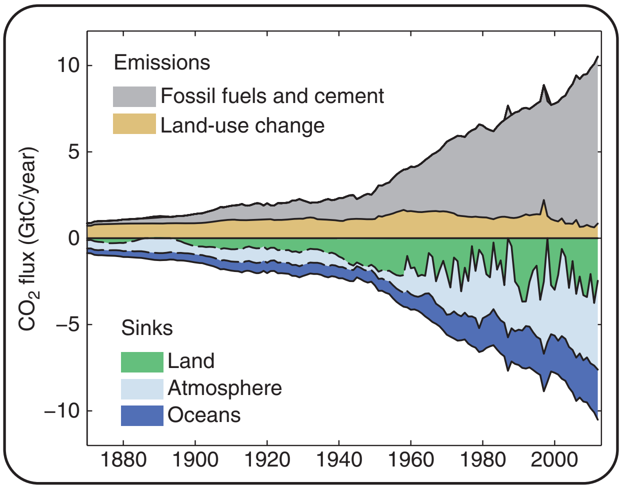

Figure 3 Anthropogenic components of the global carbon budget as a function of time, for (top) emissions from fossil fuel combustion and cement production (grey) and emissions from land-use change (tan) and (bottom) their partitioning among the atmosphere (light blue), land (green) and oceans (dark blue). All time series are in PgC yr^{-1}. Reproduced from Le Quéré *et al.*, 2014.

either net cooling or warming depending on the balance of different factors. The results can be non-intuitive, as illustrated by recent estimates of net cooling over multi-decadal timescales for high-latitude deforestation or forest fires, despite large carbon dioxide emissions (Randerson *et al*., 2006; Bala *et al*., 2007).

For 2011, the recent IPCC report predicts a total, global mean positive radiative forcing from anthropogenic greenhouse gases, aerosols and land use of 1.1–3.4 W m^{-2} relative to preindustrial conditions (IPCC, 2013). The regional heating patterns are non-uniform, with lower heating rates over Northern Hemisphere industrial regions owing to sulphate aerosol cooling. These perturbations will continue to grow, reaching mean values of 3–8 W m^{-2} by the end of this century (IPCC, 2013). **See also**: **Urban Ecology: Patterns of Population Growth and Ecological Effects**

Translating net radiative heating rate into changes in properties relevant for humans and ecosystems (e.g. surface temperature, rainfall, soil moisture, sea level, storm frequency and magnitude) is not straightforward, but there is growing evidence in contemporary observations that these properties are already shifting with time in response to human-driven climate change (IPCC, 2013). Both positive and negative climate feedbacks are involved, including clouds, sea ice, water vapour and the land surface. For example, as the surface ocean warms, greater evaporation rates are leading to elevated atmospheric water vapour that is in turn a greenhouse gas. Ocean warming and altered wind patterns are resulting in large retreat in sea ice concentration and volume regionally, which further accelerates warming because the dark, low albedo water surface can then absorb more sunlight. The largest unknowns at present arise from cloud dynamics, a complicated topic covering an enormous range of time and space scales from individual cloud droplets to tropical convective complexes thousands of miles in extent. There are also the potential for shifts in the climate system that could be triggered by, for example, a slowdown in ocean deep-water formation in the northern North Atlantic or melting of the Greenland and Antarctic ice caps.

Calculating the full impact of the climate interactions requires the use of three-dimensional, coupled Earth System models (Meehl *et al.*, 2006; Hurrell *et al.*, 2013). The models tend to have a range of up to a factor of two in overall climate sensitivity, typically computed as the change in global mean temperature, for the same forcing (IPCC, 2013). Even when the global temperature change is similar, the predicted regional to continental patterns for temperature and precipitation changes, let alone the local environment experienced by ecosystems, can differ significantly and thus should be considered somewhat uncertain, at least in detail. Improving regional near-term (decadal) and long-term (centennial) climate forecasts is a major focus of ongoing research. Those caveats aside, climate models are some of the best tools available at present to address climate change, and there is considerable ongoing efforts to validate model climate sensitivities using target process studies and comparisons against the historical climate record.

Virtually all numerical climate projections for this century show global mean surface temperature increasing, with a range for the end of the century of 2.6–4.8 °C above present-day levels for a high CO_2 emission scenario **Figure 4** (IPCC, 2013). Often a larger temperature range is reported, but this is somewhat misleading as a significant fraction of the variation depends on human behaviour, specifically how much carbon dioxide and other gases we emit to the atmosphere in the future. The lowest temperature projections occur only when emissions are reduced sharply over the next few decades. The largest projected temperature changes are concentrated over the continents and at higher latitudes during the winter season, but some level of warming will occur everywhere. Under a high emissions scenario, global mean sea level is estimated to rise due to thermal warming and melting glaciers and ice sheets by an additional +0.45–0.82 m (+1.5–2.7 ft) by 2100. Many simulations suggest a general strengthening of the hydrological cycle, with higher rainfall in the tropics, drier conditions in the subtropics, increased intensity of tropical and extra-tropical storms and an increased frequency of extreme droughts and heavy precipitation events. Other common features of a warmer climate are more El Niño-like conditions in the Equatorial Pacific, a melt back of polar sea-ice including seasonally ice-free conditions in the Arctic, retreat of glaciers and a slowdown in the formation of ocean deep water at high latitudes. A slowdown of the North Atlantic thermohaline circulation is predicted in a number of models, which could lead to lower warming rates over that ocean basin and western Europe.

Clear alterations in the Earth's climate patterns over the past several decades have already been detected from observations, and the patterns and amplitudes of these changes are inconsistent with natural variability alone but are similar to those predicted from models that include anthropogenic forcing **Figure 5**. The most recent IPCC (2013) assessment concludes that most of the observed climate changes can very likely be attributed to human activities.

On the basis of the historical and palaeoclimate records (IPCC, 2013), the global mean surface temperature in 2012 has increased by 0.85 ± 0.2 °C since the late nineteenth century, and global

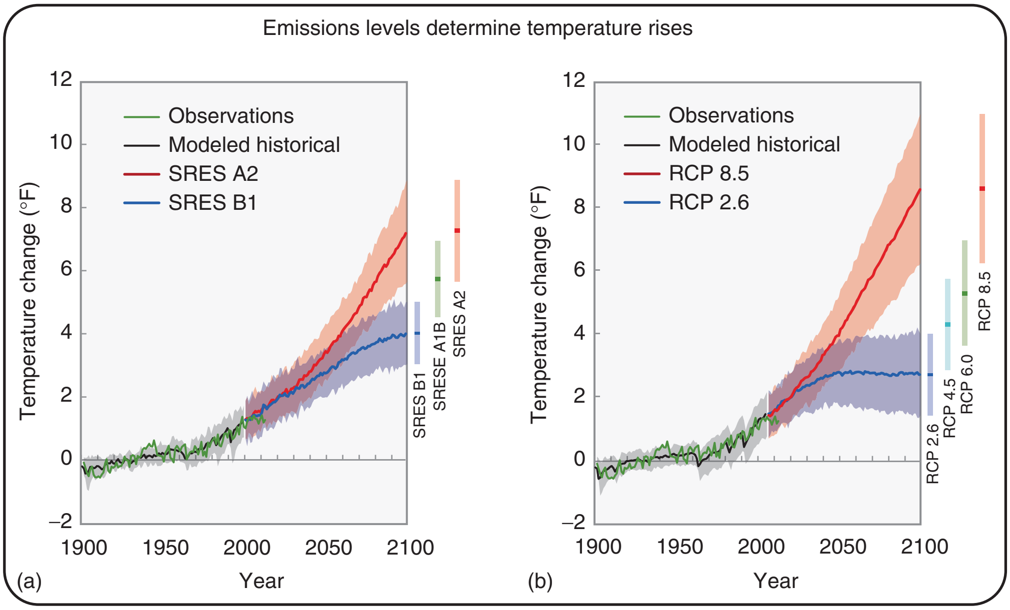

Figure 4 (a,b) Different amounts of heat-trapping gases released into the atmosphere by human activities produce different projected increases in Earth's temperature. Results from (panel a) climate models using two scenarios from the IPCC SRES Special Report on Emissions Scenarios) and (panel b) the most recent generation of climate models (CMIP5) using the most recent emissions pathways (RCPs – representative concentration pathways). Each line represents a central estimate of global average temperature rise (relative to the 1901–1960 average) for a specific emissions pathway. Shading indicates the range (5th to 95th percentile) of results from a suite of climate models. Projections in 2099 for additional emissions pathways are indicated by the bars to the right of each panel. In all cases, temperatures are expected to rise, although the difference between lower and higher emissions pathways is substantial. Reproduced from Walsh *et al.*, 2014.

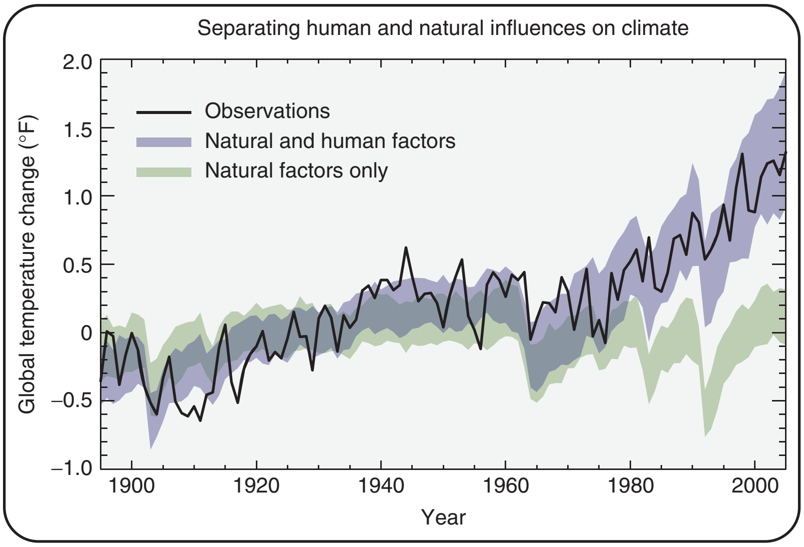

Figure 5 Observed global average temperature changes (black line), model simulations using only changes in natural factors (solar and volcanic) in green and model simulations with the addition of human-induced emissions (blue). Climate changes since 1950 cannot be explained by natural factors or variability and can only be explained by human factors. Reproduced from Walsh *et al.*, 2014 (adapted from Huber and Knutti, 2012 © Nature Publishing Group).

mean surface temperatures have been increasing at a rate of about 0.12 °C/decade from 1951 to 2012. Over the past decade and a half that started with a strong El Nino event in 1998, this rate has slowed to about 0.05 °C/decade, which has been attributed to the effects of decadal climate variability on changes in heat uptake by the oceans (Meehl *et al*., 2011; Kosaka and Xie, 2013). Present temperatures are the warmest on record going back through at least 1000 years, and we will soon be experiencing temperatures warmer than at any time in the past million years (Hansen *et al*., 2006). Subsurface ocean temperatures down to a depth of 3000 m are also on the rise, and average rates of sea-level rise over the past several decades were 1.8 ± 0.5 mm/year, with an even larger rate (3.1 ± 0.7 mm/year) over the most recent decade. Higher precipitation rates are observed at mid to high latitude and lower rates in the tropics and subtropics. Corresponding changes have been measured in surface ocean salinities. Observational data also suggests an increase in extreme weather events associated with the warming and altered hydrological cycle including more frequent heat waves, droughts (in relatively dry subtropical areas) and heavy precipitation events and flooding (in relatively wet temperate and boreal areas).

Systematic declines are also found in the global cryosphere, consistent with surface warming. Observations indicate reductions in sea-ice extent, thickness and volume in the Arctic and along the western side of the Antarctic Peninsula, retreating glaciers, shrinking Greenland and Antarctic ice sheet mass and declining seasonal snow and lake-ice cover. One of the most striking trends is in annual-mean Arctic sea-ice extent, which has declined at roughly 4%/decade from 1979 to 2012 and appears set to continue with time (http://www.arctic.noaa.gov/detect/ice-seaice.shtml). In 2012, the late-summer (September) Arctic ice-cover, the seasonal minimum, was at the lowest level ever recorded with substantial regions of open water off both North America and Eurasia, and some models predict near ice-free summer Arctic conditions by 2040. Recent studies of the Greenland ice sheet highlight an alarming increase in surface melting over the summer, and percolation of that melt water to the base of the ice sheet where the melt water could lubricate ice flow and potentially greatly accelerate ice loss and sea-level rise. The collapse of some of the major ice sheets in the west Antarctic region now appears inevitable, if gradual on century timescales, due to ice sheet melting extending beyond the marine grounding point leading to instability (Joughin *et al*., 2014). Some of these new findings have not been fully incorporated into projected sea-level rise estimates, which thus may be underestimated.

Ocean Biogeochemistry and Marine Ecosystems

The ocean is the largest carbon reservoir that exchanges with the atmosphere on decadal to millennial timescales, and the ocean will serve as the ultimate sink for about 90% of the anthropogenic carbon released to the atmosphere. On societal timescales (a few hundred years), the rate of ocean carbon uptake will depend on ocean circulation and the extent to which climate change alters ocean circulation and the natural marine carbon cycle. The latter effects are due either to evolving ocean physics or to other external forcing factors such as changes in atmospheric nutrient deposition and ocean acidification (Doney *et al*., 2014).

The physical–chemical mechanism for the dissolution of excess carbon dioxide into the ocean is well understood. Carbon dioxide combines with water to form carbonic acid and a series of acid–base products. Seawater, because of its high alkalinity, can hold a significant amount of dissolved inorganic carbon. The natural ocean carbon inventory is about 50 times that of the pre-industrial atmosphere. Higher atmospheric carbon dioxide levels drive a net carbon uptake by the ocean where the rate-limiting step is set by the replacement of surface waters with colder deep waters via ocean circulation.

The present ocean inventory of anthropogenic carbon, the excess dissolved inorganic carbon relative to preindustrial levels, is reasonably well defined based on a global ocean survey conducted over the 1990s (Sabine *et al*., 2004) and follow-on observational studies. A recent estimate of 155 ± 31 PgC for the cumulative uptake to 2010 (Khatiwala *et al*., 2013) is consistent with field-based observations and other constraints including ocean tracers (e.g. CFCs) and numerical models. The ocean's future ability to take up excess carbon, however, is not infinite; it is limited by the changes in ocean chemistry associated with the resulting higher dissolved inorganic carbon levels, the rates of ocean circulation and the alterations in circulation due to climate.

From a simple physics point of view, the changes in ocean circulation associated with global warming will tend to lower ocean carbon dioxide uptake: warmer upper ocean water holds less carbon dioxide than cold water; increased vertical stratification, due to warming at low latitudes and warming and freshening at high latitudes, reduces vertical mixing and exchange, slows high-latitude deep water formation, particularly in the North Atlantic, and decreases ventilation of the deep ocean. The situation in the Southern Ocean, where most of the anthropogenic carbon dioxide uptake occurs, is more nuanced. Reduced stratospheric ozone (the Antarctic 'ozone hole') and climate warming strengthen and contract poleward the band of easterly winds that drive the Antarctic Circumpolar Current. In some ocean model simulations, stronger winds enhance the large-scale upwelling of deep water, which in turn acts to increase the ocean uptake of anthropogenic CO_2 that is primarily limited by the rate of physical transport or exchange between the surface and the subsurface ocean. However, competing effects arise because more overturning also brings up more dissolved inorganic carbon from the deep ocean (Lovenduski *et al*., 2008), where values are elevated because of the respiration of sinking organic matter formed by photosynthesis in surface waters, the so-called biological pump.

When ocean biogeochemistry is included in coupled climate change carbon cycle simulations, future ocean uptake of carbon dioxide is increased somewhat relative to simulations with only physics, but the overall effect is still a reduction in ocean carbon uptake (Fung *et al*., 2005; Le Quére *et al*., 2010). In coupled simulations with simple plankton ecosystem and biogeochemical models, the physical effects are partly compensated by increased uptake from changes in the strength of the natural biological carbon pump. The biogeochemical response is governed by two opposing factors. For example, in the North Atlantic, where mixing is expected to decline, a reduction in the upward nutrient supply due to the increased stratification leads to decreased export production of organic matter and carbon dioxide uptake. At the same time there is a decrease in the upward vertical flux of dissolved inorganic carbon that is elevated in deep waters because of the decay of sinking organic matter from surface phytoplankton growth. The latter factor generally dominates in the present simulations, so that the effect of altered biogeochemistry in the North Atlantic is a net positive carbon dioxide uptake. In the Southern Ocean, stronger upwelling is argued to be decreasing net ocean CO_2 uptake at present, but this may reverse signs in the future when atmospheric CO_2 is so high that stronger anthropogenic CO_2 uptake outweighs stronger outgassing of natural CO_2 from deep waters (Le Quére *et al*., 2010). On a global scale, the combined effect of all these factors is estimated across a range of models as about an 11% drop in the cumulative carbon dioxide uptake over this century (Arora *et al*., 2013).

A wide variety of other mechanisms have been identified that could conceivably alter ocean carbon uptake, and not all of these processes are included in the current generation of Earth System models. However, in many cases, even the sign of the biogeochemical response, let alone the quantitative magnitude, is uncertain (Boyd and Doney, 2003). The changes in ocean physics already mentioned including warming and stratification are expected to alter patterns of marine productivity and reduce subsurface ocean levels (Bopp *et al*., 2013). Other environmental factors such as aeolian deposition of trace metals via dust, cloud cover and solar and UV irradiance, riverine and atmospheric nutrient deposition to the coastal ocean and ocean carbonate chemistry are all sensitive to climate change. A common feature of many scenarios involves shifts in planktonic community structure and the decoupling of the carbon cycle from that of the major ocean macronutrients. Specific hypotheses include elevated iron fertilisation rates of HNLC (high nitrate, low chlorophyll) regions, particularly in the Southern Ocean, higher subtropical nitrogen fixation rates, altered elemental stoichiometry of exported particulate and dissolved organic material and changes in subsurface remineralisation.

Global change impacts on the ocean extend beyond simply the responses to changes in physical climate. The excess anthropogenic carbon dioxide in the upper ocean combines with water to form carbonic acid leading to ocean acidification (Orr *et al*., 2005). Surface water pH values have already dropped by about 0.1 units from preindustrial levels and are expected to drop by an additional 0.3 units by the end of the twenty-first century (Feely *et al*., 2009). Lower pH and lower carbonate ion concentrations result in reduced calcification rates by a host of shell-forming organisms including surface and deep-water corals, many phytoplankton and zooplankton, and pteropods (marine snails) (Doney *et al*., 2009). Many of these organisms provide critical habitat and/or food sources for higher trophic levels and the overall impact of ocean acidification may be quite significant.

The Terrestrial Biosphere

Current research suggests that, on balance, terrestrial ecosystems are also a net sink for carbon dioxide at the present time. But this conclusion masks considerable complexity and uncertainty both in present fluxes and with respect to future behaviour (Schimel *et al*., 2001). Deforestation and biomass burning result in net carbon dioxide fluxes to the atmosphere, as high-carbon forests are turned into comparatively low-carbon pastures and croplands. Some regrowth is also occurring, which may partly offset deforestation's impact on atmospheric CO_2 (Pan *et al*., 2011). Deforestation is now occurring mainly in the tropics and is partially countered by temperate regrowth on abandoned farm and pastureland. In the Northern Hemisphere, regrowth of forests on former farmland (especially in the eastern United States) and in previously harvested areas causes a sink during the rapid growth phase of young forest growth. On centennial timescales, climate change may also alter the large-scale distribution of natural biomes over the planet, with significant impacts on carbon storage; for example, widespread replacement of forests with grasslands would tend to accelerate the atmospheric rise in carbon dioxide. **See also**: **Deforestation, Forest Management and Governance**

Other terrestrial processes may also cause net carbon drawdown. Fertilisation of photosynthesis by higher atmospheric carbon dioxide levels increases plant growth, in many cases substantially, and contributes to carbon uptake. Experimental studies demonstrate a significant effect of this mechanism, which, when included in global models, suggests a current sink that may continue for many decades into the future. A recent analysis, based on combining the global carbon budget with process model estimates of the CO_2 effect, suggests a global sink of >2 Pg year^{-1}, although this estimate is conservative because some of the 2 Pg may be due to climate, nutrient deposition and other processes (Schimel *et al*., 2014). Laboratory- and large-scale free-air carbon dioxide enrichment (FACE) experiments show a consistent effect of increasing carbon dioxide (Norby *et al*., 2005), although recent studies show that this experimental effect may be smaller than earlier thought (Long *et al*., 2006). Many Northern Hemisphere ecosystems are also nitrogen limited and deposition of reactive nitrogen mobilised by fossil fuel combustion may also cause increased carbon uptake (Lamarque *et al*., 2005). If the recent (Schimel *et al*., 2014) estimate is correct, that suggests a substantial negative or stabilising feedback from terrestrial ecosystems that could play an important role in the future.

The best available evidence suggests that warming causes land ecosystems to lose carbon because respiration is more temperature sensitive than photosynthesis and because of the effects of growing water stress in warmer climates. This effect is particularly important in the tropics, where warm temperatures and dry conditions appear to act synergistically to reduce carbon uptake (Wang *et al*., 2014) and increase forest mortality (Saatchi *et al*., 2013). Future warming could also release carbon from high-latitude northern regions as carbon currently preserved in permafrost is liberated. There is a wide range of estimates for how sensitive carbon stocks are to climate, with many models and data analyses suggesting significant effects (Cox *et al*., 2013; Arora *et al*., 2013). Larger effects could come if future climates lead to additional disturbances, especially fire, which could rapidly release large amounts of carbon (Van der Werf *et al*., 2010). Current attention is focusing on the balance between human- and climate-caused disturbances in controlling future ecosystem carbon storage as well as on physiological processes, such as carbon dioxide fertilisation.

As noted earlier, while early work focused on the effects of temperature on terrestrial carbon storage, changes in water balance and water available to support biological activity are now thought to be as significant (Wang *et al*., 2014). Models (Fung *et al*., 2005; Friedlingstein *et al*., 2006) and observations (Angert *et al*., 2005) show that decreasing water balance (drier soil conditions) decreases carbon uptake. The water balance can decrease either (1) when precipitation declines or (2) when evaporative demand from a warmer atmosphere increases, so that soil moisture can decline even if precipitation does not change. Analyses of coupled carbon–climate models suggest that the global effect tends to be a competition between warmer conditions favouring a longer growing season in the Northern Hemisphere, increasing carbon uptake, and drier soils in the tropics, decreasing carbon uptake.

Recent evidence shows that wildfire plays a major role in the global carbon (Van der Werf *et al*., 2004; Moritz *et al*., 2012). The increases in the growth rate of atmospheric carbon dioxide concentrations during El Niño years, long a subject of debate, now appear to be dominated by increased tropical fire associated with El Niño droughts. Evidence for this includes the location of the enhanced emissions in El Niño years (tropical land regions coincident with drought-affected areas), isotopic and chemical evidence, especially from correlations between carbon monoxide (a fire by-product) and carbon dioxide and direct satellite detection of burning plumes. These observational studies, taken together with coupled model results projecting future increases in drought with climate change, and dynamic vegetation models suggesting more fire with drier conditions, suggest a growing role for fire with climate change.

Future Research Directions

Historically, carbon cycle research of the land, ocean and atmosphere domains has occurred in a disjoint, disciplinary manner. A growing consensus, however, argues that there are strong commonalities in terms of both research techniques and information content across the domains. In essence, the 'carbon cycle is one' and is best studied through a coordinated, multi-pronged research effort. The recent launches of GOSAT and OCO-2, satellites that measure CO_2 in the atmosphere, may emphasise this unity between human, terrestrial and marine aspects of the carbon cycle. An integrated, global research framework also provides a more natural route for communicating the research results with policy makers and local to regional stakeholders who will likely be impacted by climate change (Dilling *et al*., 2003).

A number of specific scientific paths exist for improving our predictive capability with respect to climate-induced biogeochemical responses. Numerical models of terrestrial and marine biogeochemistry are pivotal and are being improved progressively as better, data-based mechanistic parametrisations are developed and evaluated. One important, emerging issue over the next few years will be matching (resolving or parametrising) within models the relevant scales (time and space) of the key biological processes. A hierarchy of models is clearly warranted incorporating both forward (prognostic) and inverse approaches. Data assimilation techniques, in particular, provide a dynamically consistent framework for melding models and observations and will provide important estimates of poorly constrained model parameters, process sensitivities and net surface fluxes and rates (Raupach *et al*., 2005; Baker *et al*., 2010). An important measure of the skill of the numerical models will be their ability accurately to hindcast observed responses to natural climate variability on timescales from the seasonal cycle to multiple decades.

Historical climate variability can, under some circumstances, directly serve as a useful analogue for future climate change (Cox *et al*., 2013). Biological time series exhibit significant variability on inter-annual to inter-decadal scales associated with physical climate phenomena such as the El Niño Southern Oscillation (ENSO) and other climate modes. Some theoretical, data analysis and modelling studies argue that climate change may project significantly on to existing natural variability modes, and thus one could infer the biogeochemical climate change response by extrapolating present distributions during climate mode extrema. The projected physical climate change for this century, however, greatly exceeds observed natural variability, and the ecosystem response to physical forcing may be timescale dependent and quite non-linear, manifested as biological regime shifts (Lenton, 2011).

The palaeoclimate record offers another valuable resource, yielding information about the state of global biogeochemistry under climate perturbations of similar magnitude (though not rate of change) to projected anthropogenic effects and to a variety of climatic conditions, both colder and warmer than today. Perhaps the best documented palaeo period is the Last Glacial Maximum (20 Ka), when atmospheric carbon dioxide levels recorded in Antarctic glacial ice were 80–90 ppm lower than during the preindustrial Holocene (Petit *et al*., 1999). The data constraints on pre-Quaternary palaeoclimate are considerably weaker, but the warm periods of the Late Palaeocene thermal maximum (55 Ma) and the Cretaceous (100 Ma) may be relevant as a window, albeit extreme, on biogeochemical response under global warming scenarios (Pearson and Palmer, 2000; Doney and Schimel, 2007). **See also**: **Ice Ages**; **Quaternary Life on Land**

Finally, long-term monitoring and field process studies are required for ongoing biogeochemical model development and evaluation. The number of historical, multi-decadal time series on either the ocean or land is limited, but their utility is almost unrivalled (Baldocchi, 2014; Ducklow *et al*., 2009). A coordinated measurement programme is needed that combines satellite remote sensing of the ocean and land surface, repeat large-scale land and ocean surveys, point time-series observations, direct surface trace gas flux measurements and an expanding atmospheric monitoring network including recently launched satellites that will be able to map the spatial distribution of carbon dioxide. New quasi-autonomous sampling platforms and physical/chemical sensors will be required. Wherever possible, the integrative nature of the atmosphere should be relied upon to smooth out the small-scale heterogeneity in land and ocean processes and reduce uncertainties associated with the severe undersampling in many areas. New remote sensing technologies also increasingly allow measurement of key carbon cycle controls, such as soil moisture, biomass and terrestrial and marine plant species composition. In order to both detect carbon cycle changes and attribute them to specific mechanisms acting over large areas, improvements in remote sensing are key. In situ and remote monitoring efforts should be augmented by focused process and manipulative studies.

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