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Part I

Chapter 1

Climate Change

The threat of human-induced climate change, popularly known as global warming, presents a difficult challenge to society over the coming decades. The production of so-called “greenhouse gases” (GHGs) as a result of human activity, mainly due to the burning of fossil fuels such as coal, oil, and natural gas, is expected to lead to a generalized warming of the Earth’s surface, rising sea levels, and changes in precipitation patterns. The potential impacts of these changes are many and varied – more frequent and intense heat waves, changes in the frequency of droughts and floods, increased coastal flooding, and more damaging storm surges – all with attendant consequences for human health, agriculture, economic activity, biodiversity, and ecosystem functioning.

Because the impacts of climate change are expected to be global and potentially severe, and because energy production from fossil fuels is a fundamental component of the world economy, the stakes in the issue are high. At the same time, a number of aspects of climate change complicate the problem. First, while much is known about the factors governing climate, considerable uncertainty remains in projections of how much climate will change, how severe the impacts will be, and how costly it would be to reduce GHG emissions. Second, because the impacts of today’s GHG emissions will be felt for decades into the future, it is not possible to wait and see how severe impacts will turn out to be before taking preventive action. Therefore, if emissions are reduced now, the costs will be borne in the near term while the benefits, which will depend on uncertain projections of future impacts, will be realized largely in the long term. Third, sources of GHG emissions are widely dispersed among nations; no single country could significantly reduce future global climate change just by reducing its own emissions. Even reductions in the more developed nations as a group, which have been responsible for the bulk of historical emissions, would not lead to stabilization of atmospheric concentrations of GHGs. Any solution to the problem must eventually be global.

This chapter briefly outlines the basic science of the greenhouse effect and its role in maintaining the Earth's global climate, the observational evidence for climate change over the past century or more, and the projections of how much and what kind of change may occur over the next 100 years as a result of human activities. It also summarizes the political history of the climate change issue.

1.1 A Climate Primer

The ultimate source of energy driving the Earth's climate is the sun. Sunlight warms the Earth's surface, which responds by re-emitting energy in the form of infrared radiation (heat). If the Earth had no atmosphere, the sun would warm the surface to only about -18 degrees Celsius ($^{\circ}\text{C}$), a temperature well below the freezing point of water and probably unable to support life. But observations show that the average surface temperature is actually about 15°C . This 33°C difference is due to a natural atmospheric phenomenon known as the greenhouse effect.

1.1.1 The greenhouse effect

A number of gases in the atmosphere are transparent to incoming sunlight but absorb the heat emitted from the Earth's surface and reradiate it in all directions, including back toward the surface. As a result, the surface is warmed more than it would have been in the absence of these gases, which serve a purpose similar to that of the glass walls of a greenhouse.[1]

The greenhouse effect forms the backbone of current understanding of the Earth's climatic history. For example, models of the life cycles of stars indicate that the sun's output has grown by about 30% over the Earth's 4.5-billion-year history. The early Earth therefore received much less solar radiation than it does today. If the atmospheric composition had been the same then as it is now, the planet would have been frozen at the same time that life was beginning to evolve. This apparent impossibility is known as the "faint young sun paradox" (Sagan and Mullen, 1972; Newman and Rood, 1977; Kasting *et al.*, 1988). Currently, the widely accepted solution to this paradox is that the early atmosphere contained much higher concentrations of GHGs than it does today, a proposition for which there is geologic evidence. The additional warming effect these gases provided kept the surface warm enough to allow life to evolve. Since that time, the Earth's broad climatic trends have been regulated in part by processes controlling the atmospheric concentration of GHGs (Walker *et al.*, 1981).

Global warming refers to the enhancement of the natural greenhouse effect due to rising atmospheric concentrations of GHGs resulting from human activities. That the Earth will warm because of the enhanced greenhouse effect is not in dispute.

Debate centers on how much and how fast warming will occur, and how serious the consequences of such changes might be. A small amount of gradual climate change may not be much cause for concern, but if the changes are larger and faster, impacts could be severe. Additionally, small shifts in global mean temperature may prove less important than larger shifts in the frequency and intensity of extreme climate events.

1.1.2 The greenhouse gases

Although GHGs play a critical role in maintaining surface temperature, they make up less than 1% of the atmosphere by volume. The most abundant GHG is water vapor. Although it is responsible for most of the Earth's natural greenhouse effect, its global abundance is not directly influenced by human activity and so it is not considered an anthropogenic (human-generated) GHG. However, its abundance is indirectly affected by anthropogenic emissions in an important way. An increase in the atmospheric concentration of GHGs traps heat near the surface. Part of that heat is expended by evaporating more water, and part warms the surface and lower atmosphere. Assuming relative humidity remains roughly constant, a warmer atmosphere will hold a larger quantity of water, which will be provided by the increased evaporation.[2] Because water vapor is a GHG, this increase in the absolute humidity will further warm the surface and atmosphere, leading to further water evaporation and so on in a positive feedback loop known as the water vapor feedback. The water vapor feedback is a key component of the estimated sensitivity of the climate system to changes in GHG concentrations.

The most important anthropogenic GHGs are carbon dioxide (CO₂), methane (CH₄), nitrous oxide (N₂O), and halocarbons. *Table 1.1* summarizes some key characteristics of these gases.

Carbon Dioxide (CO₂)

Carbon dioxide is, after water vapor, the second most important gas producing the Earth's natural greenhouse effect. Large natural fluxes of CO₂ between the atmosphere and the oceans, land plants, and soils have maintained relatively constant atmospheric concentrations over the past ten thousand years. However, over the past two centuries human activity has released CO₂ at increasing rates. The "pioneer effect" (deforestation in the Northern Hemisphere) was probably the first significant contributor, but with the acceleration of the Industrial Revolution in the 19th century the main source quickly became the burning of fossil fuels such as coal, oil, and natural gas. Over the past several decades, tropical deforestation has also become a significant source, making up about 25% of recent global emissions (Houghton, 1999).

Table 1.1. Summary of main anthropogenic greenhouse gases.

Gas	Concentration (ppb) ^a		Current growth (%/yr)	Lifetime ^b (years)	Radiative forcing (per molecule, relative to CO ₂)	100-year global warming potential (relative to CO ₂)
	Preind.	1994				
CO ₂	~280,000	358,000	0.5	~100 (50–75%) ^c >10 ⁴ (25–50%)	1	1
CH ₄	~700	1,720	0.5	12.2	21	21
N ₂ O	~275	312	0.3	120	206	310
CFC-12 ^d	0	0.503	1.4	102	15,600	6,200–7,100

^appb = parts per billion.

^bLifetime is defined as the average length of time a present emission will continue to affect atmospheric concentrations.

^cFrom O'Neill *et al.* (1997). The atmosphere's response to a CO₂ emission has a distinctly dual nature: at least half the effect of the emission is removed in about 100 years, while the remainder persists for tens of thousands of years or more. The exact fractions and time scales of persistence depend on the assumed future concentration scenario.

^dCFC-12 is used here as a representative example of the chlorofluorocarbons (CFCs), an important subclass of the halocarbons.

Source: Schimel *et al.* (1996), except as indicated in notes.

Direct measurements of atmospheric CO₂ over the past several decades, as well as measurements obtained from older air trapped in ice sheets, show that CO₂ concentrations have increased about 30% from their preindustrial level (*Figure 1.1*). A number of lines of evidence demonstrate beyond doubt that this rise has been caused by anthropogenic CO₂ emissions, and measurements of the CO₂ content of air bubbles trapped in ice sheets show that the current atmospheric level is higher than at any time in the past 420,000 years (Petit *et al.*, 1999).[3] Once in the atmosphere, CO₂ is not removed by chemical decomposition, but instead is redistributed to other carbon reservoirs. The processes controlling these transfers operate over a wide range of time scales. As a result, one-quarter to one-half of the effect of CO₂ emitted into the atmosphere is removed so slowly that it may be considered essentially permanent, while the remainder is removed in about 100 years (see *Table 1.1*).

Methane (CH₄)

Methane also contributes to the Earth's natural greenhouse effect and is emitted from a range of natural sources, most notably as a product of anaerobic respiration in wetlands. However, current anthropogenic sources are estimated to be about twice as large as natural sources and have led to more than a doubling of preindustrial atmospheric CH₄ concentrations. Methane is released from a wide array of human activities; uncertainty in the magnitude of these fluxes is higher

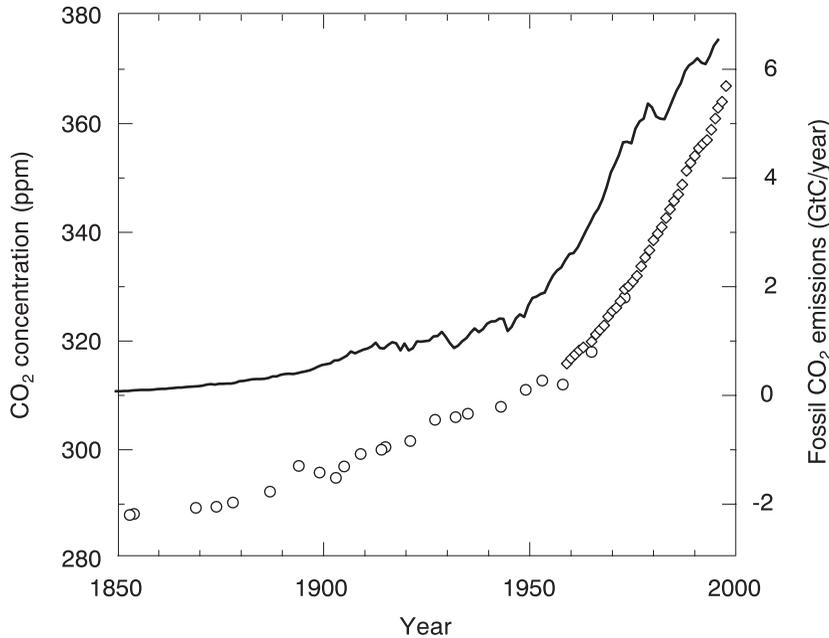


Figure 1.1. Atmospheric CO₂ concentrations (1850–1998) and emissions from fossil fuel use (1850–1996). The continuous curve shows annual CO₂ emissions from fossil fuels in gigatons carbon per year (GtC/year); diamonds are annual means of direct measurements of atmospheric CO₂ in parts per million (ppm); circles are measurements of CO₂ concentrations in air trapped in ice cores. Sources: Keeling, 1994; Neftel *et al.*, 1997; Keeling and Whorf, 1999; Marland *et al.*, 1999.

than uncertainty associated with anthropogenic CO₂ fluxes. Principal sources include livestock such as cattle and sheep (whose digestive systems use fermentation processes that produce CH₄), leakages from natural gas pipelines and coal mines, anaerobic respiration in rice paddies, and biomass burning, with a handful of other sources also contributing (Prather *et al.*, 1995).

Methane has a relatively short atmospheric lifetime of about 12 years; its effect on climate is therefore shorter lived than that of CO₂. It is removed from the atmosphere predominantly by chemical reactions with the hydroxyl radical (OH) and to a lesser extent by soil uptake. The growth rate of CH₄ in the atmosphere has slowed significantly over the past decade (Dlugokencky *et al.*, 1998). While the cause of this decline is uncertain, possibilities include the effect of temperature change (Bekki and Law, 1997), changes in anthropogenic emissions (Law and Nisbet, 1996), and the climatic influence of the 1991 Mt. Pinatubo volcanic eruption.

Nitrous Oxide (N₂O)

Nitrous oxide concentrations have increased nearly 15% due to human activity. Sources of N₂O have not been well quantified, but the largest natural fluxes to the atmosphere are thought to be from soils (particularly in tropical forests), with a smaller but significant contribution from the oceans. Anthropogenic sources are estimated to be about two-thirds as large as natural sources and are dominated by fluxes from nitrogen-fertilized agricultural fields (Smil, 1999), with additional contributions from a number of other sources including biomass burning, some industrial processes, and cattle and feed lots (Prather *et al.*, 1995). Nitrous oxide is removed from the atmosphere mainly by reactions with sunlight in the stratosphere (upper atmosphere) and has a relatively long lifetime of about 120 years.

Recent attention has focused on the potential for higher emission rates from tropical (relative to temperate) soils following fertilizer use or deposition of NO_x from burning of fossil fuels (Hall and Matson, 1999; Bouwman, 1998). Combined with the anticipation of a further shift in fertilizer and fossil fuel use toward tropical regions, this difference could significantly increase current projections of anthropogenic emissions.

Halocarbons

Halocarbons comprise a number of chlorine-, fluorine-, or bromine-containing gases with generally powerful heat-trapping properties. Halocarbons include chlorofluorocarbons (CFCs) and related compounds such as hydrochlorofluorocarbons (HCFCs), hydrofluorocarbons (HFCs), and perfluorocarbons (PFCs), as well as carbon tetrachloride, sulfur hexafluoride, and methyl chloroform. CFCs have historically been the most important halocarbons in terms of their warming effect. CFCs have no natural sources; they are synthetic compounds used as refrigerants, propellants, blowing agents in the manufacture of foams, and cleaning agents in the production of electronic components. They are essentially inert in the troposphere (lower atmosphere) and therefore have lifetimes of 50 years or more; however, they break down in the stratosphere, where they are the main culprit in stratospheric ozone depletion. Because ozone is itself a GHG, this depletion offsets some of the warming effect of the CFCs, adding a significant uncertainty to the net CFC warming contribution (Schimel *et al.*, 1996). Emissions of CFCs have fallen as the industrialized nations, which are responsible for most of the global total, have complied with the Montreal Protocol on Substances that Deplete the Ozone Layer and its amendments for phasing out production of these chemicals. However, while the HCFCs and HFCs often used as replacements are less efficient ozone depleters, they are still effective GHGs.

Tropospheric Ozone (O₃)

Although about 90% of total ozone is present in the stratosphere, tropospheric ozone (O₃) also has an important effect on climate. This “low-level” ozone is produced through the oxidation of CH₄ as well as through reactions involving a number of precursor gases (e.g., carbon monoxide, nitrogen oxides, and non-methane hydrocarbons), which are themselves produced in part by human activities like biomass burning and fossil fuel combustion. In addition, ozone is transported into the troposphere from the stratosphere. It is destroyed by reaction with ultraviolet light and hydroxyl radicals. Because production and destruction processes vary widely through space and time, ozone concentrations vary with geographical location, altitude, season, and even time of day, making estimation of global trends difficult. Available measurements and modeling studies suggest that tropospheric ozone concentrations in the Northern Hemisphere, where anthropogenic sources are greatest, may have doubled since preindustrial times. Globally averaged, the warming effect of increases in tropospheric ozone levels has probably enhanced the effect due to other GHGs by 10% or more (Stevenson *et al.*, 1998), although effects from ozone have been highly regional.

Radiative Forcing

Because the GHGs have different lifetimes and heat-trapping properties, changes in their abundances affect the Earth's energy balance to different degrees. Any change affecting this balance, whether it is a change in GHG concentrations or a change in the intensity of incoming sunlight, is described as a “radiative forcing.” The sixth column of *Table 1.1* shows the relative radiative forcings of the gases on a molecular basis. Molecule for molecule, CH₄, N₂O, and CFCs are much more effective heat-trapping gases than CO₂. The last column gives their 100-year global warming potentials (GWPs), an index that takes into account the different lifetimes of the gases by measuring the cumulative radiative forcing each would contribute over a 100-year period following equal-weight emissions. The values in the table show that, ton for ton, CO₂ emissions will contribute much less to radiative forcing over the next century than will emissions of other GHGs. This comparison should be viewed with some caution since GWPs are subject to a number of uncertainties and are sensitive to, among other things, the choice of time horizon and the future atmospheric concentrations of GHGs (Wuebbles *et al.*, 1995). Nonetheless, they provide an approximate measure of relative radiative forcing effects over a specified time period.[4]

Despite the relatively high GWPs of the other gases, CO₂ is responsible for most of the warming effect resulting from current excess concentrations of all GHGs (see *Table 1.2*). This dominance results from the much greater absolute

Table 1.2. Contribution to radiative forcing due to excess greenhouse gas concentrations.

Gas	1992 radiative forcing ^a (W/m ²)	Percentage of total 1992 greenhouse gas forcing
CO ₂	1.56	65
CH ₄	0.47	20
N ₂ O	0.14	6
CFCs and HCFCs	0.15 ^b	6
Other	0.03	1

^a Global average forcing for tropospheric ozone is estimated to be 0.4 watts per square meter (W/m²), but since its effect is highly regional it is difficult to compare with the well-mixed gases and is not included here.

^b Includes offsetting negative forcing due to stratospheric ozone depletion.

Source: Schimel *et al.*, 1996.

increase in the atmospheric CO₂ level relative to other gases. For the same reason, its share of responsibility for GHG forcing is expected to grow in the future, although the aggregate warming effect of the other gases is expected to be significant as well.

1.1.3 Other factors influencing climate change

Any climate change induced by the accumulation of anthropogenic GHGs will occur against a background of changes due to a number of other factors. Some of these factors will amplify greenhouse warming; others may offset part of it. Some may, at different times, do both. Each also complicates the task of detecting the anthropogenic component of observed climate change.

Tropospheric Aerosols

Aerosols are small airborne particles that both absorb and reflect radiation and therefore can affect climate directly. They can also affect climate indirectly by altering cloud cover. Aerosols are emitted naturally and by human activity either as particles or as gases that eventually form droplets. Natural sources include soil dust and spray from the oceans; the most important anthropogenic aerosols are sulfates formed from sulfur dioxide gas produced by the combustion of fossil fuels. Incomplete fuel combustion and burning of biomass also contribute significant aerosol fluxes in the form of soot and organic carbon (Jonas *et al.*, 1995).

Different aerosols are thought to have different net effects on climate. Soot from burning of fossil fuels is estimated to exert a positive forcing, while aerosols from burning of biomass and from sulfur emissions are thought to have a cooling

effect. Taken together, aerosols are thought to exert a cooling effect on climate and therefore may mask some of the effect of GHGs (Shine and Forster, 1999). Because they generally spend just a few days in the troposphere before returning to the surface in rainfall or through dry deposition, aerosols are not distributed uniformly in the atmosphere but are concentrated near source regions (such as heavily industrialized areas). Their climate effect is therefore regional as well, which makes determining their global contribution to the enhanced greenhouse effect difficult. It is estimated that anthropogenic aerosols may exert a cooling effect that is, on average, about 20% as large as the warming effect of GHGs released through human activity, but in some locations they could offset all of the GHG forcing (Schimel *et al.*, 1996). The uncertainty in this estimate is large, however, and some researchers have argued that because aerosols could play an even larger role, a wider range of future forcing should be considered in climate change scenarios (Hansen *et al.*, 1998a).

Volcanic Activity

Strong volcanic eruptions can inject large amounts of sulfurous gases into the stratosphere, where they are transformed into sulfate aerosols. Because the stratosphere mixes slowly with the troposphere (where aerosols are removed from the atmosphere), stratospheric aerosols disperse globally and persist for several years, reflecting incoming solar radiation and cooling the climate (Minnis *et al.*, 1993). The 1991 eruption of Mount Pinatubo in the Philippines, for example, temporarily reduced global average temperature by several tenths of a degree Celsius. By the end of 1994, however, the aerosols from Pinatubo had been removed from the atmosphere and the dip in temperature had disappeared (see *Figure 1.2* on page 13). Because the effect of volcanos on climate is short lived, future eruptions will not affect long-term climate trends resulting from global warming.

Solar Variations

The sun's output varies slightly over an 11-year cycle associated with sunspots, and presumably over longer time periods as well. Considerable attention has been focused on possible connections between solar cycles and climate, but while intriguing correlations have been found, these efforts have been unsuccessful in proposing a plausible mechanism capable of magnifying the weak magnitude of the solar variations into significant changes in climate (Shine *et al.*, 1995; Kerr, 1996). One possible link is through the influence of changes in solar ultraviolet radiation on stratospheric ozone, which could induce circulation changes in the troposphere large enough to affect surface temperature. However, model estimates of this effect are small (Shindell *et al.*, 1999a). Therefore, although trends in solar