UNDERSTANDING CLIMATE SCIENCE

2

Key points

The Review takes as its starting point, on the balance of probabilities and not as a matter of belief, the majority opinion of the Australian and international scientific communities that human activities resulted in substantial global warming from the mid-20th century, and that continued growth in greenhouse gas concentrations caused by human-induced emissions would generate high risks of dangerous climate change.

A natural carbon cycle converts the sun's energy and atmospheric carbon into organic matter through plants and algae, and stores it in the earth's crust and oceans. Stabilisation of carbon dioxide concentrations in the atmosphere requires the rate of greenhouse gas emissions to fall to the rate of natural sequestration.

There are many uncertainties around the mean expectations from the science, with the possibility of outcomes that are either more benign—or catastrophic.

Climate change policy must begin with the science. When people who have no background in climate science seek to apply scientific perspectives to policy, they are struck by the qualified and contested nature of the material with which they have to work. Part of the uncertainty derives from the complexity of the scientific issues. In the public discussion of the science, additional complexity derives from the enormity of the possible consequences, which encourages a millennial perspective. Part derives from the large effects of possible policy responses on levels and distributions of income, inviting intense and focused involvement in the discussion by those with vested interests.

The Review is not in a position to independently evaluate the considerable body of scientific knowledge; it takes as a starting point the majority opinion of the Australian and international scientific communities that human-induced climate change is happening, will intensify if greenhouse gas emissions continue to increase, and could impose large costs on human civilisation.

This chapter draws extensively on the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, and on detailed reports prepared by Australian scientists, research published since the IPCC Fourth Assessment Report and work commissioned specifically for the Review. It aims to build an understanding of the way humans can influence the climate and the limitations in our current understanding of the climate system, and introduces key terminology and concepts relevant to people who are interested in climate change policy.

In drawing on the work of the IPCC, and the large majority of Australian scientists who are comfortable working within that tradition, we are still faced with immense uncertainties, which have informed (and at times bedevilled) the Review's analysis. These perspectives may cease to be the mainstream as the development of climate science proceeds and the uncertainties narrow. At this time, the Review believes it is appropriate to give the main weight to them.

2.1 The earth's atmosphere

2.1.1 A changing atmosphere

The earth is surrounded by an atmosphere that protects it from high-energy radiation and absorbs heat to provide a moderate climate that supports life.

The earth's atmosphere has not always been the same as it is today. Billions of years ago, the atmosphere was composed mainly of ammonia, water vapour and methane, but over time release of gases from within the planet through volcanic eruptions and discharge of gases from ocean vents changed conditions so that carbon monoxide, carbon dioxide and nitrogen became dominant.

Around 3.5 billion years ago, algae-like organisms first began to use the energy from the sun to convert carbon dioxide from the air into carbohydrates, with oxygen as a by-product. Over time, oxygen accumulated and reached the current levels of 21 per cent of atmospheric volume.

2.1.2 The natural greenhouse effect

The earth's atmosphere acts like the roof of a greenhouse, allowing short-wavelength (visible) solar radiation from the sun to reach the surface, but absorbing the long-wavelength heat that is emitted back. This process is referred to as 'the greenhouse effect', and the gases that absorb the emitted heat are known as greenhouse gases. The main naturally occurring greenhouse gases are water vapour, carbon dioxide, methane, nitrous oxide and ozone (see section 2.3).

Compared to nitrogen and oxygen, which together comprise 99 per cent of the volume of the atmosphere, greenhouse gases occur only at trace levels, making up just 0.1 per cent of the atmosphere by volume (IPCC 2001a). Despite this, their presence means that the earth has an average global surface temperature of about 14°C—about 33°C warmer than if there were no greenhouse gases at all (IPCC 2007a: 946).

2.1.3 Changes in greenhouse gases and temperature over time

Records of carbon dioxide concentrations taken from proxy measures such as fossil plants and ice core data are available for the last 400 million years. These records indicate that atmospheric concentrations of carbon dioxide have fluctuated between about 180 ppm (parts per million by volume), levels similar to pre-industrial concentrations of 280 ppm, and levels higher than 4000 ppm (Royer 2006; IPCC 2007a: 444).

There is a high degree of uncertainty in historical measurements of temperature before modern times, which must be estimated from a range of indirect sources such as annual growth rings in trees and corals, and small fossils in ocean and lake sediments. During the last 2.5 million years, climate records document a saw-tooth pattern of changes in temperature and ice volume. In the last 600 000 years the fluctuations show a periodicity of around 100 000 years (Ruddiman 2008). The periods when polar ice caps were greatly expanded, which resulted in ice sheets covering large parts of the northern continents, are known as glacial periods or ice ages. Those without extended polar ice caps are known as interglacials. The last glacial maximum occurred 21 000 years ago. For the last 10 000 years, the earth has been in an interglacial period (IPCC 2007a: 447).

Glacial periods occur when solar radiation is reduced, and interglacials when solar radiation is more intense. Consistent changes in the intensity of solar radiation from regular variations in the shape of the earth's orbit and the tilt of its axis, as well as the sunspot cycle, are seen as drivers of these cyclical climate changes (Ruddiman 2008).

Fluctuations in carbon dioxide and methane concentrations have occurred in the 600 000-year period before the present, but the role of greenhouse gases in contributing or responding to glacial–interglacial fluctuations is complex.

2.1.4 How are the recent changes different?

Why are we so concerned about the current changes in climate and greenhouse gas concentrations if they have fluctuated so much over the earth's history?

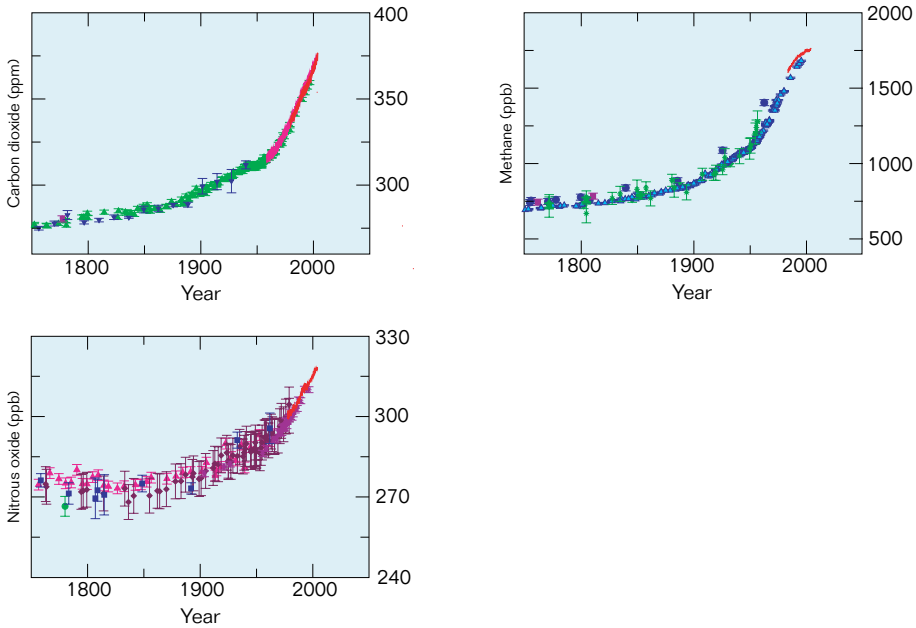
Apart from the earliest identified hominids, the history of our species has been within the period of relatively low carbon dioxide concentrations. Modern forms of our species first appeared only around 200 000 years ago. The development of agriculture, large-scale social organisation, writing, cities and the behaviours we associate with modern civilisation has occurred only in the last 10 000 years. The period in which human civilisation has developed, located within an interglacial period known as the Holocene, has been one of equable and reasonably stable temperatures.

Concentrations of carbon dioxide now exceed the natural range of the last two million years by 25 per cent, of methane by 120 per cent and of nitrous oxide by 9 per cent (IPCC 2007a: 447). The anthropogenically driven rise in carbon dioxide since the beginning of the industrial revolution (around 100 ppm) is about double the normal 'operating range' of carbon dioxide during glacial–interglacial cycling

(180–280 ppm) (Steffen et al. 2004). Trends in the atmospheric concentration of carbon dioxide, methane and nitrous oxide for the last 250 years are shown in Figure 2.1—it is not just the *magnitude* of the post-industrial increase in greenhouse gas concentrations that is unusual, but also the *rate* at which it has occurred.

There is high natural variability in global temperatures in recent millennia. The current high temperatures, though unusual over the last 1000 years, are not unusual on longer time frames. However, the rapid rate of the current warming is highly unusual in the context of the past millennium (CASPI 2007).

Figure 2.1 Trends in atmospheric concentrations of carbon dioxide, methane and nitrous oxide since 1750



Note: Measurements are shown from ice cores (symbols with different colours for different studies) and atmospheric samples (red lines).

Source: IPCC (2007a: 3), formatted for this publication.

2.1.5 Is human activity causing the earth to warm?

The development of climate-related research and modelling has allowed increasingly more definitive assessments of the human impacts on climate. The IPCC Fourth Assessment Report noted an improvement in the scientific understanding of the influence of human activity on climate change. The report concluded that the warming of the climate system is 'unequivocal' (IPCC 2007a: 5), that there is a greater than 90 per cent chance that 'the global average net effect of human activities since 1750 has been one of warming' (IPCC 2007a: 3). Confidence in the

influence of humans on other elements of climate change, such as droughts and severe weather events, is not as high, but is increasing as modelling techniques and observation databases improve.

2.2 Understanding climate change

2.2.1 Definitions of climate change

The IPCC (2007a: 943) defines climate change as 'a change in the state of the climate that can be identified (for example, by using statistical tests) by changes in the mean and/or variability of its properties, and that persists for an extended period, typically decades or longer'. Climate change may be due to natural internal processes or external influences, or to persistent anthropogenic changes in the composition of the atmosphere or land use.

By contrast, the United Nations Framework Convention on Climate Change defines climate change as 'change of climate which is attributed directly or indirectly to human activity that alters the composition of the global atmosphere and which is in addition to natural climate variability observed over comparable time periods' (UN 1992).

This report uses the IPCC definition, so that the discussion of climate change includes changes to the climate caused by natural phenomena such as volcanic eruptions.

2.2.2 The climate system

In a narrow sense, climate can be defined as the 'average weather' and described in terms of the mean and range of variability of natural factors such as temperature, rainfall and wind speed.

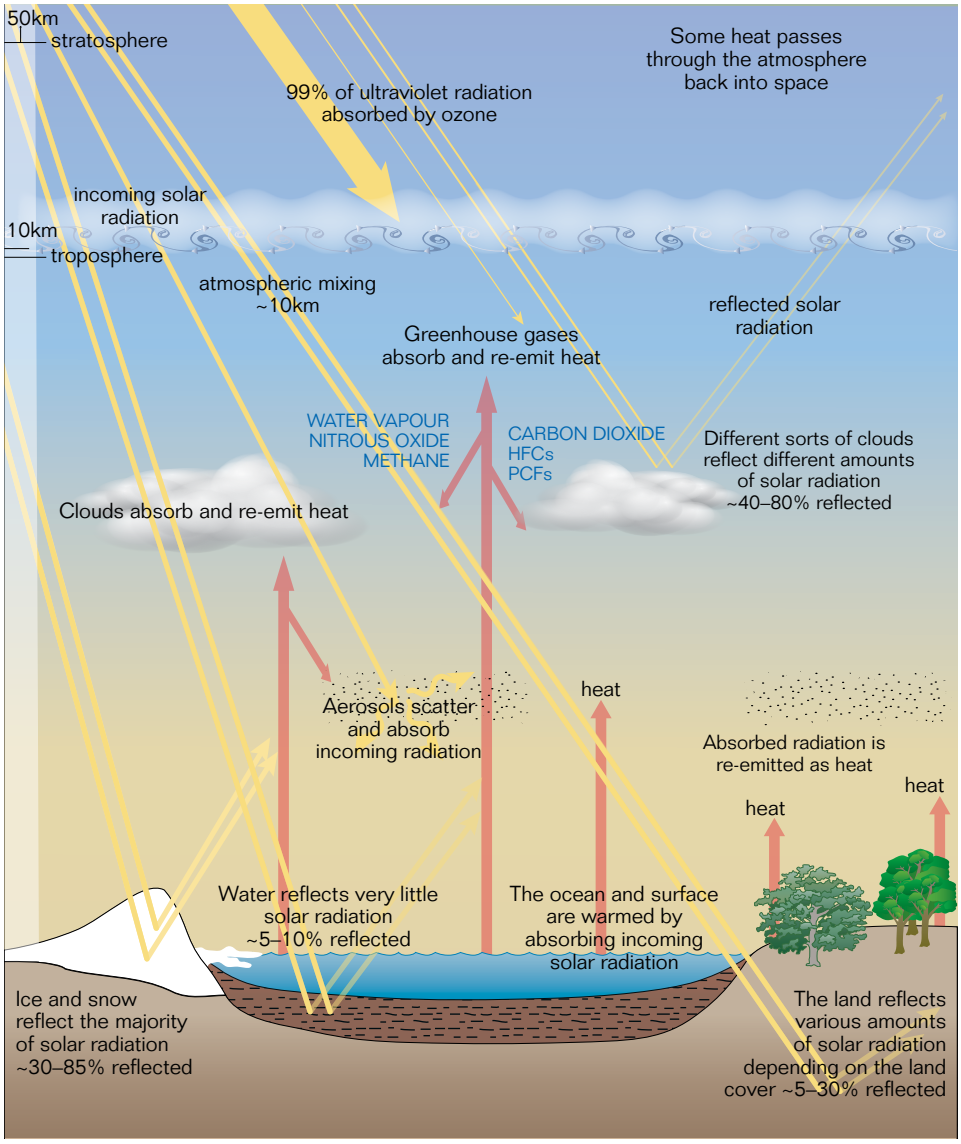
More broadly, the climate is a system involving highly complex interactions between the atmosphere, the oceans, the water cycle, ice, snow and frozen ground, the land surface and living organisms. This system changes over time in response to internal dynamics and variations in external influences such as volcanic eruptions and solar radiation.

The atmospheric component is the most unstable and rapidly changing part of the climate system. The atmosphere is divided into five layers with different temperature characteristics. The lower two—the troposphere and the stratosphere—have the most influence on the climate system.

The troposphere extends from the surface of the earth to an altitude of between 10 and 16 km. Clouds and weather phenomena occur in the troposphere, and greenhouse gases absorb heat radiated from the earth.

The stratosphere, which extends from the boundary of the troposphere to an altitude of around 50 km, is the second layer of the atmosphere. The stratosphere holds a natural layer of high ozone concentrations, which absorb ultraviolet radiation from the sun (Figure 2.2). The balance of energy between the layers of the atmosphere is a major driver of atmospheric and ocean circulation, which leads to weather and climate patterns (IPCC 2007a: 610).

Figure 2.2 A stylised model of the natural greenhouse effect and other influences on the energy balance of the climate system



2.2.3 The energy balance of the climate system

Energy enters the climate system as visible and ultraviolet radiation from the sun, and is absorbed, scattered or reflected, re-emitted as heat, transferred between different elements of the system, used by organisms and emitted back into space.

The balance between the energy entering and leaving the system is what determines whether the earth gets warmer or cooler, or stays the same. Changes in the strength of radiation from the sun change the energy that enters the system,

while the flow of energy within the system and the amount that is released can be modified by a wide range of factors.

The composition of gases in the atmosphere plays a big part in the amount of heat that is retained in the climate system, but there are many other influences. Aerosols (Box 2.1) can scatter the incoming radiation so that it never reaches the surface, and can also cause changes in cloud cover. More clouds will reflect more sunlight from their upper surfaces, but also absorb more heat radiating from the earth. Variations in land cover affect the amount of sunlight that is reflected from the surface (the 'albedo effect'), and how much is absorbed and re-emitted as heat.

Human activity can affect the energy balance of the climate system in a number of ways. Examples include changes in land use; emissions of aerosols and other pollutants; emissions of greenhouse gases from activities such as agriculture, energy production, industry and land clearing; emissions of other pollutants that react in the atmosphere to form greenhouse gases; and influences on cloud formation through aviation.

Box 2.1 Aerosols

Aerosols are tiny particles or droplets in the atmosphere, including sulphates, ash, soot, dust and sea salt, that can be natural or anthropogenic in origin. A major natural source of aerosols is volcanic eruptions. The major anthropogenic source is fossil fuel combustion.

Aerosols generally create a cooling effect, but there is great uncertainty about the magnitude of this effect. Black carbon, or soot, has a warming effect because it absorbs solar radiation. Recent research suggests this effect may be considerably higher than previously estimated (Ramanathan & Carmichael 2008). Because the lifetime of aerosols in the atmosphere is much shorter than that of greenhouse gases, the effects are more likely to be felt in the region in which the aerosol is produced.

2.2.4 Factors leading to warming of the climate system

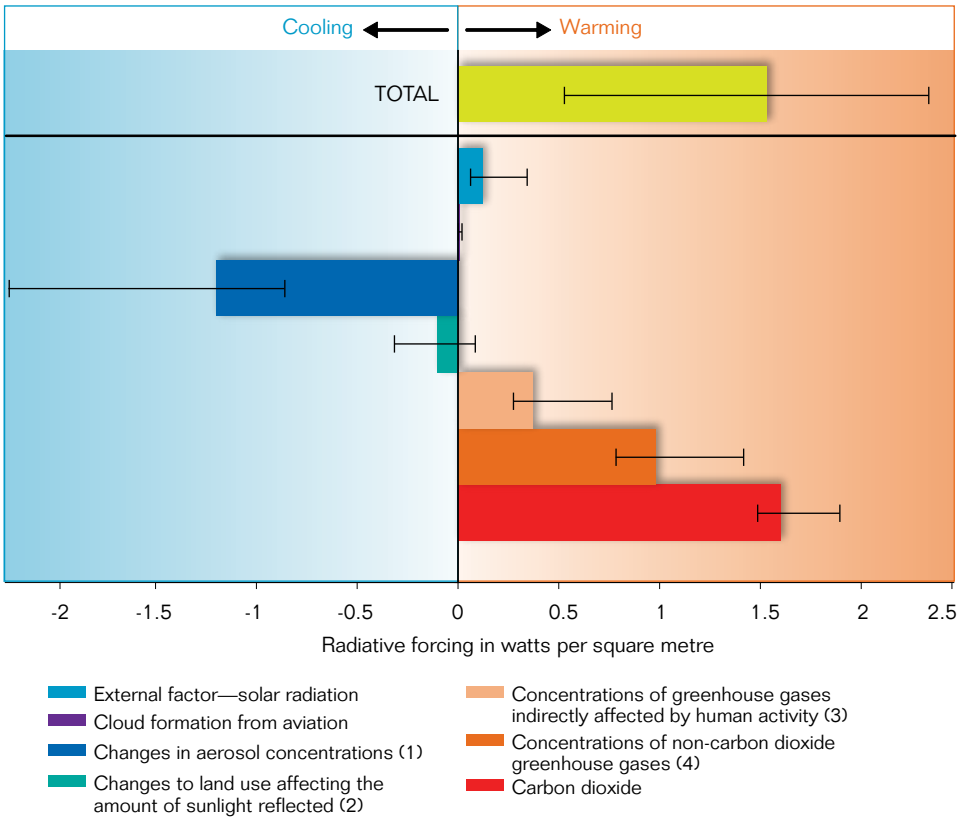
The warming of the climate system evident in the last half century is a result of the cumulative effect of all the natural and human drivers that influence the amount of warming or cooling in the system. The contribution of different factors leading to an overall warming of the atmosphere since 1750 is shown in Figure 2.3.

The dominant influence since 1750 has been an increase in concentrations of carbon dioxide. Aerosols have had a net cooling influence, although this effect is poorly understood. Natural variability in solar radiation has had a small warming influence, but there is a high level of uncertainty in the magnitude of the effect (IPCC 2007a: 192).

Even if there were no further human-induced increases in aerosols and greenhouse gases, the long-lived greenhouse gases would remain for hundreds and even thousands of years, leading to continued warming. Aerosols are removed from the atmosphere over much shorter periods, so their cooling effect would

no longer be present. Therefore, in the long term, the major influence of humans on the climate will be through activities that lead to increased concentrations of greenhouse gases in the atmosphere.

Figure 2.3 Contribution of human and natural factors to warming since 1750



Note: 'Radiative forcing' is a measure of the induced change to the energy balance of the atmosphere. Warming and cooling influences are indicated by positive and negative values respectively. When elements are grouped, uncertainty bands are approximated from the highest uncertainty in an individual element.

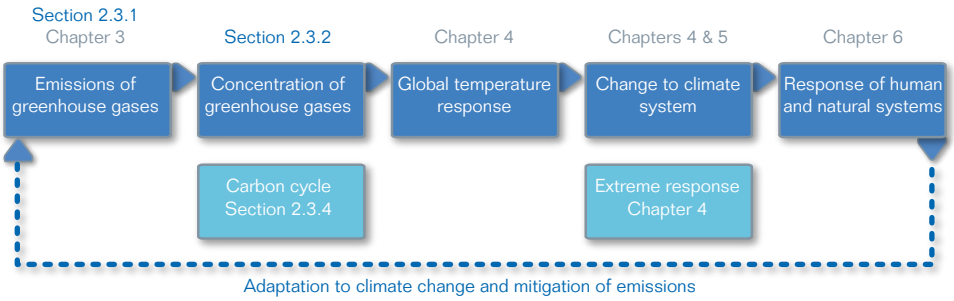
(1) Includes both the direct effect and the cloud albedo effect.
(2) Includes the cooling effect of changes to land use and the warming effect of black carbon on snow.
(3) Includes tropospheric and stratospheric ozone and stratospheric water vapour.
(4) Includes methane, nitrous oxide, HFCs, PFCs and sulphur hexafluoride.

Source: Based on IPCC (2007a: 204).

2.3 Linking emissions and climate change

The high natural variability and complex internal interactions create uncertainty in the way the climate will respond to increased emissions. Figure 2.4 illustrates the relationship between emissions from human activities and climate change as a causal chain. This causal chain does not explicitly include the feedbacks and non-linearities in the climate system that are important in its response to human forcings.

Figure 2.4 Steps in the causal chain of greenhouse gas emissions leading to climate change



2.3.1 Emissions of greenhouse gases from human activities

The greenhouse gases with the greatest influence on warming of the atmosphere are water vapour (H_2O), carbon dioxide (CO_2), nitrous oxide (N_2O), methane (CH_4) and ozone (O_3). In addition, there is a range of human-made halocarbons (such as perfluorocarbons (PFCs), hydrofluorocarbons (HFCs), chlorofluorocarbons (CFCs)) and sulphur hexafluoride (SF_6) that are present in small amounts but are potent and contribute significantly to the total warming. Table 2.1 summarises the natural and anthropogenic sources of these gases.

Not all of these gases are directly emitted through human activities. Humans have less direct control over gases such as water vapour and ozone, although concentrations of these gases can be affected by human emissions of other reactive gases.

Table 2.1 Sources of greenhouse gases

Gas	Natural sources	Dominant anthropogenic sources
Carbon dioxide	Respiration from living organisms Volcanic eruptions Forest fires Decomposition of dead animals and plants Outgassing from the ocean	Combustion of fossil fuels and cement manufacture (more than 75 per cent of the increase in concentration since pre-industrial times) (IPCC 2007a: 512) Land-use changes (deforestation and changing agricultural practices)
Methane	Oceans Termites Natural wetlands Hydrates	Fossil fuel mining Vegetation burning Waste treatment Rice cultivation Ruminant livestock Landfill
Nitrous oxide	Processes in soils and oceans Oxidation of ammonia in the atmosphere	Nitrogenous fertiliser use Biomass burning Management of livestock manure Fossil fuel combustion Industrial activities such as nylon manufacture

Table 2.1 Sources of greenhouse gases *(continued)*

Gas	Natural sources	Dominant anthropogenic sources
Fluorinated gases		
HFCs	Some PFCs and all HFCs have no detected natural sources Other PFCs, and SF_6, are present in small amounts in the Earth's crust and released into the atmosphere through volcanic activity (Harnisch et al. 2000)	Refrigeration Air conditioning Solvents Fire retardants Foam manufacture Aerosol propellants
PFCs		Aluminium production
Sulphur hexafluoride		Electricity supply industry (switches and high-voltage systems)
Water vapour	The amount of water vapour in the atmosphere is a function of temperature and tends to fluctuate regionally and on short timescales	Irrigation, artificial dams and lakes, and fossil fuel production—less than 1 per cent of emissions from natural sources (IPCC 2007a: 28)
CFCs and HCFCs	No known natural sources	Propellants in aerosol cans Refrigerants in refrigerators and air conditioners Manufacture of foam packaging
Tropospheric ozone	Chemical reaction between other gases ('precursor species'), including carbon monoxide, methane and nitrogen oxides	Humans have limited direct influence, but influence concentrations through the emission of precursor species such as methane, nitrogen oxides and organic compounds from industry, power generation and transport

2.3.2 Accumulation of greenhouse gases in the atmosphere

The accumulation of greenhouse gases in the atmosphere that leads to warming is a function of both the rate of emissions and the rate of natural removal from the atmosphere.

Each greenhouse gas has specific characteristics that affect how long it stays in the atmosphere. The 'lifetime' of a gas in the atmosphere is a general term used for various timescales that characterise the rate of processes affecting the destruction or removal of trace gases.

With the exception of carbon dioxide, physical and chemical processes generally remove a specific fraction of the amount of a gas in the atmosphere each year. In some cases, the removal rate may vary with atmospheric properties such as temperature or background chemical conditions (IPCC 2007a: 23).

The lifetime of a gas affects the speed at which concentrations of that gas change following changes in the level of emissions. Gases with short lifetimes such as methane respond quickly to changes in emissions. Other gases with longer

lifetimes respond very slowly. Over the course of a century, half of the carbon dioxide emitted in any one year will be removed, but around 20 per cent will remain in the atmosphere for millennia (IPCC 2007a: 824).

Carbon dioxide

After water vapour, carbon dioxide is the most abundant greenhouse gas in the atmosphere. Most gases are either removed from the atmosphere by chemical reaction or destroyed by ultraviolet radiation. Carbon dioxide, however, is very stable in the atmosphere. When it enters the atmosphere, carbon dioxide exchanges rapidly with plants and the surface ocean, and is then redistributed over hundreds to thousands of years through various forms of carbon storage, or 'sinks', as part of the carbon cycle (see section 2.3.4).

The concentration of carbon dioxide in the atmosphere increased from about 280 ppm in 1750 to 383 ppm in 2007. Over the last 10 years, carbon dioxide in the atmosphere has increased at an average rate of 2 ppm per year (Tans 2008).

Methane

Methane is a reactive compound of carbon and hydrogen that is short lived in the atmosphere, where it reacts to form water vapour and carbon dioxide. Considerable methane is stored in frozen soils and as methane hydrates in ocean sediments.

The methane concentration in the atmosphere has more than doubled since pre-industrial times to about 1774 ppb (parts per billion) after a slow fluctuation between around 580 ppb and 730 ppb over the last 10 000 years. Since the early 1990s growth rates have declined, and virtually no growth occurred between 1999 and 2005 (IPCC 2007a: 27), but more recent data for 2007 show concentrations rising for the first time since 1998 (NOAA 2008).

Nitrous oxide

Nitrous oxide is relatively stable in the atmosphere, but is eventually destroyed in the stratosphere when it reacts with ultraviolet light and charged oxygen molecules.

Nitrous oxide concentrations have increased by about 18 per cent to 319 ppb since pre-industrial times.

Fluorinated gases

Generally, fluorinated gases (HFCs, PFCs, sulphur hexafluoride) are non-flammable and non-toxic and have a low boiling point, which makes them useful for a number of manufacturing processes. HFCs and PFCs have been developed as replacements for ozone-depleting gases being phased out under the Montreal Protocol on Substances that Deplete the Ozone Layer, which entered into force in 1989.

Fluorinated gases are predominantly human in origin. They are destroyed and removed from the atmosphere mainly through reaction with ultraviolet light and

other agents in the atmosphere such as chlorine; uptake in ocean surface waters; and chemical and biological degradation processes. Concentrations of these gases are relatively small but increasing rapidly (IPCC 2007a: 28).

Water vapour

Water vapour is the most abundant and important greenhouse gas in the atmosphere, accounting for around 60 per cent of the natural greenhouse effect for clear skies (IPCC 2007a: 271). Humans have a limited ability to directly influence its concentration (IPCC 2007a: 135).

Stratospheric water vapour has shown significant long-term variability. It is produced by the oxidation of methane, and the rate of reaction increases as methane concentrations rise. However, this effect is estimated to be equivalent to only about 1 to 4 per cent of the total change caused by long-lived greenhouse gases (IPCC 2007a: 28).

Human activities that contribute to warming indirectly affect the amount of water vapour in the atmosphere, because a warmer atmosphere can hold more water vapour than a cooler one. Increases in global temperature as a result of climate change are likely to affect water vapour concentrations substantially (IPCC 2007a: 135).

The lack of understanding of the way water vapour will respond to climate change, specifically its role in cloud formation, is a key factor in the uncertainty surrounding the response of the climate to increased temperatures.

Chlorofluorocarbons and hydrochlorofluorocarbons

Chlorofluorocarbons and hydrochlorofluorocarbons (CFCs and HCFCs) are human-made gases that are odourless, non-toxic, non-flammable and non-reactive. These gases are at low concentrations but have a strong warming effect: as a group they contribute about 12 per cent of the warming from long-lived greenhouse gases.

The low reactivity of CFCs and HCFCs allows them to remain in the atmosphere for long periods and cycle up to the stratosphere. Through a series of chemical reactions, CFCs destroy ozone and other similar compounds in the stratosphere.

The Montreal Protocol has caused a substantial reduction in emissions of these gases. Emissions of CFC-11 and CFC-13 decreased substantially between 1990 and 2002 (IPCC 2007a: 513), but with lifetimes of 45 and 85 years, respectively, their concentrations in the atmosphere are falling much more slowly.

The Montreal Protocol, which aims to phase out emissions of CFCs and HCFCs by 2030, has removed the need to discuss these gases in detail in the context of greenhouse gas mitigation policy.

Tropospheric ozone and precursor species

Tropospheric ozone is produced by chemical reaction between gases rather than being directly emitted. Increases in tropospheric ozone have accounted for 10–15 per cent of the positive change in the earth's energy balance since pre-industrial times (IPCC 2007a: 204). Tropospheric ozone is short-lived, and concentrations are likely to be localised around the main sources of precursor emissions.

2.3.3 How is the warming from different gases compared?

Global warming potential is an index that compares the radiative forcing from a given mass of greenhouse gas emissions to the radiative forcing caused by the same mass of carbon dioxide (CASPI 2007). Actual emissions of different gases are multiplied by their global warming potential to give a value for the mass of emissions in carbon dioxide equivalent (CO_2-e). Global warming potential depends both on the intrinsic capability of a molecule to absorb heat, and the lifetime of the gas in the atmosphere.

Global warming potential values take into account the lifetime, the concentration at the start of the time period and the warming potential of the gas. Sulphur hexafluoride has the highest global warming potential of all gases at 22 800 times that of carbon dioxide, but has a low impact on overall warming due to its low concentrations.

Global warming potential is used under the Kyoto Protocol to compare the magnitude of emissions and removals of different greenhouse gases from the atmosphere. It is also the concept used in the design and implementation of multi-gas emissions trading schemes for calculating the value of a trade between the reductions in emissions of different greenhouse gases.

2.3.4 The carbon cycle

The 'carbon cycle' refers to the transfer of carbon, in various forms, through the atmosphere, oceans, plants, animals, soils and sediments. As part of the carbon cycle, plants and algae convert carbon dioxide and water into biomass using energy from the sun (photosynthesis). Living organisms return carbon to the atmosphere when they respire, decompose or burn. Methane is released through the decomposition of plants, animals and other hydrocarbon material (fossil fuels and waste) when no oxygen is present.

Carbon sinks

The parts of the carbon cycle that store carbon in various forms are referred to as 'carbon sinks'. The majority of carbon that was present in the early atmosphere is now stored in sedimentary rocks and marine sediments. Other carbon sinks are the atmosphere, oceans, fossil fuels such as coal, petroleum and natural gas, living plants and organic matter in the soil.

Carbon dioxide dissolves in the ocean and is returned to the atmosphere through dissolution in a continuous exchange. Dissolved carbon dioxide is carried deep into the oceans through the sinking of colder water and waste and debris from dead organisms, where it is either buried or redissolves. The transfer of carbon to the deep ocean is slow. Water at intermediate depths mixes with the surface water over decades or centuries, but deep waters mix only on millennial timescales and thus provide a long-term carbon sink.

Table 2.2 provides estimates of the amount of carbon stored in different sinks in 1750 and how they changed up to the end of the 20th century.

Table 2.2 Estimates of the amount of carbon stored in different sinks in 1750 and how they have changed

Carbon sink	Gigatonnes carbon stored in 1750	Percentage of total cycling carbon	Net change in sink between 1750 and 1994	Percentage change
Atmosphere	597	1.3	165	27.6
Vegetation, soil and detritus	2 300	5.1	-39	-1.7
Fossil fuels	3 700	8.3	-244	-6.6
Surface ocean	900	2.0	18	2.0
Marine biota	3	0.0	–	–
Deeper ocean	37 100	82.9	100	0.3
Surface sediments	150	0.3	–	–
Sedimentary rocks	>66 000 000	n/a	–	–

Note: Due to the very slow exchange with other carbon sinks, percentages of total carbon do not include storage in sedimentary rocks.

Sources: For sedimentary rocks, UNEP & GRID–Arendal (2005); for all others, IPCC (2007a: Figure 7.3).

The deep and surface ocean accounts for more than 85 per cent of the carbon being cycled more actively. Although the atmosphere accounts for just over 1 per cent of carbon storage, it shows the largest percentage increase since pre-industrial times. Vegetation and soil have had a net decrease in carbon stored—a considerable loss from land-use change has been partially offset by carbon uptake by living organisms.

The major change to the carbon cycle from human activity is increased emissions of carbon dioxide to the atmosphere from the burning of fossil fuels. The rate of exchange between the ocean and the atmosphere has increased in both directions, but with a net movement to the ocean. Terrestrial ecosystems are also a significant carbon sink. However, absorption by both the ocean and land is less than emissions from fossil fuels, and almost 45 per cent of human emissions since 1750 have remained in the atmosphere. It is generally accepted that future climate change will reduce the absorptive capacity of the carbon cycle so that a larger fraction of emissions remain in the atmosphere compared to current levels (IPCC 2007a: 750).

Carbon–climate feedbacks

Carbon–climate feedbacks occur when changes in climate affect the rate of absorption or release of carbon dioxide from land and ocean sinks. Examples of climate–carbon feedbacks include the decrease in the ability of the oceans to remove carbon dioxide from the atmosphere with increasing water temperature, reduced circulation and increased acidity (IPCC 2007a: 531); and the weakening of the uptake of carbon in terrestrial sinks due to vegetation dieback and reduced growth from reduced water availability, increased soil respiration at higher temperatures and increased fire occurrence (IPCC 2007a: 527; Canadell et al. 2007).

Large positive climate–carbon feedbacks could result from the release of carbon from long-term sinks such as methane stored deep in ocean sediments and in frozen soils as temperatures increase (IPCC 2007a: 642). There is a high level of uncertainty about how the carbon cycle will respond to climate change.

2.3.5 Greenhouse gases and temperature rise

How do different greenhouse gases change the energy balance?

The radiative forcing of a greenhouse gas represents the change in the effect of that gas on the energy balance of the atmosphere. It takes into account the greenhouse gas's concentration in the atmosphere at the start of a period (in this report, pre-industrial), the amount the concentration has changed due to human activities, and the way a molecule of that gas absorbs heat. Because radiative forcing measures change between two specific points in time, the lifetime of a gas in the atmosphere has a considerable influence on future forcing.

For most gases, removal from the atmosphere is minimally influenced by changes in the climate system. However, for the calculation of carbon dioxide concentrations, the complexity of the carbon cycle and the uncertainty over how it will change over time must be taken into account.

Carbon dioxide molecules absorb heat in a particular range of wavelengths, and as concentrations increase the additional heat of those wavelengths gets absorbed. If concentrations keep growing, carbon dioxide added later will cause proportionately less warming than carbon dioxide added now. The same amount of warming will occur from a doubling from 280 ppm (pre-industrial levels) to 560 ppm as from another doubling from 560 ppm to 1120 ppm.

The forcing due to carbon dioxide, methane, nitrous oxide and halocarbons is relatively well understood. However, the contributions of ozone at different levels in the atmosphere, aerosols and linear clouds from aviation are poorly understood (CASPI 2007).

A measure commonly used in the literature and policy discussions is the concept of carbon dioxide equivalent of a gas *concentration*, measured in parts per million. This is a different but related measure to carbon dioxide equivalent *emissions* calculated using the global warming potential index. The former is the concentration of carbon dioxide that would cause the same amount of radiative

forcing as a particular concentration of a greenhouse gas. This term is often used in discussions of global stabilisation or concentration targets.

The total radiative forcing of the long-lived greenhouse gases is 2.63 (± 0.26). In terms of carbon dioxide equivalence, this equates to a concentration of around 455 ppm CO_2-e (range: 433–477 ppm CO_2-e) (IPCC 2007b: 102).

However, the warming that would result from this is offset by the cooling effects of aerosols and land-use changes, which reduce the concentration to a range of 311 to 435 ppm CO_2-e, with a central estimate of about 375 ppm CO_2-e (IPCC 2007b: 102).

Climate sensitivity

'Climate sensitivity' is the measure of the climate system's response to sustained radiative forcing. More precisely, it is the global average surface warming (measured in degrees Celsius) that will occur when the climate reaches equilibrium following a doubling of carbon dioxide concentrations above the pre-industrial value. Such doubling of carbon dioxide levels is approximately equivalent to reaching 560 ppm carbon dioxide, which is twice pre-industrial levels of 280 ppm. Climate models predict a wide range of climate sensitivities due to differing assumptions about the magnitude of feedbacks in the climate system. Feedbacks include the response of water vapour to increased temperatures, changes in cloud formation, and the implications of the melting of ice and snow for the amount of heat absorbed by the surface.

Climate sensitivity relates to the equilibrium temperature reached when all elements of the climate system have responded to induced changes. Due to the long timescale of response, this may not occur for thousands of years. The IPCC estimates that it is likely (that is, a greater than 66 per cent chance) that a doubling of carbon dioxide will lead to a long-term temperature increase of between 2°C and 4.5°C (IPCC 2007a: 12). It is considered unlikely that climate sensitivity will be less than 1.5°C. Values substantially higher than 4.5°C cannot be excluded, but these higher outcomes are less well supported (IPCC 2007a: 799). The best estimate of the IPCC is about 3°C (IPCC 2007a: 12).

When considering the impacts of climate change on human society in the coming century, and how to respond to those impacts, temperature change over shorter time frames is more relevant. The effective climate sensitivity reflects the warming occurring in the short term, and takes into account climate feedbacks at a particular time. Assumptions on the rate of warming of the oceans in different models have a considerable effect on the short-term temperature outcomes.

Climate sensitivity is the largest of the uncertainties affecting the amount of warming when a single future pathway of greenhouse gases is selected (IPCC 2007a: 629).

Rate of change in global temperatures

Much of the analysis of projected climate change impacts focuses on the temperature increase under a certain emissions pathway by a point in time, often 2100. However, it is not just the magnitude of temperature rise, but also the rate at which it occurs that determines climate change impacts, as a higher rate of change in temperature affects the adaptive capacity of natural and human systems (Warren 2006; Ambrosi 2007; IPCC 2007a: 774).

2.3.6 Changes to the climate system

Elements of the climate system respond in different ways to changes in the energy balance. The climate response can vary considerably over time and space.

Many considerations of climate response focus on the changes to mean climate at a particular point in time. These considerations do not address the inherent variability of the climate system or the long time frames over which the system will continue to respond even if the energy balance ceases to change.

Climate variability

Climate variability refers to the natural variations in climate from the average state, and occurs over both space and time.

Natural climate variability occurs as a result of variations in atmospheric and ocean circulations, larger modes of variability such as the El Niño – Southern Oscillation (see Box 2.2), and events such as volcanic eruptions and changes in incoming solar radiation.

Box 2.2 Large-scale patterns of climate variability

Analysis of variability in global climate over time has shown that a significant component can be described in terms of a relatively small number of large-scale patterns of variability in atmospheric and oceanic circulation (IPCC 2007a: 39).

A key example is the El Niño – Southern Oscillation, which is a coupled fluctuation in the atmosphere and the equatorial Pacific Ocean. The El Niño – Southern Oscillation leads to changed conditions in surface temperature across the central equatorial Pacific Ocean and are thought to be once every three to seven years, which in turn leads to changes in rainfall, floods and droughts on both sides of the Pacific. It is characterised by large exchanges of heat between the ocean and atmosphere, which affect global mean temperatures but also have a profound effect on the variability of the climate in Australia.

Another example relevant to Australia is the Southern Annular Mode, which is the year-round fluctuation of a pattern with low sea surface pressure in the Antarctic and strong mid-latitude westerlies (IPCC 2007a: 39). The Southern Annular Mode has a considerable influence over climate in much of the southern hemisphere, affecting surface temperatures in the Southern Ocean and the distribution of sea ice. It influences rainfall variability and storm tracks in southern Australia.

The extent of natural climate variability differs from place to place. In Australia, a notable feature of the climate is the high variability in rainfall from year to year, influenced by the El Niño – Southern Oscillation. This affects the ability of scientists to identify long-term trends in the climate system and to establish whether the changes result from human activities.

Sustained changes to the energy balance of the climate system will cause a change in the long-term means of elements of the climate system such as temperature and rainfall, but may also lead to a change in the pattern of variability about a given mean. In Australia, observed changes in the climate suggest that the frequency of extremes in rainfall events is increasing at a faster rate than the mean (CSIRO & BoM 2007).

Severe weather events

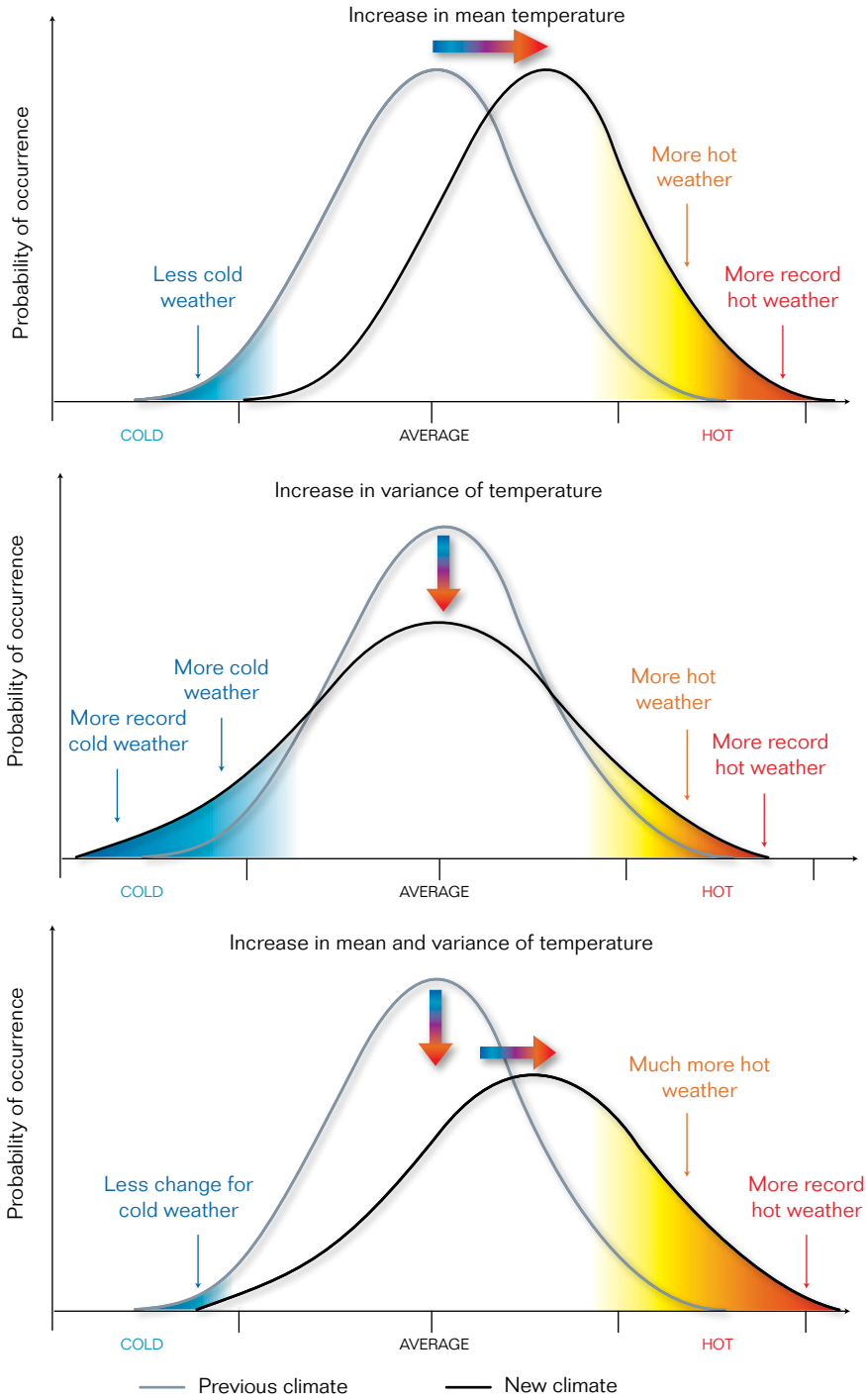
Natural variability in the weather is reflected in the occurrence of 'severe weather events', defined in this report as an event of an intensity that is rare at a particular place and time of year. Examples of severe weather events include:

- hot days and nights (including heatwaves)
- cold days and nights (including frosts)
- heavy rainfall events
- droughts
- floods
- hail and thunderstorms
- tropical cyclones
- bushfires
- extreme winds.

The characteristics of what is called severe weather may vary from place to place in an absolute sense—for example, the temperature required to define a heatwave in Hobart would be lower than in Darwin. Weather events may also be considered severe if they cause extensive damage due to timing or location, even if they are not considered rare in terms of their likelihood.

Climate change may result in changes to the frequency, intensity and distribution of weather events that are considered 'severe' today. Figure 2.5 demonstrates the effect of a change in mean temperature on the lower probability temperature events—in some cases, an increase in the frequency at one end of the probability (more heatwaves) will be associated with a decrease in the opposite end (fewer frosts). However, if the variability (or variance) were to change rather than the mean, hot days and frosts would both increase. If both the variability and the mean were to change, there would be fewer frost events but a more significant increase in the number of hot days.

Figure 2.5 Effect on extremes of temperature from an increase in mean temperature, an increase in variance, and an increase in both mean temperature and variance

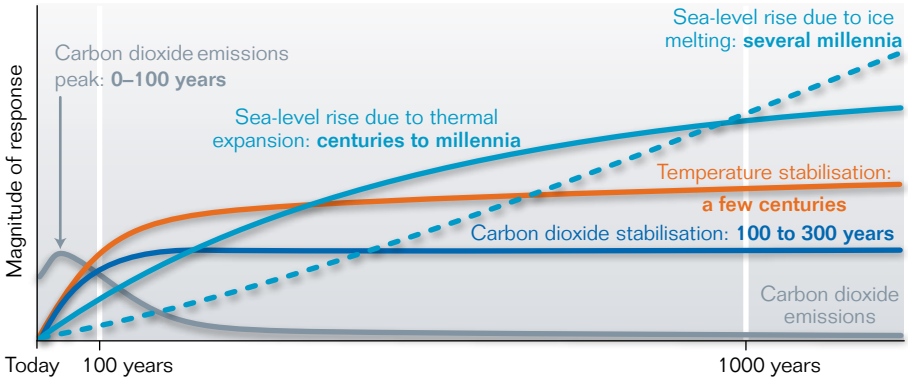


Source: IPCC (2001b: Figure 4.1), reformatted for this publication.

The slow response of the climate system

Figure 2.6 shows estimates of the time it takes for different parts of the climate system to respond to a situation where emissions are reduced to equal the rate of natural removal. While greenhouse gas concentrations stabilise in around a hundred years, the temperature and sea-level rise due to thermal expansion of the oceans takes much longer to stabilise. The melting of ice sheets is still increasing the sea level even after a thousand years.

Figure 2.6 Inertia in the climate system



Source: IPCC (2001b: Figure 5.2), reformatted for this publication.

2.4 The task of global mitigation

Goals for global mitigation have typically been cast in terms of stabilisation of greenhouse gas concentrations in the atmosphere. Article 2 of the United Nations Framework Convention on Climate Change (UNFCCC) states as its ultimate objective:

> [s]tabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system. Such a level should be achieved within a time frame sufficient to allow ecosystems to adapt naturally to climate change, to ensure that food production is not threatened and to enable economic development to proceed in a sustainable manner (UN 1992).

However, the UNFCCC does not define the point at which 'dangerous anthropogenic interference with the climate system' or 'dangerous climate change' might occur. Even if the climate change resulting from a given pathway of future emissions were known with certainty, there would be different approaches to defining 'danger', and interpretation of the UNFCCC goal will not be defined only by the science. Ethical, economic and political judgments will also be required (IPCC 2007b: 99).

The Review's terms of reference required it to analyse two specific stabilisation goals: one at which greenhouse gases are stabilised at 550 ppm CO_2-e and one at which they are stabilised at 450 ppm CO_2-e. A stabilisation target of 450 ppm

CO_2-e gives about a 50 per cent chance of limiting the global mean temperature increase to 2°C above pre-industrial levels (Meinshausen 2006), a goal endorsed by the European Union (Council of the European Union 2007) among others. Stabilisation at 500 ppm or 550 ppm CO_2-e would be less costly than a more ambitious target, but is associated with higher risks of dangerous climate change.

Based on a best-estimate climate sensitivity of 3°C, stabilisation at 550 ppm CO_2-e is likely to lead to an equilibrium global mean temperature increase of 3°C above pre-industrial levels (IPCC 2007b; Meinshausen 2006).

2.4.1 What is stabilisation?

Stabilisation of a greenhouse gas is achieved when its atmospheric concentration is constant. For a group of greenhouse gases, stabilisation is achieved when the combined warming effect (radiative forcing) of the gases is maintained at a constant level.

Stabilisation of long-lived greenhouse gases does not mean the climate will stop changing—temperature and sea-level changes, for example, will continue for hundreds of years after stabilisation is achieved (see Figure 2.6).

How does the lifetime of a gas influence stabilisation?

For all greenhouse gases, if emissions continue to increase over time their atmospheric concentration will also increase. However, the way in which the concentration of a gas will change in response to a decrease in emissions is dependent on the lifetime of the gas (IPCC 2007a: 824). Stabilisation of greenhouse gas emissions is therefore not the same as stabilisation of greenhouse gas concentrations in the atmosphere.

Carbon dioxide is naturally removed slowly from the atmosphere through exchange with other parts of the carbon cycle. The current rate of emissions is well above the natural rate of removal. This has caused the accumulation of carbon dioxide in the atmosphere. If carbon dioxide emissions were stabilised at current levels, concentrations would continue to increase over this century and beyond. To achieve stabilisation of carbon dioxide concentrations, emissions must be brought down to the rate of natural removal.

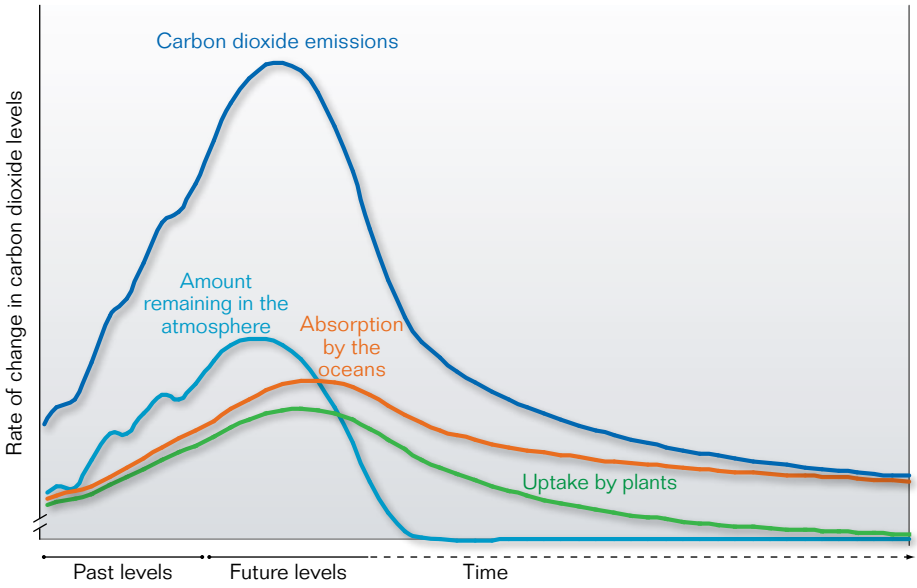
The rate of absorption of carbon by sinks depends on the carbon imbalance between the atmosphere, the oceans and the land, and the amount already contained in these sinks. Once stabilisation in the atmosphere is reached, the rate of uptake will decline (Figure 2.7). Long-term maintenance of a stable carbon dioxide concentration will then involve the complete elimination of carbon dioxide emissions as the net movement of carbon dioxide to the oceans gradually declines (IPCC 2007a: 824; CASPI 2008).

The response of other greenhouse gases to decreases in emissions is more straightforward: the level at which concentrations are stabilised is proportional to the level at which emissions are stabilised. For gases with a lifetime of less than a century (such as methane) or around a century (such as nitrous oxide), keeping emissions constant at current levels would lead to the stabilisation of

concentrations at slightly higher levels than today within decades or centuries, respectively (IPCC 2007a: 824). If anthropogenic emissions of these gases were to cease completely, their concentration levels would eventually return to pre-industrial levels.

For greenhouse gases other than carbon dioxide with lifetimes of thousands of years (such as sulphur hexafluoride), stabilisation would only occur many thousands of years after emissions stopped increasing. In the policy context they are treated in the same way as carbon dioxide, with the long-term aim of bringing emissions to zero in order to stabilise their warming effect.

Figure 2.7 Response of different carbon sinks to the rate of emissions over time



Source: Based on CASPI (2008).

How can stabilisation be achieved?

Any number of emissions pathways could lead to stabilisation of a gas at a given concentration. For carbon dioxide, these pathways generally involve a trade-off between the level at which emissions peak and the maximum rate of reductions required in the future. Figure 2.8 shows some of the possible emissions pathways to achieve the same stabilisation target. These curves are stylised—in the real world, annual emissions would fluctuate. The pathways that have a higher peak in emissions have a much greater rate of reduction at a later point in time, shown by the steepness of the curve.

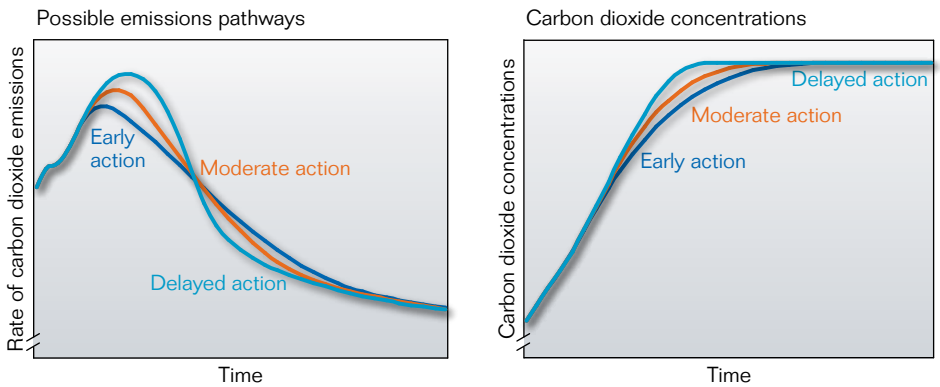
The timing of emissions reductions influences the efficiency of uptake of carbon dioxide by sinks, the rate of temperature increase and potentially the

timing of climate–carbon feedback effects. Delays in the realisation of a specified stabilisation target through more rapid mitigation would, if they were large, give greater environmental benefits (O'Neill & Oppenheimer 2004).

Is a target of 450 ppm CO_2-e or below achievable?

The concentration of long-lived greenhouse gases in the atmosphere for 2005 is equivalent to the warming effect of 455 ppm of carbon dioxide (see section 2.3.5). However, when the cooling influence of aerosols is included, the equivalent carbon dioxide concentration is estimated at 375 ppm CO_2-e. The concentration of carbon dioxide in 2007 was 383 ppm (Tans 2008).

Figure 2.8 Different pathways of emissions reductions over time to achieve the same concentration target



Source: Based on CASPI (2008).

Due to the short lifetime of aerosols in the atmosphere, it is not appropriate to include their influence in a long-term target. The presence of aerosols is expected to lessen through a reduction in the burning of fossil fuels as a result of climate change policies as well as through separate efforts to reduce air pollution.

The 2005 long-lived greenhouse gas concentration of 455 ppm CO_2-e includes the warming influence of gases such as methane, which can be reduced in a relatively short period of time. If the target were set for some point in the future (such as 2050 or 2100), it would be scientifically feasible to bring CO_2-e emissions down to a target level of 450 ppm CO_2-e if immediate and deep cuts were made in emissions of most greenhouse gases.

However, for a target of 450 ppm CO_2-e to be achieved, global emissions would have to peak and then fall almost immediately at a high rate. In the context of current emissions trends, it would take a major change in many countries' positions for early international agreement on such dramatic changes in emissions to be feasible.

Overshooting

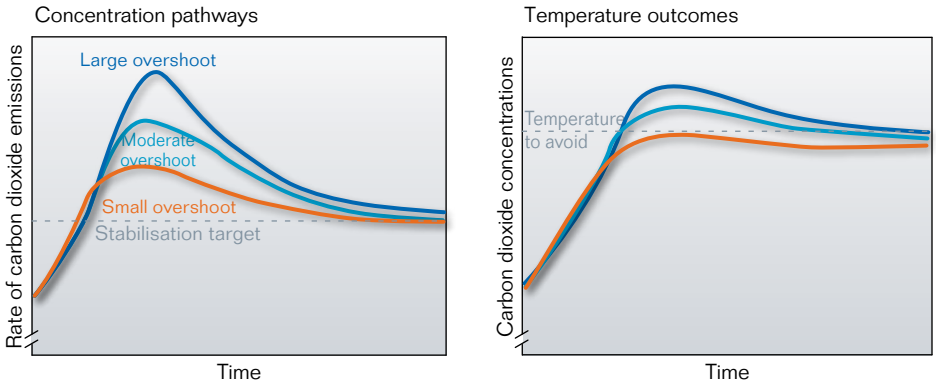
There is increasing recognition in both science and policy communities that stabilising at low levels of CO_2-e (around or below 450 ppm) requires 'overshooting' the concentration target (den Elzen et al. 2007; Meinshausen 2006; IPCC 2007a: 827).

The climate change impacts of the higher levels of greenhouse gas concentrations reached in an overshoot profile are dependent on the length of time the concentrations stay above the desired target, and how far carbon dioxide overshoots.

Figure 2.9 shows the different temperature outcomes for a range of cases of overshooting. All three cases show stabilisation at the same level in a similar time frame, but with varying amounts of overshooting. The temperature output demonstrates that while the 'small overshooting' case remains under the target temperature, the other cases do not. Hence, due to inertia in the climate system, a large and lengthy overshooting will influence the transient temperature response, while a small, short one will not (den Elzen & van Vuuren 2007).

Increasing attention is being paid in the environmental and scientific communities to low stabilisation scenarios. In particular, a number of organisations in Australia have suggested that the Review should focus as well on a 400 ppm objective. They argue that the risks of immense damage to the Australian environment, including the Great Barrier Reef and Kakadu National Park, are unacceptably high at 450 ppm. Some scientists have also expressed the view that stabilisation at 450 ppm is too high (Hansen et al. 2008). For any such scenarios to be feasible, there would need to be a considerable period of overshooting.

Figure 2.9 Temperature outcomes of varying levels of overshooting



Source: Concentration and temperature pathways developed using SIMCAP (Meinshausen et al. 2006).

What is a peaking profile?

An overshooting profile requires a period in which emissions are below the natural level of sequestration before they are stabilised. Another mitigation option is to follow a 'peaking profile'.

Under a peaking profile, the goal is to cap concentrations at a particular level (the peak) and then to start reducing them indefinitely, without aiming for any explicit stabilisation level. Stabilisation is therefore not conceived as a policy objective for the foreseeable future.

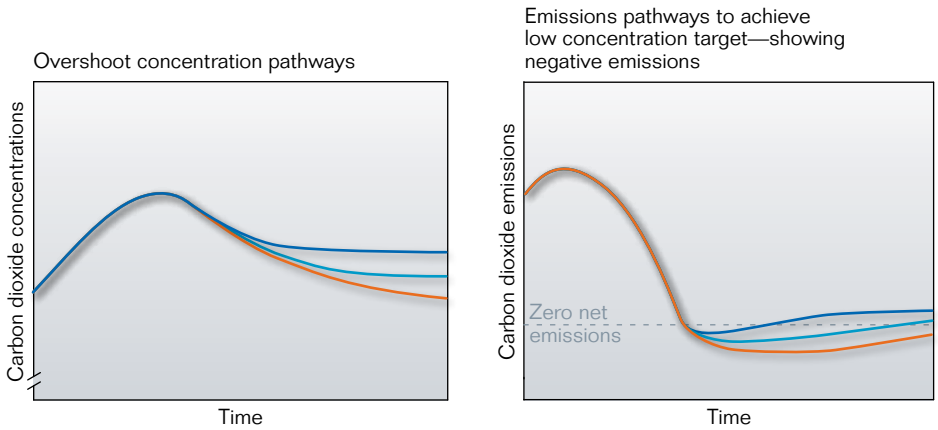
The key benefit of a peaking profile is that it allows concentrations to increase to or above the level associated with a given long-term temperature outcome, but reduces the likelihood of reaching or exceeding that temperature outcome. The higher level of peak concentrations means that current trends in emissions growth do not need to be reversed as quickly to achieve any given temperature goal. This decreases the costs of meeting a given temperature target (den Elzen & van Vuuren 2007).

Following a peaking profile could be a disadvantage if the climate is found to be more sensitive to increases in greenhouse gases than anticipated. Due to the higher concentrations reached under a peaking profile, there is less flexibility to adjust to a lower concentration target at a later point in time, so there is greater risk that a threshold may be crossed.

Is overshooting feasible?

Designing a mitigation pathway—whether an overshooting or a peaking profile—that requires a decrease in the concentration of greenhouse gases assumes that emissions can be brought below the natural level of sequestration. Figure 2.10 shows the emissions pathways required to achieve a low concentration target following an overshoot. A lower concentration target following an initial overshoot will require negative emissions net of natural sequestration for a longer period.

Figure 2.10 Emissions pathways required to achieve a low concentration target following an overshoot



Source: Based on CASPI (2008).

The costs of reducing emissions below natural sequestration levels would be lower if controls on gross emissions were supported by cost-effective means of removing carbon dioxide from the atmosphere. Bringing emissions below the natural rate of sequestration would require rigorous reduction of emissions from all sources, but might also require extraction of carbon dioxide from the air. Possible methods include:

- increasing absorption and storage in terrestrial ecosystems by reforestation and conservation and carbon-sensitive soil management
- the harvest and burial of terrestrial biomass in locations such as deep ocean sediments where carbon cycling is slow (Metzger et al. 2002)
- capture and storage of carbon dioxide from the air or from biomass used for fuel
- the production of biochar from agricultural and forestry residues and waste (Hansen et al. 2008).

The simplest way to remove carbon dioxide from the air is to use the natural process of photosynthesis in plants and algae. Over the last few centuries, clearing of vegetation by humans is estimated to have led to an increase in carbon dioxide concentration in the atmosphere of 60 ± 30 ppm, with around 20 ppm still remaining in the atmosphere (Hansen et al. 2008). This suggests that there is considerable capacity to increase the level of absorption of carbon dioxide through afforestation activities. The natural sequestration capacities of algae were crucial to the decarbonisation of the atmosphere that created the conditions for human life on earth, and offer promising avenues for research and development.

Technologies for capture and storage of carbon from the combustion of fossils fuels currently exist, and the same process could be applied to the burning of biomass.

As yet, there are no large-scale commercial technologies that capture carbon from the air. Some argue, however, that it will be possible to develop air capture technologies at costs and on timescales relevant to climate policy (Keith et al. 2006). Research and development in Australia on the use of algae is of global importance. Captured carbon dioxide could be stored underground or used as an input in biofuel production.

Under a carbon price applying broadly across all opportunities for carbon dioxide reduction and removal, and with strong research and development support, there will be more rapid commercial development of both existing and new technologies to achieve negative emissions on a large scale.

2.4.2 Geo-engineering

'Geo-engineering' is a term used to describe 'technological efforts to stabilize the climate system by direct intervention in the energy balance of the earth' (IPCC 2007b: 815).

A range of geo-engineering proposals have been put forward, including:

- the release of aerosols into the stratosphere to scatter incoming sunlight (Crutzen 2006)
- cloud seeding through the artificial generation of micro-meter sized seawater droplets (Bower et al. 2006)
- fertilisation of the ocean with iron and nitrogen to increase carbon sequestration (Buesseler & Boyd 2003)
- changes in land use to increase the albedo (reflectivity) of the earth's surface (Hamwey 2005).

Geo-engineering proposals appear to have several advantages. First, they may be cheap in comparison to reductions in greenhouse gas emissions. They can be implemented by one or a small number of countries and thus do not require the widespread global action that stabilisation of greenhouse gases will require (Barrett 2008). They may act quickly, with a lag from implementation to impact of months rather than decades. Geo-engineering techniques could potentially be deployed to avoid reaching a tipping point related to temperature increase.

However, such proposals also have disadvantages.

- Those that focus on reducing solar radiation will do nothing to prevent the acidification of the ocean as a result of increased atmospheric concentrations of carbon dioxide, and therefore leave untouched part of the wider environmental problem.
- Geo-engineering techniques are generally untried. Some studies have been undertaken including through small-scale experiments on ocean fertilisation (Buesseler & Boyd 2003), investigation of similar natural phenomenon such as the release of aerosols from Mount Pinatubo in 1991, and computer simulations (Wigley 2006; Govindasamy & Caldeira 2000). However, there will always be the risk of unanticipated consequences, which need to be considered analytically.
- The fact that these solutions can be implemented unilaterally may also give rise to risks of conflict.

So far, the disadvantages of geo-engineering approaches have tended to outweigh the advantages in most minds that have turned to the issue. However, in recent years such proposals have received more support from a number of prominent scientists and economists, with calls for more research into their feasibility, costs, side effects and frameworks for implementation (IPCC 2007c: 79; Crutzen 2006; Cicerone 2006; Barrett 2008).

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