CHAPTER

3.4

The Greenhouse Effect, Aerosols, and Climate Change

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3.4.1 FUNDAMENTALS

Over the past 50 years, the earth's climate has changed at a rapid pace relative to the observed climate variability of the past 5000 years. The global mean surface air temperature has increased by about 0.8 °C, sea level has risen by about 12 cm, and precipitation patterns have shifted. This global warming and associated climate change has been driven by increasing concentrations of gases that absorb and emit infrared radiation in the atmosphere, principally carbon dioxide and methane. The warming has been reduced to some degree by increasing aerosol pollution, which has its own associated impacts on precipitation.[1] Continued emissions of greenhouse gases at present rates can be expected to produce continued warming, which would also continue after emissions have ceased, until thermodynamic equilibrium is achieved. In this chapter, we will discuss the chemistry of these gases and particles, focusing on the processes that determine their lifetime in the atmosphere and their fate when they are removed from the atmosphere, and then review the physics of climate, focusing on the role of greenhouse gases and aerosols. We will also discuss some proposed technological interventions to remove greenhouse gases from the atmosphere or counteract their impact on climate.

Fig. 3.4.1 shows the annual mean global mean surface temperature change from 1880 through 2016, as a difference from the average over the years 1950–80, on the left and the rise in greenhouse gas concentration since the Industrial Revolution on the right. There is nothing particularly surprising about the agreement between the two figures: the idea that increases in carbon dioxide would lead to warming of approximately the observed magnitude has been around for over a century. However, the complexity of the climate system, and the presence of year-to-year and decade-to-decade ups and downs in the observed temperature record makes prediction of the exact magnitude of future change for a given amount of greenhouse gas emissions difficult to predict in advance, and the accompanying changes to precipitation and sea level have also required much scientific work to monitor, explain, and predict.

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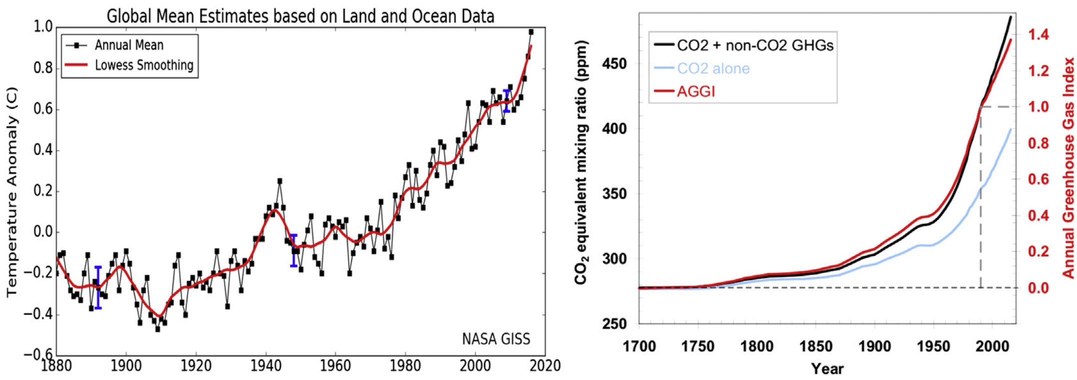

FIGURE 3.4.1 Left: Global average near-surface air temperature, 1880 to present, expressed as the difference from the temperature averaged from 1950 to 1980. The solid *black line* is the global annual mean and the solid *red line* (gray in print versions) is the 5-year LOWESS smoothed curve. The blue (dark gray in print versions) uncertainty bars (95% confidence limit) account only for incomplete spatial sampling.[38,39] Right: In red (gray in print versions), the annual greenhouse gas index, the ratio of the total direct radiative forcing due to long-lived greenhouse gases for any year for which adequate global measurements exist to that which was present in 1990.[40]

3.4.2 SOURCES AND SINKS OF GREENHOUSE GASES

Atmospheric composition changes that influence climate can be divided into two main categories: changes in molecular gases that interact with infrared radiation and changes in aerosols (small liquid or solid particles) that absorb or reflect visible radiation. In addition, ozone (O_3), which forms naturally in the atmosphere from molecular oxygen (O_2), interacts with both ultraviolet and infrared radiation, and its concentration is controlled by chemistry that has been changed substantially by human emissions of various pollutants.

Concentrations of CO_2, CH_4, and N_2O have all been rising rapidly since before 1900 and are now more abundant than their preindustrial levels by factors of 1.4, 2.5, and 1.5, respectively.[2] The impact of human emissions of these gases on climate depends on two main factors: the strength of interaction of gases with infrared radiation and the lifetime of gases in the atmosphere. For example, a ton of CH_4 added to the atmosphere increases the atmosphere's interaction with infrared radiation much more strongly than a ton of CO_2 added to the atmosphere [mostly because CH_4 is only present at a concentration of 1.8 parts per million by volume (ppmv), whereas CO_2 is present at about 400 ppmv], but CO_2 is much more persistent, so the overall impact of the CH_4 on climate will be about 110 times larger than that of the CO_2 in the first year after emission, about 30 times larger 40 years after emission, and only about 4 times larger 100 years after emission, when most of the CH_4 will have been lost to chemical processes.[3]

Anthropogenic greenhouse gas emissions arise from a range of human activities: industrial, agricultural, and domestic. Carbon dioxide emissions derive primarily from combustion of fossil fuels, in nearly equal parts each from burning coal, oil, and natural gas,[4] and secondarily from deforestation. Deforestation results in emission of carbon dioxide (and methane) because much of the woody biomass cut down when forests are cleared for conversion to agricultural land is either burnt or decomposes, converting carbon stored as cellulose in

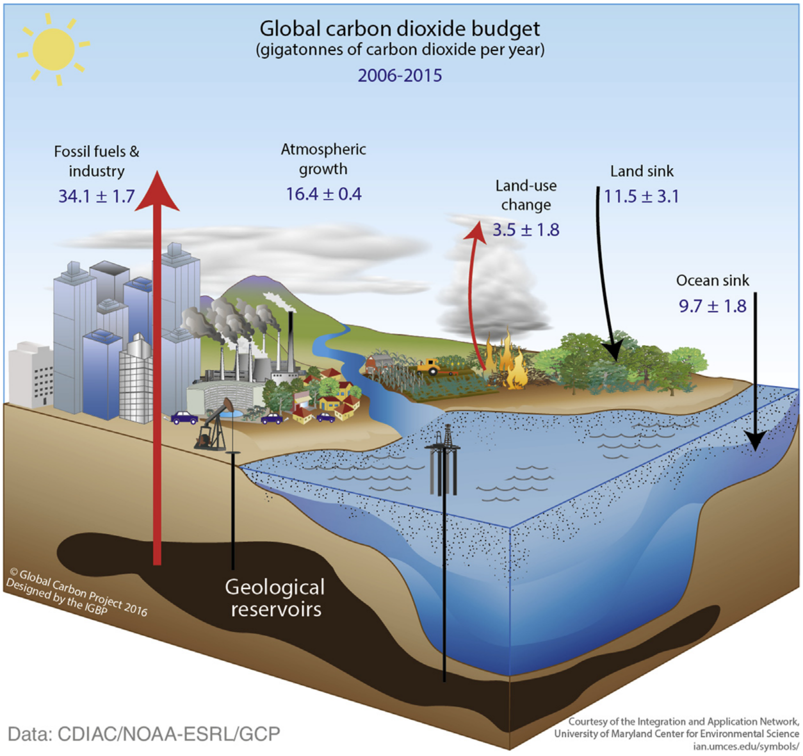

FIGURE 3.4.2 Fluxes of CO_2 averaged over the past decade.[41] Burning of oil, coal, and natural gas is the predominant cause of increasing atmospheric CO_2, with some additional CO_2 added by deforestation, especially in tropical rainforest areas. Of the CO_2 emitted from this combustion and deforestation, a bit less than half remains in the atmosphere, whereas the rest is either absorbed by the oceans or is taken up in forests and soils.

wood into carbon dioxide. In addition, chemical processes such as cement manufacturing, in which $CaCO_4$ is heated to form CaO_2, release CO_2. Fig. 3.4.2 shows the relative contribution of these sources to global CO_2 emissions. Methane emissions arise from fermentation either in the rumens of livestock or in anoxic agricultural environments such as rice paddies and manure piles or human waste treatment facilities, from leakage of natural gas (which is mostly methane) from wells and pipelines, and from burning of forests and fields. Methane emissions and concentrations rose exponentially through the 20th century along with the population, but leveled off in the late 20th century with improvements in rice farming practices that involved briefer periods of flooded fields, and reduction in leakage from natural gas drilling, before resuming their increase in the 2010s as these technological improvements were completed while population and wealth continued to increase.[5] Nitrous oxide emissions arise principally from soil bacteria, and are enhanced by the heavy use of nitrogen fertilizers.[6]

Greenhouse gases, once emitted to the atmosphere, do not remain there forever. We define the "atmospheric lifetime" of a chemical in the atmosphere as follows: if we add an increment of that chemical to the atmosphere so that its concentration increases from x kg/m^2 to $(x + y)$ kg/m^2, then the atmospheric lifetime is the time it would take for the concentration of that chemical to decrease back to $(x + ye^{-1})$ kg/m^2. In other words, the atmospheric lifetime of a chemical is the time it takes for a perturbation to the atmospheric concentration of that chemical to decrease by a factor of *e*. The three major greenhouse gases have rather different atmospheric lifetimes.

N_2O largely is nonreactive in the atmosphere below about 20 km, but it is destroyed by ultraviolet radiation (and secondarily by reaction with atomic oxygen, O, when it mixes into the stratosphere). This fairly slow removal mechanism gives it a fairly long atmospheric lifetime of about 116 years.[7]

CO_2 is subject to dissolution in the oceans and to uptake into those parts of the biosphere where carbon-based substances are accumulating (e.g., land areas that were cleared for agriculture in the past, but are now returning to forest). The fraction of CO_2 that remains in the atmosphere, making up the increment in concentration each year, had been close to 46% of what is emitted each year for the past several decades, with the remaining 54% being split between absorption in the ocean and in the land biosphere. There is much ongoing research about whether this fraction will remain constant over time—a change in the fraction remaining in the atmosphere would have major implications for how much CO_2 can be emitted in the future while preventing the global mean temperature from exceeding some threshold.[8]

The initial oceanic uptake of carbon dioxide takes places very quickly. One can think of the ocean mixed layer as simply a second box, adjacent to the atmosphere into which CO_2 flows. The amount of CO_2 that flows into this "box" is limited by the fraction of the ocean that turns over on the time scale of a year or so. CO_2 in the ocean quickly reacts with water in the following two reactions[9]:

$$CO_{2(aq)} + H_2O \rightleftharpoons HCO_3^- + H_3O^+$$

$$HCO_3^- \rightleftharpoons CO_3^{-2} + H_3O^+$$

The concentration of CO_2, HCO_3^-, and CO_3^{-2}, referred to collectively as dissolved inorganic carbon (DIC), is close to equilibrium with atmospheric CO_2, so as atmospheric CO_1 rises, the concentration of DIC and the concentration of hydronium ions (H_3O^+) increases. This implies an increase in the acidity of the ocean, as has been observed (Fig. 3.4.3). Thus the fraction of CO_2 emitted by people that fills the ocean mixed layer "box" does not influence the earth's climate, so the ocean absorption reduces the climate impact of a given amount of CO_2 emitted. However, the acidification of the ocean accompanying increasing DIC has its own set of ecological consequences, since it makes it more difficult for the microscopic animals and plants that form the base of the ocean food chain to make their shells.[10] This latter process leads to a slow sink of carbon from the ocean mixed layer to the ocean sediment: some fraction of those bits of plankton and dead plants and animals that are not decomposed and recycled within the mixed layer may eventually settle to the ocean floor and be buried, removing the carbon they contain completely from the ocean-atmosphere system, a process referred to as the "biological pump." If human emissions of carbon dioxide stop completely, over thousands of years this process will return atmospheric CO_2 concentrations to preindustrial levels.[1]

On land, CO_2 can be taken up in woody biomass (trees and bushes) or in soils (leaf litter that accumulates without decomposing completely), whereas it can be released when forests

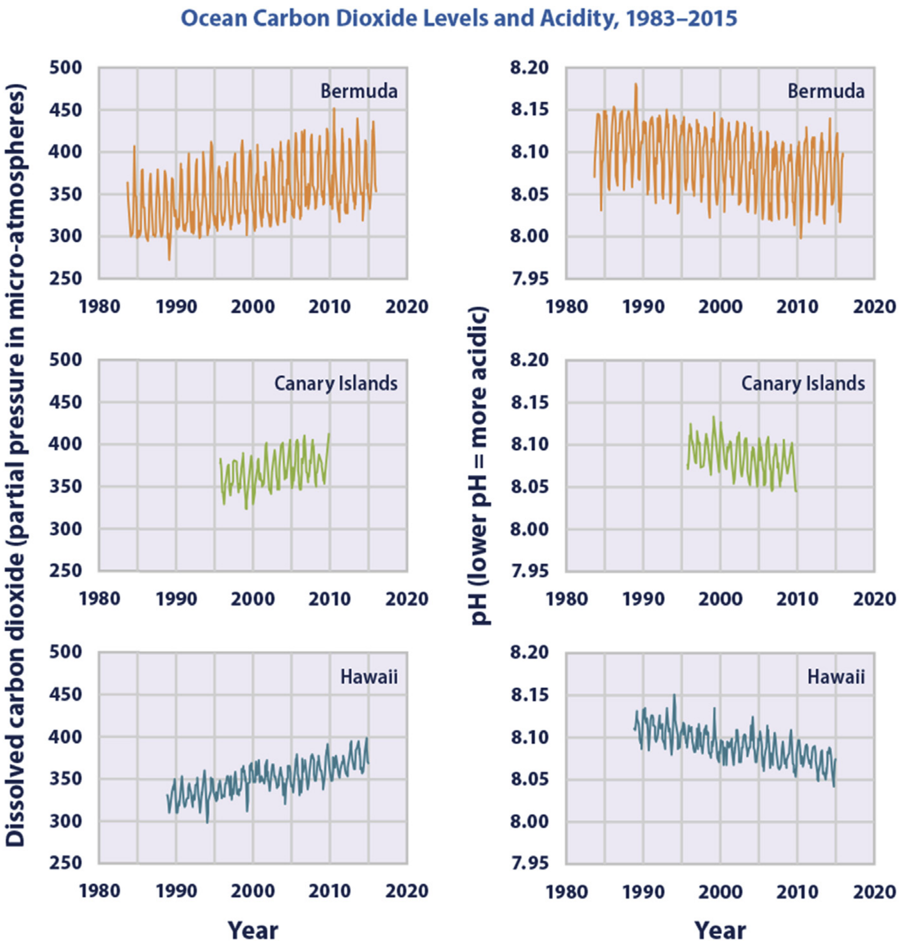

FIGURE 3.4.3 Direct observations of dissolved carbon dioxide concentration and pH of seawater sampled and the locations indicated in each graph. Trends toward greater acidity (lower pH) and higher carbon dioxide concentration are visible in each location.[42] A strong seasonal cycle is also present, associated with the seasonal cycle in atmospheric concentration of CO_2.

are burned or cut and allowed to rot, when organic soils are tilled and exposed to air, or when frozen peatlands thaw. While the tropical landmasses are significant net sources of CO_2 due to forest clearing (South America: 1.8 Pg CO_2/year, Africa: 0.4 Pg CO_2/year, Asia: 2.0 Pg CO_2/year), North America is a significant net sink (1.7 Pg CO_2/year) due to regrowth of forests cut down in the 19th and 20th centuries and due to growth of existing forests partly because of fire suppression.[11]

CH_4, unlike CO_2, is subject to chemical loss in the atmosphere. Most importantly, reaction with the OH (hydroxyl) radical yields CO_2 and water. This reaction yields an atmospheric lifetime for CH_4 of only about 10 years. Thus sustained high concentrations of CH_4 over more than a few decades requires sustained high emissions of CH_4.[5]

3.4.3 AEROSOLS AND CLIMATE

Volcanic aerosol emissions are a powerful natural driver of year-to-year and decade-to-decade variations in climate. Volcanoes emit sulfur dioxide gas (SO_2), which reacts with water in the atmosphere to form sulfuric acid (H_2SO_4). When volcanic plumes are emitted powerfully enough to reach the stratosphere,[a] the H_2SO_4 can form a persistent haze of liquid droplets, reflecting away sunlight and cooling the earth for a year or two. Human activities can have a similar cooling effect. Coal tends to contain substantial amount of sulfur, so that burning of coal for heat and power releases SO_2. Such pollution does not reach the stratosphere, so the added SO_2 has a fairly short atmospheric lifetime; however, the mass of emissions is large enough that a substantial cooling effect by H_2SO_4 droplets results. Burning of fossil fuels and of forests and agricultural fields results in large amounts of particle emission consisting of complicated carbon compounds. All these particles can also act to nucleate cloud particles, changing the lifetime and radiative properties of clouds.

The net effect of these anthropogenic (human-caused) aerosols is to cool the planet, but the complexity of their interactions with clouds, and their high variability due to their short lifetime, makes their quantitative effect on climate highly uncertain. They are estimated to cause a cooling about one-third as strong as the warming due to greenhouse gases, but the range of uncertainty is anywhere from zero to two-thirds of the greenhouse gas warming.[1]

3.4.4 PHYSICS OF CLIMATE

The climate of the earth is constrained by the earth's energy exchange with the universe. The earth receives essentially all[b] its energy from the universe in the form of ultraviolet, visible, and infrared radiation from the sun and returns energy to the universe in the form of infrared radiation. Because the energy received from the sun varies by less than 0.15% from year to year (most variability being associated with the approximately 11-year-long sunspot cycle),[12] the earth's climate is typically close to equilibrium: the annually averaged incoming and outgoing energy are equal to within about 1%. The bulk of yearly averaged net outward or inward energy transfer is caused by changes in atmospheric composition. These changes are large enough to be directly observed in satellite observations of the earth's outgoing radiative flux following volcanic eruptions, when increased aerosol concentrations in the stratosphere result in increased reflection of solar radiation to space.[13] For the somewhat smaller, but sustained and therefore more consequential, radiative response to increasing greenhouse gas concentrations, the energy imbalance can be calculated from the increase in heat stored in the oceans[14] and from radiative transfer models of the atmosphere.[15]

[a]The stratosphere is the layer of the atmosphere above 12–18 km in which the temperature rises with height, due to the absorption of solar ultraviolet radiation by ozone. This temperature structure suppresses strong vertical motions, so that air that reaches the stratosphere tends to remain there for a few years. Air enters the stratosphere in the tropics and returns to the troposphere (the layer between the stratosphere and the ground) in the polar regions.

[b]The next largest source of energy is the moon, from which the earth never receives more than about $1/(5 \times 10^4)$ as much energy as it receives from the sun (and that only when the moon is full).

3.4.4.1 Radiative Balance

Consider a small metal sphere orbiting the sun at about the earth's distance from the sun. We are using a metal sphere because it conducts temperature well, so we can assume that it has a single temperature. Its energy budget could be expressed as the change in heat energy resulting from the difference between the absorption of energy from the sun and the loss of energy by radiation emitted by the sphere:

$$\frac{4}{3}\pi r^3 \rho C \frac{dT}{dt} = (1-\alpha)S\pi r^2 - 4\pi r^2 \varepsilon \sigma T^4 \quad (3.4.1)$$

where ρ is the density of the metal, C is its heat capacity per unit mass, r is the radius of the sphere, T is its temperature, t is time, S is the flux of radiation from the sun that would fall on a plane facing the sun, α is the albedo of the sphere (the fraction of solar radiation reflected away from it, equal to 1 for a white sphere and 0 for a black sphere), σ is the Stefan-Boltzmann constant, and ε is the emissivity of the sphere (equal to 1 for a perfect blackbody).

Simplifying, we get:

$$\frac{r}{3}\rho C \frac{dT}{dt} = (1-\alpha)\frac{S}{4} - \varepsilon \sigma T^4 \quad (3.4.2)$$

If we assume the sphere's temperature starts at zero, the second term on the right-hand side of Eq. (3.4.2) will also be zero, so the right-hand side will be positive and the temperature will start to rise due to the absorption of solar radiation. As its temperature rises, the sphere will radiate ever more energy according to the second term on the right (the Stefan-Boltzmann law). Eventually the sphere will approach equilibrium, when the temperature will change only infinitesimally over time, so that the left-hand side is near zero. Then we can write

$$(1-\alpha)\frac{S}{4} = \varepsilon \sigma T^4 \text{ or}$$

$$T = \left[(1-\alpha)\frac{S}{4\varepsilon\sigma}\right]^{1/4} \quad (3.4.3)$$

If we assume that the sphere can radiate as well as a blackbody (so that $\varepsilon = 1$), and has the same albedo as the earth (so that $\alpha = 0.3$), and noting that the flux of radiation falling on a plane facing the sun at the earth's distance from the sun is $S = 1360\ \mathrm{W/m^2}$, we get a temperature $T = 255\mathrm{K} = -18\ ^\circ\mathrm{C}$, which is in the ballpark of the earth's temperature: it is the temperature of the air near 5 km elevation in the atmosphere.

However, the average temperature at the earth's surface is about 15 °C or 288K. What explains the difference? The peak wavelength of radiation emitted by a blackbody can be derived from Planck's law to be approximately $\lambda = 2900\ \mu\mathrm{m\,K}/T$, which gives about 11 μm for radiation emitted by our imaginary sphere. Substances like rock and water have emissivities of about 0.9, whereas snow, ice, and vegetation have emissivities of 0.97 or higher at wavelengths near 11 μm,[16] but even plugging a value of $\varepsilon = 0.9$ into Eq. (3.4.3) gives a temperature of just 260K. Assuming the earth's actual temperature and solving for the emissivity gives $\varepsilon = 0.61$. What is it that reduces the ability of the earth's surface to radiate infrared radiation to this level, resulting in its higher temperature?

The answer turns out to be the atmosphere. Because there are substances in the atmosphere, including water vapor, clouds, and a number of trace gases, that absorb and emit infrared radiation, an instrument placed outside the earth's atmosphere and looking toward the earth's surface will measure radiation that, depending on its wavelength, arises in part from the surface of the earth and in part from gases and clouds at various levels in the atmosphere. The average temperature at these levels in the atmosphere, weighted by the radiation they emit, corresponds approximately to the temperature we derived above, 255K.

One way to build intuition about this concept is to consider a two-level model of the earth's climate.[c] We can approximate the earth's surface as a perfect emitter (a blackbody) with $\varepsilon = 1$ and the earth's atmosphere as a partially absorbing layer with a finite emissivity, $\varepsilon = \varepsilon_a$. If we assume that the atmosphere is transparent to radiation from the sun, except for some that is reflected away by clouds and does not act to warm the climate, we can write two new energy balance equations:

$$\frac{4}{3}\pi r^2 D \rho_s C_s \frac{dT_s}{dt} = (1-\alpha)S\pi r^2 - 4\pi r^2 \sigma T_s^4 + 4\pi r^2 \varepsilon_a \sigma T_a^4 \tag{3.4.4}$$

$$4\pi r^2 H \rho_a C_a \frac{dT_a}{dt} = 4\pi r^2 \varepsilon_a \sigma T_s^4 - 2 \times 4\pi r^2 \varepsilon_a \sigma T_a^4 \tag{3.4.5}$$

The terms on the right-hand sides of these equations are shown in the diagram on the left of Fig. 3.4.4. Eq. (3.4.4) represents the energy budget of the earth's surface: any positive change in the thermal energy of the surface (taken here to be a layer of water of depth D, density ρ_s, and heat capacity C_s spread on a sphere—the earth—of radius r), is equal to the solar heating of the surface minus the radiative energy emitted by the earth's surface plus the radiation emitted downward by the atmosphere. Eq. (3.4.5) represents the energy budget of the entire atmosphere, represented, contrary to the fact, as having a single temperature, a single density, and a finite thickness H. Any positive change in the thermal energy of the atmosphere is equal to energy radiated by the earth's surface (σT_s^4) and absorbed by the atmosphere with its characteristic emissivity ε_a (which we take to equal to its absorptivity, its fractional ability to absorb infrared radiation), minus the energy radiatively emitted by the atmosphere, both upward and downward (thus the factor of 2 in the second term on the left).

Simplifying, and assuming a steady solution with time derivatives equal to zero, we get from Eq. (3.4.5) that $T_a^4 = \frac{1}{2}T_s^4$. Substituting this back into Eq. (3.4.4) yields:

$$\sigma T_s^4 - \varepsilon_a \frac{1}{2} T_s^4 = \left(1 - \frac{\varepsilon_a}{2}\right)\sigma T_s^4 = \frac{S}{4}(1-\alpha). \tag{3.4.6}$$

[c]A slightly more complicated version of this model that includes the possibility of some atmospheric absorption of solar radiation can be found in Marshall and Plumb.[36]

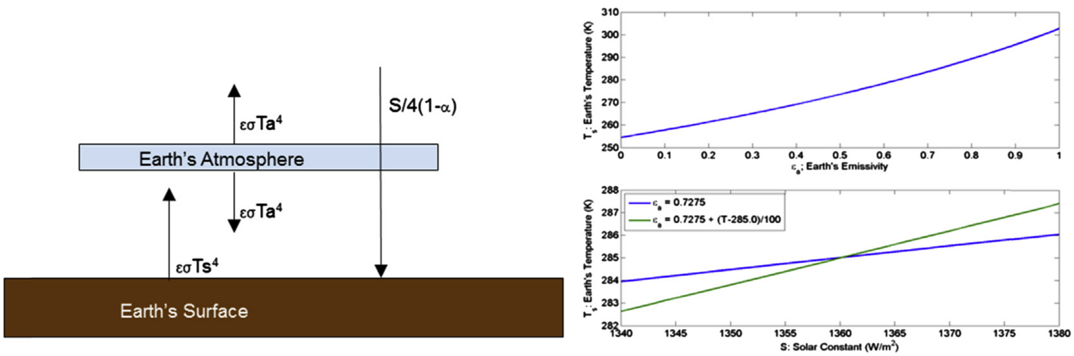

FIGURE 3.4.4 The diagram on the left shows the conceptual model of the earth's climate expressed in Eqs. (3.4.4) and (3.4.5). The top figure on the right shows the resulting steady-state temperature T_s as a function of the value of ε_a, the emissivity of the earth's atmosphere. The bottom figure on the right shows the T_s for a range of values of S, and again for the case where ε_a is a function of T_s. This shows the effect of a climate feedback: if the emissivity increases as a function of temperature, then the planet will respond more to an increase in solar forcing than it does for constant emissivity. This concept will be explored further in Section 3.4.4.4.

Solving for T_s gives

$$T_s = \sqrt[4]{\frac{S(1-\alpha)}{4\sigma\left(1-\frac{\varepsilon_a}{2}\right)}}. \tag{3.4.7}$$

Note that Eq. (3.4.6) implies that the emissivity of the earth-atmosphere system can be written as $\varepsilon = 1 - \frac{\varepsilon_a}{2}$: as the emissivity of the atmosphere increases, the emissivity of the earth as a whole decreases. Increasing atmospheric opacity to infrared radiation makes the earth's surface less able to shed radiation to space. So we should expect that as the atmosphere's emissivity increases, the temperature will need to increase, so that the system can shed the heat received from the sun and remain in equilibrium. The graph on the right of Fig. 3.4.4 shows the dependence of T_s on ε_a: as the atmosphere's ability to absorb and emit infrared radiation increases, the temperature of the surface does indeed increase. Thus gases such as carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), and ozone (O_3) that absorb and emit infrared radiation are referred to as greenhouse gases, because, like the glass of a greenhouse, they reduce the net transfer of energy from the earth to space, warming the earth below.[d]

The upper limit of about 303K applies only to this model with its single-layer atmosphere with uniform temperature: a more realistic model with a continuous atmosphere can achieve much higher surface temperatures, as in fact occurs on Venus.[17] To understand why higher emissivity leads to higher temperature from a dynamical point of view, we can inspect the differential Eqs. (3.4.4) and (3.4.5) directly. If the emissivity increases, Eq. (3.4.4) tells us

[d]Actual greenhouse glass keeps greenhouses warm mostly by reducing *convection*, air motions that transfer heat away from the earth's surface into the atmosphere above,[37] whereas *greenhouse gases* warm the planet through radiative interactions, as we will discuss later.

that we should expect the resulting increase in downward radiation from the atmosphere to warm the surface. Eq. (3.4.5) tells us that we should expect the increase in radiative heat loss due to the atmosphere's emission of infrared radiation to be compensated by an increase in absorption of infrared radiation from the surface. The eventual increase in surface temperature leads to net warming of the atmosphere.

3.4.4.2 Radiative Transfer

The real atmosphere does not have a single emissivity. Rather, the ability of the atmosphere to absorb and emit radiation depends strongly on the wavelength of the radiation. This wavelength dependence in turn arises from the molecular properties of the atmosphere's various constituents, and from bulk properties of the atmosphere: temperature, density, and pressure. The phenomena associated with emission, absorption, and transmission of radiation in the atmosphere are collectively referred to as radiative transfer. Absorption and emission of energy by molecules is quantized. Radiative energy arrives at and departs from molecules in the form of photons with particular energies, and only those photons whose energy corresponds to a permitted energy transition of the molecule can be absorbed or emitted. For energies corresponding to infrared radiation, these transitions correspond to changes in the rotational and vibrational energy states of molecules in the atmosphere. Thus constituents like argon, which is present as individual atoms, and oxygen and nitrogen, which are present as homonuclear diatomic molecules, do not interact with infrared radiation: their symmetry means they do not present an electric dipole to the oscillating electromagnetic field associated with infrared radiation. Carbon dioxide obtains a dipole moment when it is warm enough to vibrate, as it is at terrestrial temperatures, so it has a rotational-vibrational set of absorption lines centered on the frequency of its bending mode at 667 cm^{-1}. Methane has similar rotational vibrational-rotational absorption bands. Water vapor has these, but, in addition, it has a strong pure rotational mode that extends over a broad region of the infrared spectrum below 600 cm^{-1}, because it possesses a dipole moment in its ground state, due to its bent geometry.

Fig. 3.4.5 shows the spectrum of radiation emitted by the earth to space as a function of location and season. Regions of the spectrum where atmospheric gases have high emissivity appear as low points in the spectrum (e.g., around 650 cm^{-1}), because radiation emitted by the warm surface is absorbed strongly by constituents of the atmosphere (in this case carbon dioxide, CO_2), whereas the radiation observed from space is emitted from the highest parts of the atmosphere, which are cold, and do not emit much radiation there. Regions of the spectrum where atmospheric gases have low emissivity (e.g., between 800 and 950 cm^{-1}) appear to touch up against a smooth curve, which would be close to the emission by a blackbody at a temperature close to that of the earth's surface, as predicted by Planck's law. The dips in radiation emitted to space in Fig. 3.4.5 are referred to as absorption lines and collectively as bands. For example, the region between 570 and 750 cm^{-1} is referred to as the CO_2 absorption band. Increasing the concentration of carbon dioxide increases the strength of its band, increasing the effective emissivity of the atmosphere as a whole.

Carbon dioxide is relatively abundant in the atmosphere at 400 ppmv, so the carbon dioxide band is strong, and increasing the concentration of the gas in the atmosphere mostly acts to increase the emissivity at the edges of the band, since the atmosphere is already nearly

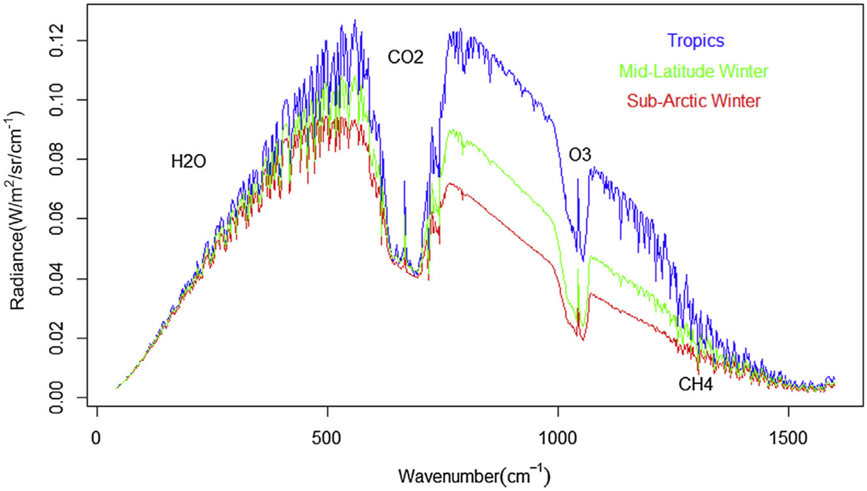

FIGURE 3.4.5 Overview of the earth's outgoing infrared radiation as a function of wave number (the inverse of wavelength) and latitude.[43] Radiances for this figure were calculated using Modtran and a web interface developed by David Archer available here: http://climatemodels.uchicago.edu/modtran/.

opaque to radiation at wavelengths in the center of the band. As a result, the radiative forcing of carbon dioxide (the extra heating at the surface due the presence of the gas, equivalent to $\varepsilon_a \sigma T^4$ in Eq. 3.4.5) goes as the natural logarithm of the concentration of the gas in the atmosphere. This is why one often hears about climate change referenced to a doubling of CO_2: for any doubling of the concentration of the gas, whether from 275 to 550 ppmv or from 300 to 600 ppmv, the natural log of the concentration increases by a fixed amount and the radiative forcing increases by about 3.7 W/m^2. On the other hand, a greenhouse gas like sulfur hexafluoride (SF_6), whose concentration is very low (8.3 parts per trillion), has very narrow, weak absorption lines. The impact of each additional molecule is very large (23,000 times larger than CO_2) because the lines can get deeper and broader with each additional molecule.[1] Its radiative forcing increases linearly with concentration. Methane, at 1.8 ppmv, is an intermediate case: its radiative forcing increases as the square root of its concentration.[18]

We humans are emitting carbon dioxide in large amounts (about 36×10^{12} kg, or 36 GT CO_2 per year from fossil fuel combustion and an additional 6 GT CO_2 that results from the burning or decomposition of trees when forecasts are cleared for agriculture). The *rate* of emissions has been increasing over the past 150 years, so that the concentration of CO_2 has increased exponentially, by about 0.5% per year since the mid-20th century. This exponential increase in concentration, combined with a logarithmic dependence of radiative forcing on concentration has resulted in a near-linear increase in radiative forcing due to CO_2. Expressed mathematically, if R is the warming due to CO_2 in W/m^2, a is a constant of proportionality, C is the present concentration of CO_2, and C_0 is the concentration at t_0, b is the exponential growth rate of the gas concentration, and t is the time in years, then:

$$R = a \ln (C/C_0) = a \ln \left(C_0 e^{b(t-t_0)}/C_0\right) = a(bt - bt_0) \tag{3.4.8}$$

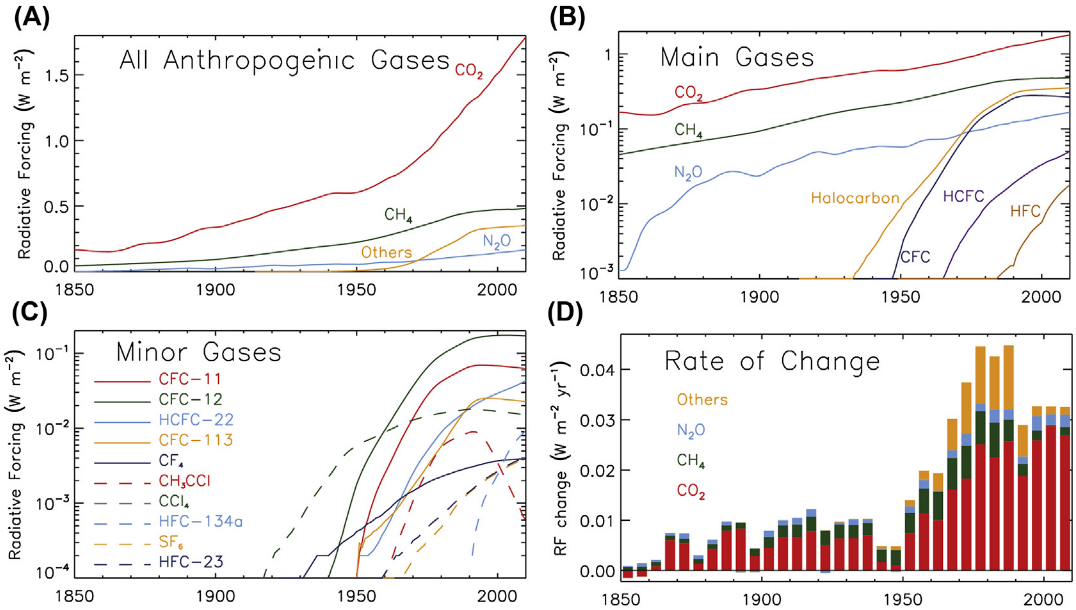

FIGURE 3.4.6 (A) Radiative forcing (RF) from the major well-mixed greenhouse gases (WMGHGs) and the sum of the minor gases such as chlorofluorocarbons and SF_6,1850 to 2011. (B) as (A) but with a logarithmic scale, (C) RF from the minor WMGHGs from 1850 to 2011 (logarithmic scale). (D) Rate of change in forcing from the major WMGHGs and groups of halocarbons from 1850 to 2011 (Fig. 8.6 in the IPCC Fifth Assessment Report[1]).

Fig. 3.4.6 shows the radiative forcing and its annual rate of change for the major greenhouse gases emitted by human activities: CO_2, CH_4, N_2O, and the sum of all other anthropogenic greenhouse gases. Since about 1960 the increase in radiative forcing due to CO_2 has been nearly linear, whereas the rate of increase for CH_4 has slowed and the rate for N_2O has accelerated, but from a low level. The large rates of change in the "other" category in the 1970s and the 1980s arose from the chlorofluorocarbons, gases that were found to damage the stratospheric ozone layer, and were banned by the 1989 Montreal Protocol: their rate of increase decreased sharply thereafter.

3.4.4.3 Anthropogenic Climate Change in Space and Time

Comparing the left panel of Fig. 3.4.6 with the left panel of Fig. 3.4.1, we note that as radiative forcing has increased in a nearly linear fashion over the past 50 years, temperature has also increased nearly linearly with time. We will now consider some of the questions this relationship suggests. Should we expect the temperature to continue to rise linearly as the forcing increases? What do we observe to be the pattern of temperature and other climate changes associated with this global mean temperature rise? And how should we expect these patterns to change over time?

In the following discussion, we will refer often to the "climate system," by which we mean the atmosphere, the oceans, and the parts of the land surface (including the living things that grow on or in it) that can interact with the atmosphere and ocean over time scales of days to thousands of years: soil, forests, fields, lakes, rivers, glaciers, and ice sheets. We often talk about the climate system responding to "external forcing," by which we mean either things not included in the earlier description (the sun, the orbital change caused by the earth's gravitational interaction with other planets, volcanoes, continental drift) or things done by human beings. We humans are not, of course, truly external to the climate system, but we will respect this useful fiction for the sake of clarity.

Understanding the response of the climate system to external forcing requires that we write down the equations that govern the system and then solve them. These equations involve derivatives of climate variables in space and time.[e] Since they do not have simple analytic solutions, we typically solve them in approximate form as difference equations using computers. Having expressed and assembled our understanding of the climate system in this way, we can pose questions to these climate or Earth system models: what do we expect to happen to temperatures in Siberia when a powerful volcanic eruption occurs in the Philippines? What do we expect to happen to rainfall in Iowa when carbon dioxide doubles? We can compare the answers that climate models give with what has been observed to occur in the historical record and gain a sense of how predictable the impacts of various forms of external forcing may be. For example, the volcanic eruption of Mount Pinatubo in 1991 resulted in a global averaged negative radiative forcing of about $-4\,\mathrm{W/m^2}$, about the same magnitude as a doubling of carbon dioxide, but of opposite sign and lasting for only about 18 months.[19] Volcanoes act to cool the earth mostly by injecting sulfur dioxide into the stratosphere, where it is converted to a haze of sulfuric acid droplets. This haze was thick enough to be visible to the naked eye as whitening around the sun on clear days in the year following the eruption. This forcing resulted in global average cooling of about $-0.4\,°\mathrm{C}$, peaking about 1 year after the eruption. This global average cooling was also produced by climate models, and the range of cooling observed in the models can be compared to the models' predicted warming under increasing carbon dioxide.[18] Other effects of such large climate forcing events, such as changes in wind and precipitation patterns, can also be compared in models and observations.[20]

3.4.4.4 Feedbacks and Climate Sensitivity

One of the primary questions that climate science has sought to address over the past few decades is: how much warmer will the earth get as a result of an increase in carbon dioxide? To answer this question, climate scientists first needed to state it carefully. This led to a few important definitions. The earth's *equilibrium climate sensitivity* (*ECS*) is the global average change in temperature of the air 2 m above the surface between two long periods (e.g., a few centuries), of which during one period the atmosphere's CO_2 concentration was close to its preindustrial value of 275 ppmv, and during the other it was twice that value, 550 ppmv. The *transient climate response* (*TCR*) is the change in temperature of the earth during a period when the CO_2 concentration is rising steadily from its preindustrial value of 275 ppmv, between the preindustrial average temperature and the temperature averaged over a few years centered on the time when the CO_2 concentration reaches two times its preindustrial values (550 ppmv). The TCR and the ECS are different because the earth's heat capacity, C_s in Eq. (3.4.4), is not zero, which means that the earth's temperature takes some time to respond to a radiative forcing (e.g., from a change in CO_2 concentration).

[e]For a thorough introduction to these equations, see Marshall and Plumb.[36]

The right-hand panels of Fig. 3.4.4 help to illustrate the concept of ECS. The upper panel shows the response of the surface temperature in the simple model of Eqs. (3.4.4) and (3.4.5) to changing emissivity (the simple model equivalent of changing the CO_2 in the real atmosphere). As the emissivity increases, the equilibrium temperature increases, by a nearly fixed amount for each increase in emissivity. The lower panel shows two different curves. In this case, the sensitivity being measured is to a change in the solar constant (S, the radiation received from the sun by a flat plane facing the sun at the position of the earth's orbit) rather than a change in emissivity. The blue and green curves show the sensitivity of two different model planets. The blue curve shows how the temperature changes as S increases for the model in Eqs. (3.4.4) and (3.4.5). The green curve shows how the temperature changes for a modified model in which the emissivity itself increases when the temperature increases:

$$\varepsilon_a = \varepsilon_0 + (T_s - 285\mathrm{K})/(100\mathrm{K}) \tag{3.4.9}$$

where ε_0 is the emissivity such that $T_s = 285\mathrm{K}$ in the upper panel ($\sim$0.7275). In this model, a little extra heat from the sun (higher S), results in a larger increase in temperature than in the original model, because the warmer temperature also goes along with a higher emissivity, which means a little more radiation coming down from the atmosphere in Eq. (3.4.5), and thus an extra increase in surface temperature. This kind of mechanism, which increases the sensitivity of the model to the same change in energy from the sun, is referred to as *positive feedback*. This extra dependency of ε_a on T_s makes the model a bit harder to solve: one has to replace the derivatives on the left-hand sides of Eqs. (3.4.4) and (3.4.5) with time differences (setting $dT_s = T_{st+\Delta t} - T_{st}$, where Δt is a finite amount of time), solve for $T_{st+\Delta t}$ in terms of T_{st}, and then step through time until T_s stops changing.

Note that the existence of the aforementioned positive feedback does not *necessarily* imply an unstable climate, where any initial push causes a response that magnifies the initial push, so that the temperature rises or falls infinitely. Such a system is possible: simply replace the term (100K) in Eq. (3.4.8) with (50K) and the model climate will quickly shift to a state where either $\varepsilon_a = 0$ or $\varepsilon_a = 1$, assuming ε_a is required to lie in that range. And indeed, there have been times in the earth's history, most recently about 800 million years ago, when, because the sun emitted less radiation, a run-away to an "icehouse" climate occurred. As a result, the entire planet became covered with ice, and remained so until enough CO_2 was emitted by volcanoes so that the run-away operated in reverse, warming the planet swiftly and exposing enough ocean water to allow the CO_2 to be drawn down into the ocean and ocean floor deposits of calcium carbonate.[21] As the sun continues to brighten over the coming hundreds of millions of years, it is anticipated that we will again enter an unstable regime, when all the earth's oceans will evaporate into the atmosphere, leaving the earth with a Venus-like climate (unless of course intelligent life remains on the planet and is able to intervene effectively in this process with its own stabilizing feedback).[22]

However, in the current climate, we are in a regime where the existence of positive feedbacks merely weakens the strong negative feedback of the Stefan-Boltzmann law, so that warming temperatures result in increasing outgoing radiation, but the outgoing radiation increases with temperature *at a lower rate* than it would in the absence of positive feedbacks. Note that these feedbacks need not be active only for infrared radiation. The *ice-albedo feedback* describes the climate impact of the fact that warmer temperatures tend to result in smaller areas of ice and snow on Earth, so that as the temperature increases and the snow cover decreases, less solar radiation is reflected away from the earth and the earth warms more in response to a change in greenhouse gases or in solar radiation than it would otherwise. Cloudiness changes in response to warming temperatures can also cause changes in both long-wave emissions (since clouds act to increase infrared emissivity just as greenhouse gases do) and in short-wave emissions (since increasing cloud cover increases reflection of solar radiation). Changes in cloud properties with global temperature are among the most uncertain aspects of climate dynamics.[1]

The various climate feedback mechanisms can be compared quantitatively if we introduce a simplified form of the climate energy balance expression, in which we add a "forcing" term F representing some external change in the earth's radiative balance, for example, a change in the solar constant or a change in greenhouse gas concentration, that results in a measurable change in the balance of radiation at the top of the earth's atmosphere, expressed in W/m^2, and a "response" term, R, that represents the change of radiative balance at the top of the atmosphere associated with the warming or cooling of the earth in response to the forcing F, due to feedbacks other than blackbody. In equilibrium, the response will equal the forcing. In the "normal" climate, $\alpha = 0.31$, $\varepsilon_a = 0.61$, $S = 1360\ W/m^2$, and $T_0 = 288K$. We can write:

$$\frac{S}{4}(1-\alpha) + F = \varepsilon_a \sigma (T_0 + T)^4 + R,$$

where $\frac{S}{4}(1-\alpha) = \varepsilon_a \sigma T_0^4$, and since $\Delta T << T$, we can write:

$$F = 4\varepsilon_a \sigma T_0^3 \Delta T + R$$

Next we expand R into various responses to changing global temperatures:

$$F = 4\varepsilon_a \sigma T_0^3 \Delta T + R_{wv} + R_{\text{cloud}} + R_{\text{snow}} + R_{\text{lapserate}}$$

and express these responses as linear functions of the global mean temperature:

$$F = 4\varepsilon_a \sigma T_0^3 \Delta T - \lambda_{wv}\Delta T - \lambda_{\text{cloud}}\Delta T - \lambda_{\text{snow}}\Delta T - \lambda_{\text{lapserate}}\Delta T$$

The feedbacks are subtracted from the primary Planck feedback to preserve the convention that a *negative* feedback will result in more outgoing radiation (or less absorbed incoming radiation), and thus act to make the equilibrium temperature *lower*.

Rearranging and solving for ΔT, we get:

$$\Delta T = \frac{F}{4\varepsilon_a \sigma T_0^3 + \lambda_{wv} + \lambda_{\text{cloud}} + \lambda_{\text{snow}} + \lambda_{\text{lapserate}}} \tag{3.4.10}$$

For a doubling of CO_2, $F = 4\ W/m^2$, and since $4\varepsilon_a \sigma T_0^3 = 3.3\ W/m^2\ K$, Eq. (3.4.10) indicates that a doubling of CO_2 would lead, in the absence of any other feedbacks, to a warming of $\Delta T = 4.0\ W/m^2 / 3.3\ \ W/m^2\ K = 1.2\ K$. However, the other feedback mechanisms, when evaluated in climate models and in observational studies, are found on balance to be negative, meaning that warming temperature is associated with either reduced outgoing infrared radiation (e.g., because of increased atmospheric water vapor) or increased absorbed solar radiation (e.g., because of reduced snow cover and thus lower albedo). According to recent estimates from climate models, typical values of the climate feedback parameters are $\lambda_{wv} = 1.5\ \mathrm{W/m^2}$, $\lambda_{\mathrm{cloud}} = 0.5\ \mathrm{W/m^2}$, $\lambda_{\mathrm{snow}} = 0.3\ \mathrm{W/m^2}$, and $\lambda_{\mathrm{lapserate}} = -0.3\ \mathrm{W/m^2}$. Subtracting these from the Planck feedback gives a denominator of $1.3\ \mathrm{W/m^2}$ in Eq. (3.4.9) resulting in an ECS of about 3K, typical of many climate model estimates of the ECS.[23,24]

The ECS is one important aspect of the earth's climate system; it tells us how *much* the average temperature of the air just above the earth's surface will warm in response to a given radiative forcing, if we wait for the system to adjust itself completely. For practical purposes, however, it is important to understand how *fast* this adjustment will be. If all the extra energy from doubling CO_2 were mixed immediately throughout the oceans, the adjustment to equilibrium would take thousands of years and we might conclude that we could afford to do nothing to reduce emissions for hundreds of years, waiting for a perfect technological solution that would allow us to suck CO_2 out of the atmosphere and store it somewhere away from the atmosphere. However, in fact, the observed behavior of the climate system, both in response to greenhouse gas accumulation and in response to volcanic eruptions, shows that the leakage of heat into the deep oceans is quite slow. As a result, much of the approach to equilibrium is accomplished in the first tens of years, and involves only the land surface and the upper few hundred meters of the ocean. One way to quantify simultaneously the rate and magnitude of the climate response to greenhouse warming is the TCR. The TCR can be estimated from the rate of change of the global mean temperature observed so far, in combination with an estimate of the change in radiative forcing due to both greenhouse gases and aerosol pollution, compared to the preindustrial period. It can also be modeled using coupled climate models. Current (2016) global mean surface air temperature is about 1K warmer than in preindustrial times, and is warming at about 0.2K/decade, and CO_2 is currently increasing at 30 ppmv/decade from a value of 404 ppmv. A very simple approach to estimating the TCR would be to extrapolate these rates of increase to find the expected time of CO_2 doubling and the temperature at that time. At the present rate of increase, we expect that CO_2 would rise from its preindustrial value of 275 to −550 ppmv by about the year 2065, and if we extrapolate the present rate of global temperature increase, we would expect to reach a warming of 2K at that time. So the present trends are consistent with a TCR of 2K, assuming that aerosol cooling has approximately canceled warming from the greenhouse gases other than CO_2.[1] More detailed recent estimates of the TCR that account in detail for all known changes in radiative forcings do in fact result in a TCR of 2K ± 0.8K.[25]

3.4.4.5 Consequences of Global Warming

When the earth's average temperature changes, other climate system changes should be expected. For example, when the earth was about 4K colder at the coldest point of the last ice age, some 20,000 years ago, ice sheets expanded far equatorward, the sea level fell by about 125 m, precipitation was reduced over many midlatitude continent regions, and global dustiness increased dramatically.[26] What changes in climate variables other than temperature should we expect in a warmer world?

We will start from the most obvious consequences of a globally warmer world and move to consequences that are more complicated and more uncertain. In the first place, it has turned out to be the case that as our world has warmed globally by a degree or so, almost all locations have experienced warming. This need not have been the case: because of changes in ocean circulation that have accompanied global warming, a few locations such as the North Atlantic Ocean in the area around Greenland and the Southern Ocean off Antarctica have experienced some cooling. At times it has been speculated that these areas might have been larger, and might have included large section of Northern Europe, but the observational record has shown that these areas of cooling have been very limited.

However, the distribution of warming has not been, and is not expected to be, uniform. Warming is larger over land than over oceans for two reasons. As greenhouse gases increase and the infrared radiation received by the earth increases, the land and the ocean respond differently. First, the heat capacity of the ocean is larger than that of the land, both because water has a larger heat capacity than soil or rock, and more importantly, water can move and mix heat down from the surface, whereas heat can only diffuse slowly through rock and soil. So the extra heat received from the atmosphere is shared over a large depth of the ocean, but only a small volume of rock and soil on land. Second, evaporation from the ocean allows more heat to be carried away from the ocean (in the form of the latent heat of evaporation) than by turbulent mixing of sensible heat (proportional to temperature) alone. While this also occurs over moist soil, most land surfaces are less than fully saturated with water, so less evaporation can occur, and the land surface must warm to a higher temperature to balance the increase in downwelling infrared radiation.[27]

Warming is also larger over the high latitudes than in the tropics or over the South Pole. This is largely due to the ice-albedo and snow-albedo feedback, where warming results in less ice and snow, which in turn results in a darker surface of the earth that can absorb more radiation from the sun. The feedback does not function very strongly in Antarctica, which is so cold that no melting occurs (except on the Antarctic Peninsula), but has been very effective over the Arctic Ocean, where melting sea ice replaces a very reflective sea ice surface with very dark ocean water.[28] These effects are visible for both observed warming and warming projected by climate models in Fig. 3.4.7: warming is greater over the land than over the sea and greater at high latitudes than in the tropics.

Warming has occurred and is expected to occur in all seasons of the year and all times of the day, but in most locations, warming is stronger at night than in the day and slightly stronger in the winter than in the summer. This is because daytime conditions and summer conditions experience more *deep convection*, by which air rises vertically in thunderstorms. This process moves heat from the surface to the upper atmosphere more efficiently than radiation or the slanted motions that occur in wintertime storms.[1]

3.4.4.6 Sea Level Rise

Beyond warmer surface temperature, a globally warmer world implies that the upper layers of the ocean are warmer, and warmer water takes up more space than colder water. The density of seawater decreases by approximately 0.34 kg/m^3K. Since a cubic meter of sea water has a mass of about 1017 kg, for every 1K of warming of a 100-m-thick ocean layer, we expect about

$$\left(\frac{1\ \mathrm{m}^3}{1016.66\ \mathrm{kg}} - \frac{1\ \mathrm{m}^3}{1017.0\ \mathrm{kg}}\right) \times 1017\ \mathrm{kg} \times 100\ \mathrm{m} = 0.036\ \mathrm{m}$$

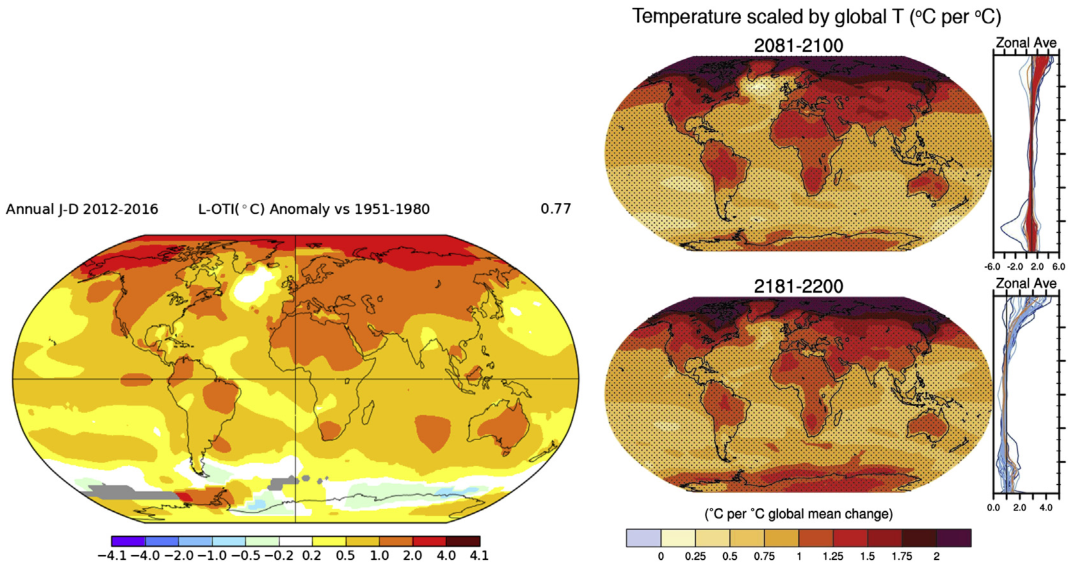

FIGURE 3.4.7 Warming observed to date (left)[38] and the distribution of warming expected for each degree of global average warming.[1] That is, the maps on the right show the amount of warming at each location in the world, divided by the global mean warming, averaged over the predictions of a large number of climate models, for two different times in the future: the end of the 21st century (top) and the end of the 22nd century (bottom). This emphasizes the strong agreement among the models that high latitude and land areas warm more than low latitude and ocean areas. The agreement with the observed pattern is also good, except for the observed cooling near the coast of Antarctica.

of sea level rise. The planet has warmed about 1K since the middle of the 20th century, and during this time, the sea level has risen by about 15 cm. The temperature of the ocean has been measured carefully and thoroughly down to about 1000 m since the turn of the 21st century; it turns out that the warming of the upper levels of the ocean accounts for about one-sixth of the observed sea level rise since 2003, as shown in Fig. 3.4.8. Other factors must be contributing to this rise. A warmer world is a world in which the snow line on mountains is higher up and glaciers are typically arranged with half their mass above the snow line and half their mass below. As the snow line rises, the total mass of the glacier will decrease (since the glacier cannot retreat above the top of the mountain!). Mountain glaciers are observed to be shrinking worldwide; this process has contributed a little more than 1 cm of sea level since 1960. Finally, warming air and sea temperatures are acting to increase melting of the continental ice sheets in Greenland and Antarctica, and to accelerate the flow of ice from these ice sheets into the ocean. This process has been especially strong in Greenland, whereas in Antarctica it has been at least partially counteracted by an increase in snowfall.[36] Although the contribution from this process has been a bit less than 1 cm of sea level rise since 1960, it is being closely monitored by satellite observations and direct measurement because the potential for sea level rise is so large. Melting the Greenland Ice Sheet would result in about 7 m of sea level rise, and we know that this has occurred in the past, as recently as the last warm period before the last ice age, 125,000 years ago. Such a continental-scale melting could not occur very rapidly, but

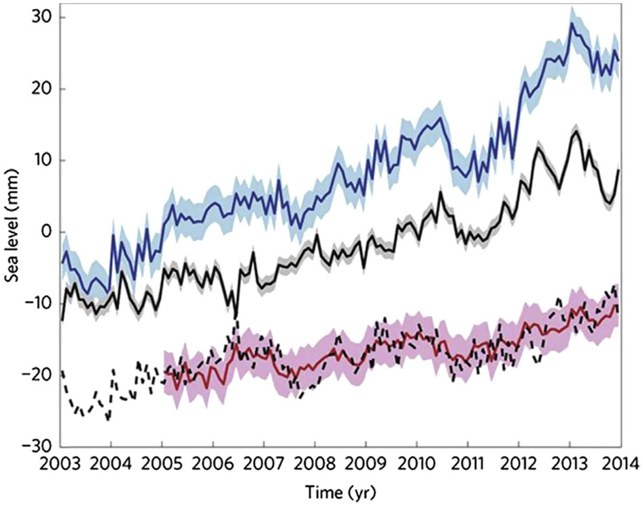

FIGURE 3.4.8 Contributions to sea level rise from reduced density of warming water [red (gray in print versions) curve, with pink (lightest gray in print versions) uncertainty bounds], from loss of ice on land (black curve with gray uncertainty bounds) and total sea level rise [blue (dark gray in print versions) curve, with light blue uncertainty bounds]. The *black (dark gray in print versions) dashed line* is a consistency check: it shows the difference of the *blue* (dark gray in print versions) and *black curves* (observed by different satellite methods) and agrees well with the *pink curve* (lightest gray in print versions) (calculated from observed temperature changes).[44]

up to 1 m of sea level rise due to melting of continental glaciers alone is thought to be possible by the year 2100, an amount sufficient to inundate large regions of coastal lands with high populations and large agricultural productivity.[1]

3.4.4.7 Ecological Consequences of Warming

Warming temperatures are causing changes in the range of various animal and plant species and stressing some populations. For example, warming of wintertime temperatures in western North America has led to a wider range of the Mountain Pine Beetle, which has led to blight and increased tree mortality over the Rocky Mountains of British Columbia, Canada.[29] Similarly, higher wintertime temperatures have led to the expansion of insect predators and forest damage to New Jersey pinelands and spruce forests in Alaska.[30] On the other hand, rising CO_2 concentrations themselves make it easier for plants to generate biomass via photosynthesis. Plants obtain CO_2 from the atmosphere by opening pores in their leaves called *stomates*, but when the stomates are open, they allow water to be lost from the leaves. With higher CO_2 concentration, plants can gain the needed carbon from the atmosphere with less loss of water vapor, allowing for better tolerance of high temperatures and water stress. In addition, warmer temperatures themselves extend plants' growing season in cool climates. As a result, the overall forest growth appears to be increasing over time in regions where forests are not being cut back.[31]

Warming is also associated with changes in atmospheric chemistry that can increase tropospheric ozone formation, a process discussed in more detail in Chapter 3.2.

3.4.4.8 Changes in Precipitation

Besides higher temperatures and their immediate consequences, a warming climate is observed and expected to differ from the preindustrial climate in its patterns of storminess and precipitation as well. These impacts are more uncertain, for two main reasons. First, precipitation is more variable from year to year than temperature in many locations, so it is harder to be confident of the detection of a long-term trend: the signal-to-noise ratio is lower. Second, precipitation is more difficult to predict using the weather models that form the core of climate models, so we are less confident of the mechanisms for changes in precipitation than for temperature and there is a wider disagreement among the various climate models developed around the world.

With this caveat in mind, there are some observed and predicted changes in precipitation of which we are reasonably confident. First, the distribution of precipitation is shifting from more frequent light events to less frequent and heavier events. Second, models and observations tend to show increases in precipitation in cold high latitude regions, and a decrease in precipitation in regions on the edge of the Tropics: the Mediterranean region in Europe and North Africa, Mexico, the southern United States, Southern Africa, and Australia. Third, changes are expected in the distribution and intensity of hurricanes and typhoons. These storms draw their energy from the contrast in temperature between the warm ocean surface and the cold upper atmosphere, and climate models generally predict that some regions of the planet, especially in the Pacific Ocean, will be able to support more intensive hurricanes in the future. The models generally predict a larger number of more intense storms, and a smaller number of less intense storms.[32] Fig. 3.4.9 shows an average over the models consulted by the Intergovernmental Panel on Climate Change (IPCC) of changes in precipitation, scaled by the change in temperature in each model. A general pattern of increasing precipitation near the equator, reduced precipitation in the subtropics (and extending notably into the Mediterranean region) and increased precipitation in the high northern and southern latitudes is evident.

3.4.5 TECHNOLOGY TO REDUCE GREENHOUSE GAS EMISSIONS

In light of the impacts discussed in the previous sections and reviewed in detail by the IPCC,[33] economists have estimated the damage done to the global human economy for each ton of CO_2 emitted today. A recent review calculated a range of values (due to uncertainty in economic damage of given effects, together with uncertainty in the amount of climate change for given level of greenhouse gas emissions) of \$7 to \$77 per ton of CO_2, with a most likely value of \$31 per ton.[34] The figure is often referred as the *social cost of carbon.* A consequence of this calculation is that an effort to reduce emission of CO_2 or an equivalent amount (in terms of climate impact averaged over some period of time) of another greenhouse gas, would be likely to have a net benefit to the human economy if it could be accomplished for less than \$32 per ton of reduced emissions.

Many such measures can be imagined. For example, it may be possible to add insulation to poorly insulated houses in such a way that the owner saves more money from reduced heating and cooling costs than the cost of the insulation and the labor to put it in place.

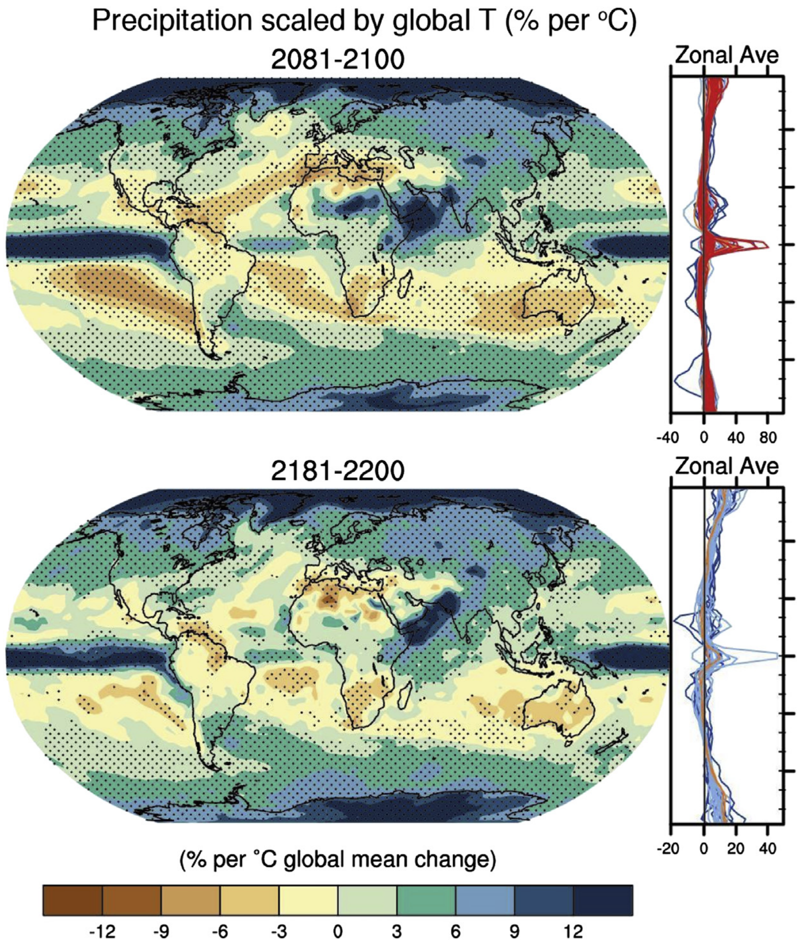

FIGURE 3.4.9 Patterns of observed precipitation change since 1970 and predicted patterns of precipitation changes per degree of global warming.[1]

In that case, CO_2 emissions are reduced for zero cost. Alternatively, one could design a device to remove CO_2 from the air and condense it into liquid form, using electricity generated by a hydroelectric dam. A demonstration facility to do just this was put in operation in Squamish, BC, in 2016.[35] The cost of such a process would depend on the success of a long process of learning by doing, but at present is likely much higher than $31 per ton CO_2. Other means of reducing CO_2 emissions are further along toward industrial-scale development. In locations that are windy or sunny, electricity generated from wind by wind turbines or from sunlight by photovoltaic panels is as cheap as any other source of electricity. A tax imposed on CO_2 emissions, set at something close to the estimated social cost of carbon, is often discussed as a simple measure that would tend to drive the economy toward lower carbon modes of production. Depending on the size of the tax, and the

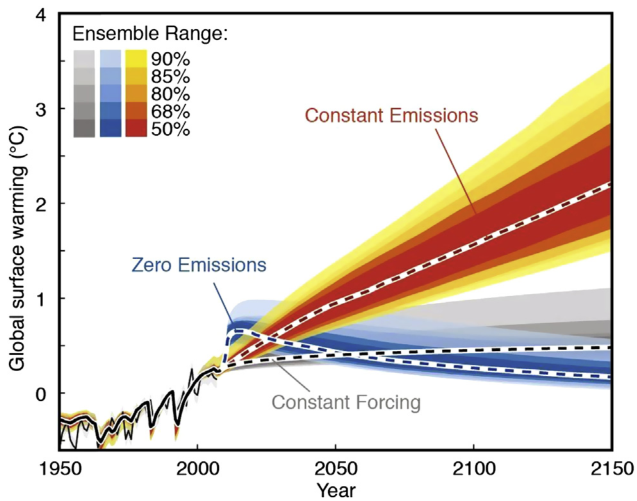

FIGURE 3.4.10 Probable results for global average temperature from various future greenhouse gas emission pathways.[1] The colors indicate the range of climate model uncertainty for each of three different emissions pathways. The red (gray in print versions) and yellow (light gray in print versions) colors show the range of climate outcomes for emissions of greenhouse gases frozen at 2011 levels, with the *dashed line* being the most probable outcome. The blue (dark gray in print versions) colors show the range of climate outcomes if emissions were suddenly halted. The gray colors show the outcome if emissions were halted in such a way as to produce constant climate forcing after 2011. The blue (dark gray in print versions) trajectories jump suddenly because emissions of greenhouse gases are accompanied by emissions of short-lived aerosols that tend to cool the planet. When emissions are halted, the aerosols decline sharply to preindustrial levels over a few years, resulting in a sudden jump in temperature, which decays as methane is converted to CO_2 in the atmosphere and CO_2 is drawn down into the ocean. Note that greenhouse gas emissions have been increasing steadily for decades, so all these pathways imply some social effort to reduce emissions, or at least to halt the growth of emissions. "Business as usual" with continued increasing emissions, would result in even warmer conditions than shown for the red (gray in print versions)-shaded emission scenario.

fraction of the world's economic activity on which it could be imposed, and on the response of human scientific, engineering, and economic activity, such a tax might be expected to slow the growth of emissions, to freeze emissions, or to actually reduce or even reverse emissions, with the accompanying range of results for climate shown in Fig. 3.4.10 below.

The human economy is far more complex and far more difficult to predict than the climate system. How rapidly we will be able and willing to reduce greenhouse gas emissions will determine how dramatic the consequences of their rising concentration on natural environments and human communities will become. Climate science has achieved a remarkable success in correctly predicting, well in advance, that fossil fuel combustion would lead to large changes in the earth's climate. Political and diplomatic action in response to this knowledge has led to the rapid growth in new non-fossil fuel energy industries. Our descendants will judge whether our response meets the demands of this knowledge and opportunity.

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